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HANDBOOK

OF

WEATHER FORECASTING

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PREFACE

The Handbook of Weather Forecasting was written mainly for distribution within the Meteorological Office to provide forecasters with a comprehensive and up-to-date reference book on techniques of forecasting and closely related aspects of meteorology. The work, which appeared originally as twenty separate chapters, is now re-issued in three volumes in loose-leaf form to facilitate revision.

Certain amendments of an essential nature have been incorporated in this edition but, in some chapters, temperature values still appear in degrees Fahrenheit. These will be changed to degrees Celsius when the chapters concerned are completely revised.

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TEMPERATURE

14.1. 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 every wavelength, the maximum possible radiation at a given temperature is known as a "black body". Planck's Law states that the energy distribution in the spectrum of radiation from a black body is represented by the formula

$$E_\lambda = \frac{C_1\lambda^{-5}}{\mathrm{Exp}\left(C_2/\lambda T\right) - 1},$$

where λ is the wavelength, C_1 and C_2 are constants and T is the absolute temperature. The total black-body radiation is obtained by integrating Planck's formula for all wavelengths and is given by the expression σT^4, where σ is a constant and T is the absolute temperature. This is known as Stefan's Law. The wavelength for which E_λ is a maximum is given by the formula

$$\lambda_{max} = \frac{C_3}{T}$$

where C_3 is a constant. This is Wien's Law. For λ_{max} in microns and T in degrees absolute, $C_3 = 2940$.

Matter both absorbs and emits radiation. The absorptivity ($a_{\lambda T}$) 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 emmissive power on both wavelength and temperature. Kirchoff's Law states that at a given temperature the ratio of emissive power to absorptivity $e_{\lambda T}/a_{\lambda T}$ for a given wavelength is a constant for all bodies. It follows from Kirchoff'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 Kirchoff'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

$$I = I_o \text{ Exp } (-k_\lambda m) \text{ ,}$$

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.

14.2. SOLAR RADIATION

The prime source of energy on the earth and in its atmosphere is solar radiation. For the purpose of this handbook the sun may be regarded as a black body radiating at an effective temperature of about 6,000°A. From Planck's Law it follows that 99 per cent of the radiated energy from the sun is contained between the wavelengths 0.15 to 4μ. About one half of this energy is contained between the wavelengths 0.4 to about 0.8μ, that is, within the visible spectrum. In accordance with Wien's Law the wavelength of the maximum intensity of solar radiation is about 0.5μ. Solar radiation traverses space and reaches the outer limits of the earth's atmosphere where its intensity (measured perpendicularly to its beam) is fairly constant at a value of about 2.0 calories per square centimetre per minute at mean solar distance. This is the solar constant. Its value is difficult to measure very accurately with certainty and there may be variations of a few per cent. If these variations are real it would be natural to expect some relationship between them and global weather conditions but the current state of knowledge on this is very meagre and inconclusive. The amount of solar radiation incident upon unit horizontal area at the limit of the earth's atmosphere varies both with latitude and with season.

After solar radiation impinges on the atmosphere it is absorbed, scattered and reflected to varying extents by the atmosphere before it reaches the ground where it becomes available for warming the earth's surface. The following gaseous constituents of the atmosphere are known to have absorption bands in the wavelengths of solar radiation: carbon dioxide, oxygen, ozone and water vapour. The absorption bands of carbon dioxide and oxygen are small and these gases absorb only a very small fraction of incident radiation. Ozone which is generally present in the upper atmosphere has a marked absorption band below 0.3μ and effectively absorbs all solar radiation of shorter wavelengths. Water vapour has a number of absorption bands in the solar spectrum and is the principal absorber in the troposphere. Ozone and water vapour probably absorb about 15 per cent of incident solar radiation.

When radiation reaches the atmosphere molecular scattering causes some radiation to be scattered back to space and some to be scattered forward into the atmosphere. An approximate estimate is some 10 per cent of the incident radiation scattered back to space and therefore lost to the atmosphere and some 10 per cent scattered earthwards.

Radiation is further depleted by reflection from the earth's surface and clouds. The fraction of incident radiant energy which a body reflects is known as its albedo. The albedo of the earth's surface depends on the nature of that surface and varies through a wide range of values. Although present forecasting techniques do not use values of the albedo, Table 1 has been included so that forecasters may have a qualitative understanding of the effects of reflection.

TABLE 1 *Albedo of various surfaces*[1]*

Land	*Albedo*	*Water (direct sun only)* Z °	*Albedo*	*Cloud*	*Albedo*
Desert	.24 - .28	0 (i.e. sun overhead)	.02	Stratus overcast 0-500 ft. thick	.05 - .63
Fields (green wheat etc.)	.03 - .15	20	.021	Stratus overcast 500-1,000 ft. thick	.31 - .75
Fields (dry, ploughed)	.20 - .25	40	.025	Stratus overcast 1,000-2,000 ft. thick	.59 - .84
Grass (various conditions)	.14 - .37	50	.034	Stratocumulus overcast	.56 - .81
Ground (bare)	.10 - .20	60	.060	Altostratus, occasional breaks	.17 - .36
Mould (black)	.08 - .14	70	.134	Altostratus overcast	.39 - .59
Sand (dry)	.18	80	.348	Cirrostratus overcast	.44 - .50
Sand (wet)	.09	85	.548	Cirrostratus and altostratus overcast	.49 - .64
Snow or ice	.46 - .86	90	1.0		

Z = sun's distance in degrees from the zenith

In general wet surfaces have a lower albedo than dry surfaces but snow cover causes the greatest change in the albedo of the ground. Freshly fallen snow may have an albedo in excess of 0.8. This decreases with age and for old wet snow the albedo probably falls within the range of 0.46 to 0.65. The albedo of extensive water surfaces depends very largely on the elevation of the sun but the roughness of the surface also has a slight effect, the albedo of a rough water surface being generally somewhat higher than that of a smooth glassy surface.

Except when the ground is snow covered, clouds are the most important reflectors of solar radiation. The albedo of cloud varies through a wide range of values depending on thickness and cloud amount. Neiburger[2] made observations on stratus clouds which affect the Californian coast in summer. He found a rapid variation of albedo for stratus clouds less than 1,000 feet thick. Although there was a considerable scatter in his observations particularly for thin clouds he found albedos as low as 0.1 for clouds about 200 feet thick but his average albedo for stratus 500 feet thick had risen to over 0.4. For clouds 1,000 feet thick the albedo rose to about 0.6 and approached a value of about 0.8 as thickness increased beyond 1,500 feet. Thus for a moderately thick overcast the albedo will be about 0.6 and for very thick cloud layers the albedo may be as high as 0.8.

Angström has given the following formula for the relation of albedo to cloud amount, n, where $\bar{n}$ is the fraction of the sky covered by cloud:

$$\text{Albedo} = 0.17 + 0.53n$$

Thus the amount of solar radiation reaching the surface of the earth and becoming available for heating that surface is but a fraction of the incident solar radiation. This fraction varies widely from day to day. Thick layers of

* The superscript figures refer to the bibliography at the end of this chapter.

cloud normally have the greatest effect but a complete cover of snow on the ground can exert about the same effect.

14.3. TERRESTRIAL RADIATION

The earth emits terrestrial radiation over a range of wavelengths appropriate to its temperature in accordance with Planck's Law. To fix ideas assume an earth temperature of 300°A. (about 80°F.). Terrestrial radiation is almost wholly contained in the wavelengths 3 to 80μ. About half its energy is within the wavelength 8 to 14μ and, according to Wien's Law, the maximum intensity of radiation is at about 10μ. Compared with the range of solar radiation (0.15 to 4μ with a maximum at 0.5μ) wavelengths of terrestrial radiation are much longer. In meteorology it has become customary to refer to solar radiation as short-wave and terrestrial radiation as long-wave radiation. 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 17μ. Within the wavelengths of terrestrial radiation the absorption by water vapour is very complex. Water vapour has strong absorption bands between 5.5 and 7μ and from about 14μ upwards. It is partially absorbent within the ranges 7 to 8.5μ and 11 to 14μ. Between 8.5 and 11μ 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[3] 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 7μ and from 14μ upwards. For air with a dew point of 50°F. and temperatures between 50 and 68°F. this amount of water is present at ground level in a column of air about 100 feet 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°A. (about −27°F.). 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 minimum temperature from falling to the value it would reach with clear skies.

14.4. SKY RADIATION

Another factor which must be considered is the sky radiation. Carbon dioxide and water vapour absorb long-wave (and, to a much more limited degree, short-wave) radiation. By Kirchoff'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 layer. 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.

Angström[4,5] found that the downward flux of radiation (R_o) to the earth from a clear sky could be represented by a formula:

$$R_o = \sigma T^4 (A - B \times 10^{-\gamma e}),$$

where $A = 0.75$, $B = 0.32$, $\gamma = 0.069$, e is the partial pressure in millibars of water vapour at the surface and σT^4 is the total black-body radiation at temperature T.

Brunt[6] suggested a different formula:

$$R_o = \sigma T^4 (a + b\sqrt{e}).$$

Several series of observations are available from which a and b can be determined and it is found that the values for different series of observations vary widely. Mean values are $a = 0.43$ and $b = 0.08$ when e is in millibars.

Values of T(°A.) and e (millibars) at the surface are widely available so that the formulae can be readily applied. The units of R will correspond with those for the particular value of Stefan's constant which is inserted, that is, if σ is taken as 1.38×10^{-12} gm.cal.cm.$^{-2}$sec.$^{-1}$ then R is obtained in gm.cal.cm.$^{-2}$sec.$^{-1}$. A defect in both formulae is that only surface values of temperature and humidity are used. More elaborate calculation of sky radiation can be made using an Elsasser[7] or a Kew[8] radiation chart which requires information regarding the humidity and temperature distribution in the vertical.

A cloud layer radiates as a black body and the downward flux of sky radiation is then greatly increased. Asklöff[9] modified Angström's formula to apply to cloudy skies. The formula is

$$R = \sigma T^4 - \alpha\,(\sigma T^4 - R_o)$$

where α is a factor depending on the height of the cloud base and R_o is the downward flux from a clear sky. The following values of α were determined from observational data:

Cloud height (km.)	1.5	3	7
α	0.14	0.25	0.80

Philipps[10] calculated rather different figures:

Cloud height (km.)	2	5	8
a	0.17	0.38	0.45

Thus from the formula the lower the cloud the smaller is a and so the greater is R - the downward flux.

14.5. TRANSFER OF HEAT IN AND NEAR THE GROUND

Having reviewed the various radiative processes effective at the surface of the earth and its overlying atmosphere attention will now be given to the radiant energy absorbed by the surface of the earth. This energy will warm the surface and so modify the thermal gradient in the underlying surface. Heat will be conducted down the thermal gradient and the precise rate depends on the thermal properties of the surface layers. The heat flux through the surface of the earth can be represented by the equation

$$Q_{\circ} = -k_1 \rho_1 c_1 \left(\frac{\partial T}{\partial z}\right)_{\circ},$$

where k_1 is the thermometric conductivity, ρ_1 is the density and c_1 is the specific heat of the underlying matter. k_1, ρ_1, and c_1 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[11] values of specific heats range from 0.83 for clay to 0.27 for sandy loam. Densities are very variable but Sutton[12] gives the following typical values which are approximate.

	$gm.cm.^{-3}$
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

As solar radiation impinges on the surface, heat will be conducted downwards and the temperature of the ground near the surface will rise. At night the surface of the earth will lose heat by radiation and after some time the temperature gradient in the layers near the surface will be reversed and heat is conducted upward to the surface. Observations show that this diurnal wave of temperature does not penetrate very far into the ground and is virtually negligible at about

50 to 60 centimetres depth. The long-term annual variation penetrates to much greater depths. At the surface of the ground the mean diurnal variation is large. From observations made at Kew during May 1953 - May 1954 Roach[13] concluded that the diurnal variation of temperature at the surface of ground covered by close cropped grass is just about twice that of the screen temperature all the year round. In summer diurnal ranges of 30°C. (54°F.) were not uncommon and he considered that a range up to 40°C. (72°F.) was quite possible - especially in periods of summer drought - and would be most likely to occur between about mid May and mid June. Extreme values of the ground surface temperature at Kew for the year May 1953 - May 1954 were 39°C. (102°F.) and −12°C. (10°F.). Roach considered that probable absolute extremes at Kew are 47°C. (117°F.) (but 52°C. (126°F.) over parched brown turf) and −18°C. (0°F.). On a clear night the excess of temperature in the screen over that at the ground surface can reach 5°C., and more over snow, but is more often 2°-3°C. On a hot day in June the deficit of screen temperature below ground temperature is usually around 10°C. but may reach 15°C. The mean times of the occurrence of daily maxima and minima at the ground surface are substantially earlier than those of screen temperatures: the maxima are about 2½ hours earlier in summer and one hour in winter, while the minima are about 1½ hours earlier in summer and two hours in winter.

Some measurements of the diurnal variations of temperatures at various depths in the soil have recently been made at sites near Cambridge using an apparatus described by Rider[14]. From measurements made at depths of 1, 5, 10, 20 and 40 centimetres below three different types of surface Rider has prepared graphs, as yet unpublished, of the monthly means of the diurnal temperature range for July 1954 and January 1955 for depths of up to 40 centimetres. The graphs are reproduced in Figure 1. Mean variations in July 1954 ranging from 8° to 14°C. (15° to 24°F.) at 1 cm. depth are reduced rapidly to about 4° to 7.5°C. (8° to 14°F.) at 5 cm., 2° to 4°C. (4° to 8°F.) at 10 cm. and about 1° to 2°C. (2° to 4°F.) at 20 cm. At 40 cm. mean diurnal variations are less than 1°C. (2°F.). The mean daily range of screen temperatures at the same site in July

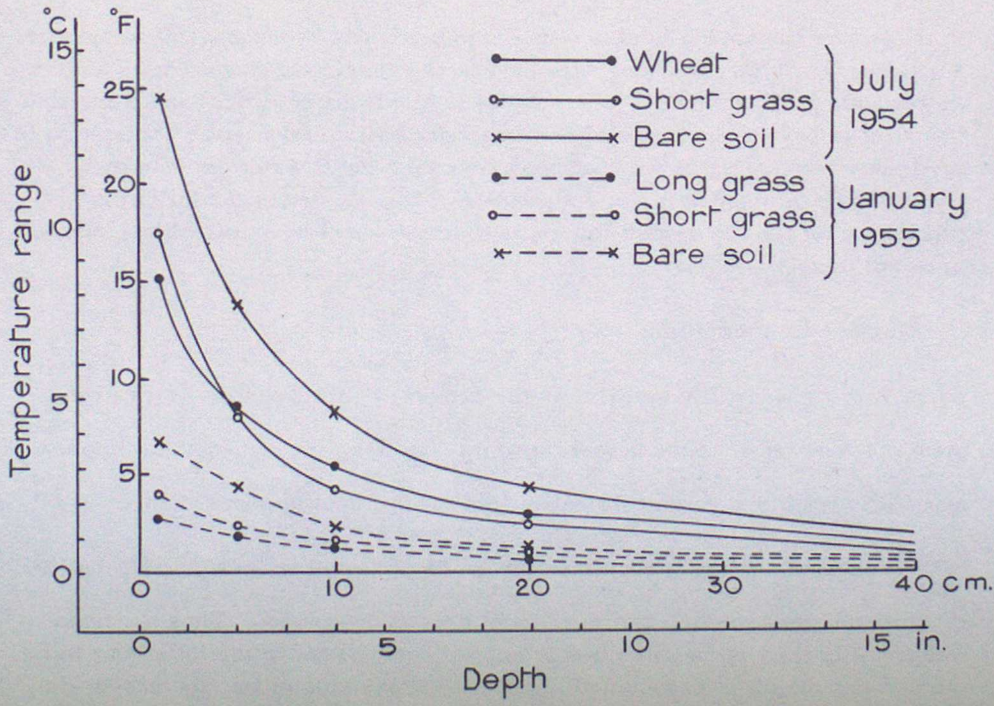

Figure 1 *Means of diurnal temperature ranges in the ground for three surfaces at Cambridge for July 1954 and January 1955 (After Rider[14])*

1954 was 7.7°C. (14.0°F.). It is seen that in July 1954 the mean variation at 1 cm. under a surface on which wheat was growing exceeded the mean variation in the screen, and at 1 cm. under bare soil was nearly twice this value. July 1954 was a dull cool month so that the values found by Rider would be an underestimate of mean conditions in sunny weather in summer. Measurements of ground and surface temperature at Cambridge are not available but by combining the Kew and Cambridge results forecasters can see that quite large temperature gradients can and do exist in the ground near the surface. Day-time gradients are substantial and are reversed at night.

The ground surface heats the air in direct contact with it and during sunny periods large temperature gradients are set up in the air in the lowest few centimetres. These lapse rates may be many times the dry adiabatic. Best[15] found the following values for midday mean temperature differences over grass at Porton on clear days.

Layer	*Temperature differences*		*Temperature differences corresponding to dry adiabatic lapse rate*
	June	*December*	
cm.	°*F.*	°*F.*	°*F.*
2.5 - 30	−3.96	−1.44	−0.0054
30 - 120	−1.73	Values in error	−0.016

Negative values indicate temperatures decreasing upwards

Air with such high lapse rates is unstable and it is now believed that, owing to their buoyancy, bubbles of heated air rise from the ground and ascend into the cooler air above. Discrete bubbles of air probably leave the ground at the rate of a few a minute according to estimates by Swinbank.[16] This convective process transfers heat from the surface of the earth upwards within the vertical limits of convection.

Heat from the ground is also transferred vertically by mechanical turbulence. A great deal of effort has been expended on the development of theories and understanding of turbulence. The subject is highly mathematical and a detailed treatment is beyond the scope of this handbook particularly as the application to day-to-day forecasting of the equations governing the transfer of atmospheric properties by turbulence is not yet practical. Brunt[17] quotes the following expression for the net upward flux of heat across unit horizontal surface due to turbulent transfer:

$$\text{Flux of heat (upwards)} = -K\rho c_p\left(\frac{\partial T}{\partial z} + \Gamma\right),$$

where K is the eddy diffusivity, ρ is the density, c_p the specific heat at constant pressure, $-\frac{\partial T}{\partial z}$ is the actual lapse rate and Γ is the dry adiabatic lapse rate. It is readily seen that the sign of the flux of heat upwards changes as $\frac{\partial T}{\partial z} + \Gamma$ changes through zero. When the lapse rate $-\frac{\partial T}{\partial z}$ is less than Γ, that is, the atmosphere is stable, then the flux of heat is downwards. Thus the transference of heat by eddies in a stable atmosphere will cause the lower part to be warmed and the upper part cooled; hence the lapse rate will be increased. In the absence of other factors turbulent transfer of heat will continue to modify the lapse rate of the layer of the atmosphere through which turbulence is effective until that lapse rate becomes dry adiabatic.

Variation of temperatures in the surface layers of seas and extensive inland waters is markedly different from that of the land. As the albedo for water surfaces is small, except when the sun's elevation is low, a large part of the incident solar radiation is absorbed by the water mainly in the first few feet. About 70 per cent is absorbed in the first six feet in clear oceans and this rises to about 85 per cent for turbid coastal waters. Owing to this penetration of solar radiation, to conduction of heat and to mixing due to turbulent motion of the water near the surface, the sensible heat for warming the water is spread through a considerable layer. At night when outgoing radiation causes surface waters to cool, heat is transferred from lower layers mainly by convection and mixing. Thus, except for very shallow waters of limited size, the diurnal variation of temperature of water surfaces is small compared with land surfaces. Annual variations of temperature are smaller and occur later in the year. A knowledge of these differences between land and sea is necessary for forecasting temperatures particularly as no part of the British Isles is very far from the sea.

14.6. VARIABILITY OF TEMPERATURES

Before rules are given for forecasting temperatures it is desirable that forecasters should have some knowledge of the variability of temperatures. The range, scale and periods of fluctuations and oscillations of temperature vary widely. In the large scale of major frontal depressions the range may be some 10°-20°F., the scale of the order of 1,000 kilometres and the period a few days. In the medium scale there are fluctuations of a few degrees on a horizontal scale of some 300 kilometres which persist for a few hours and are normally revealed by the synoptic network of upper air and surface reporting stations. These variations within an air mass arise from local variations of topography, precipitation or surface conditions and, when observed, some allowance can be made for the horizontal advection of colder or warmer air. Variations of smaller scale tend to slip through the network - particularly of upper air stations. Even where the variation is observed the interpretation of the information is difficult owing to the limited size of the fluctuation.

The small-scale fluctuations in temperature in clear air have been observed by the Meteorological Research Flight (MRF) over a number of years. From an analysis of a number of earlier flights made at various levels between about 700 and 800 millibars, Frith[18] concluded that "closed temperature and humidity patterns, on a scale measured in terms of miles, exist, with some degrees of persistence, in the free atmosphere. There is some similarity between temperature and humidity isopleths, drier air tending to be warmer than moist. Temperature variations are probably usually of the order of 1 to 2 degrees (F.) but may be much more. The variations in frost point may be anything up to 30 degrees (F.)."

Later flights by MRF using aircraft equipped with an ultra rapid thermometer have been analysed by Grant[19] who draws the following conclusions:

"(*a*) Horizontal variations of air temperature up to at least 1°F. can exist over distances of the order of 100 feet at all heights up to at least 30,000 feet.

(*b*) These fluctuations are very common near inversions.

(*c*) There is always bumpiness in the region in which appreciable temperature fluctuations are present.

(d) There are occasionally present in the atmosphere at all heights up to at least 17,000 feet patches of warm and cold air with horizontal dimensions of 200 - 2,000 feet and temperature 2°-4°F. different from the surrounding air.

(e) There are sometimes discontinuities of temperature in the horizontal of up to 2°F., particularly near inversions."

Grant also reaches some rather more speculative conclusions of which the the following are of direct interest to practical forecasters.

"*(f)* Large patches of cold air occur above the tops of cumulus cloud whose growth has been stopped by a stable layer.

(g) Large patches of warm and cold air occur on days of extreme instability.

(h) Small-scale fluctuations occur in the stratosphere as well as the troposphere."

That variations in temperatures at screen level are always present in the same air mass is readily apparent from inspection of any synoptic chart or pairs of records from neighbouring observing stations. These variations are the integrated effect of a variety of differences in the nature and state of the soil, the type and denseness of vegetation, topography etc.

These variations are real and demonstrate that the atmosphere is far from homogeneous. Nevertheless on the synoptic scale and in the general smoothing and coarseness of synoptic analysis this relatively small-scale heterogeneity is smoothed out. Synoptic charts show a broad homogeneity of characteristics within an air mass which enables forecasts of temperatures for a particular place or area to be made with an accuracy which appears to satisfy present requirements and which, bearing in mind the small-scale fluctuations, is reasonably indicative of the temperatures likely to be recorded.

14.7. FORECASTING TEMPERATURES NEAR THE GROUND

From the preceding considerations of the various heat processes which are effective in determining the temperature of the atmosphere it would be possible to write down equations which state the balance existing in the atmosphere at the time under consideration. In the present state of knowledge this is of no practical value to the operational forecaster. The values of the various quantities are not readily observed, for example, the density and specific heats and conductivities of the various soils in the area for which a forecast is to be made, the coefficient of eddy diffusivity, the various radiative fluxes - none of these observations is made in the United Kingdom at a network of stations, nor would it be practical to do so. Even if the observations were available the manipulation of the equations would preclude their numerical application to practical forecasting at the present time. The essence of the practical approach to day-to-day forecasting lies in experience of the temperatures accompanying varying conditions of cloud and air mass. Diurnal and other temperature changes can be anticipated by consideration of the physical processes involved and by use of empirical formulae which have been devised for simple meteorological situations. The variations of temperatures 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. From 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, strength and speed of winds, cloud amounts and thicknesses, precipitation etc., that extreme value is modified quantitatively sometimes by objective methods (although these objective methods depend on the subjective forecast) but more often the value has to be adjusted in the light of the forecaster's experience and interpretation of the physical processes at work.

14.7.1. Day heating

14.7.1.1. Maximum day temperature. A ~~The~~ method ~~most~~ widely used to forecast the maximum day temperature in the neighbourhood of the United Kingdom is based on *Professional Note No. 63* by Gold.[20] In this note Gold made broad assumptions regarding the radiative balance at the ground and in the lowest layers of the atmosphere during the period from dawn to 3 p.m. (the approximate time of maximum day temperature) and calculated the probable maximum energy actually 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.

For his calculations Gold starts with the approximate amount of solar radiation entering the earth's atmosphere in about latitude 50°N. in one day near the middle of each month. To allow for the earth's radiation which is not absorbed by the lower layers of the atmosphere he assumed that 30 per cent of the total earth radiation would escape. He had to estimate temperatures for the radiating surface of the earth on a clear day to compute the flux of radiation but variations of up to 10°F. did not substantially modify the results. Values for diffuse sky radiation were taken and added to incoming solar radiation. Reflected solar radiation was taken as 10 per cent of the total solar radiation. Finally an allowance for evaporation of water from the ground and vegetation was made. By these means Gold was able to estimate the energy available for warming the lower layers of the atmosphere. As energy is represented on a tephigram by an area, the available energy can be regarded as an area on a tephigram and this provides the simple practical application of the method.

Table 2 gives the critical values of the areas corresponding to the middle of each month for application of Gold's method to the 1955-63 ~~56~~ editions of the tephigram. In *Professional Note No. 63* the areas were given in units of "squares" on the tephigram - each square being formed by the intersection of isolines of temperature and potential temperature at 10°F. intervals. Tephigrams now have centigrade scales and the squares of 10°C. differences are inconveniently large. Table 2 contains values for the critical area in square centimetres.

TABLE 2 *Areas on a tephigram corresponding to energy available on a clear day for warming the lower layers of the atmosphere to 1500 local time*

Form No.	*Jan.*	*Feb.*	*Mar.*	*Apr.*	*May*	*June*	*July*	*Aug.*	*Sept.*	*Oct.*	*Nov.*	*Dec.*
						square centimetres						
~~Form 2810 1955 edition and~~ Large scale insert on Met forms 2810A, 2810B 1956 edition	3.2	6.1	9.2	12.4	14.4	16.1	14.8	13.3	10.5	7.3	3.9	2.1
Met Form 2810 19~~56~~ 63 edition	~~1.7~~ 1.8	~~3.2~~ 3.4	~~4.8~~ 5.2	~~6.4~~ 6.9	~~7.5~~ 8.1	~~8.3~~ 9.0	~~7.7~~ 8.3	~~6.9~~ 7.4	~~5.4~~ 5.9	~~3.8~~ 4.1	~~2.0~~ 2.2	~~1.1~~ 1.2
Small scale section of Met Forms 2810A, 2810B 1956 edition	0.8	1.5	2.3	3.1	3.6	4.0	3.7	3.3	2.6	1.8	1.0	0.5

It is also useful to know for a few times of the day the amount of energy which is likely to have become available since sunrise for warming the lower layers of the atmosphere on clear days. The values given in Table 3 are based mainly on some figures given by Wallington[21] and are applicable to Southern England.

TABLE 3 *Areas on a tephigram corresponding to energy which has become available since sunrise, for various times during a clear day, for warming the lower atmosphere.*

Form No.	*Local time*	*Jan.*	*Feb.*	*Mar.*	*Apr.*	*May*	*June*	*July*	*Aug.*	*Sept.*	*Oct.*	*Nov.*	*Dec.*
		Month											
		square centimetres											
~~Form 2810 1955 edition and~~ Large scale insert on Met Forms 2810A, 2810B 1956 edition	0800	0.0	0.2	0.8	2.3	4.5	5.8	4.6	3.0	2.3	0.8	0.2	0.0
	1000	0.7	1.9	3.5	5.3	7.5	9.2	7.7	6.2	3.9	3.1	1.6	0.6
	1200	2.1	2.8	6.4	9.1	11.2	13.9	11.6	9.2	6.9	5.5	3.2	1.8
Met Form 2810 1963 edition	0800	0.0	0.1	0.4	~~1.2~~ 1.3	~~2.3~~ 2.5	~~3.0~~ 3.2	~~2.4~~ 2.6	~~1.6~~ 1.7	~~1.2~~ 1.3	0.4	0.1	0.0
	1000	0.4	1.1	2.0	3.0	4.2	5.2	4.3	3.5	2.2	1.7	0.9	0.3
	1200	1.2	1.6	3.6	5.1	6.3	7.8	6.5	5.2	3.9	3.1	1.8	~~0.9~~ 1.0
Small scale section of Met Forms 2810B, 2810A 1956 edition	0800	0.0	<0.1	0.2	0.6	1.1	1.5	1.1	0.7	0.6	0.2	<0.1	0.0
	1000	0.2	0.5	0.9	1.4	1.9	2.3	1.9	1.5	1.0	0.8	0.4	0.1
	1200	0.5	0.7	1.6	2.3	2.8	3.5	2.9	2.3	1.7	1.4	0.8	0.5

It is readily apparent from the assumptions made, that the energy which is actually available on any one day may vary significantly from the figures given. Experience and the test of time have shown that Gold's values give quite reasonable results for clear days. The method for clear days without change of air mass is:

(i) Examine the synoptic charts and, from the air flow in the lowest two or three thousand feet, determine the location of the air mass which will lie over the forecast area.

(ii) Examine all available upper air temperatures for this air mass and estimate any modifications to the lapse rate of temperature and humidity due to subsidence, advection over sea, etc. Sketch in this lapse rate expected at dawn over the forecast area.

(iii) Determine from Table 2, interpolating as necessary for days not near the middle of the month, the area on the tephigram corresponding to the resulting available energy.

(iv) Estimate the representative moisture-content line for the lowest 2,000 feet of the atmosphere at mid-afternoon.

(v) Draw a dry adiabatic line from the surface pressure until it meets the environment curve or until it meets the moisture-content line. If it meets the moisture-content line before reaching the environment curve then extend it along a wet adiabatic curve to meet the environment curve. The area between the initial estimated environment and the constructed dry and wet adiabatic curve has to be equal to the area determined at (iii) above. In practice this is readily performed on a trial and error basis. Two scales which may assist in the assessment of areas are shown in Figure 2.

(vi) The temperature at which the dry adiabatic intersects the surface pressure on the tephigram gives the estimate of the maximum temperature.

If, in carrying out the operation described in (v) above, the dry adiabat intersects the representative moisture content line the forecaster will normally expect some convection cloud to form and it will be necessary to make an estimate of the amount by which the probable maximum temperature is likely to be reduced due to the development of convection cloud. (Other aspects of forecasting temperatures on days which are not clear are referred to in Para. 14.7.1.3).

The application of either scale of Figure 2 for estimating areas for an ascent curve plotted on a tephigram is illustrated in Figure 3. Let PQRS be an actual ascent curve and PAD the surface isobar. If the isothermal AC is drawn intersecting the curve at Q and the dry adiabatic BD intersecting the curve at R so that the area PAQ is equal to the area QRB then the area PADRQP is equal to

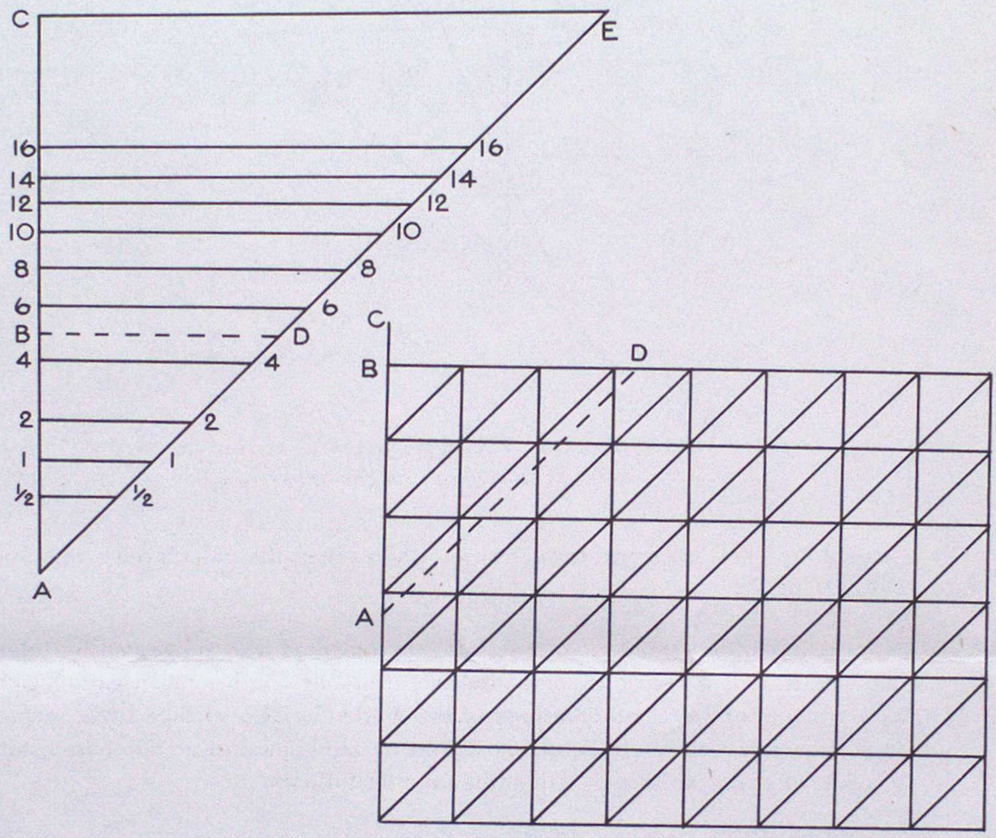

FIGURE 2 *Two scales for assessing areas on the tephigram*

the area ADRBQA. The curve DRS will approximate to the ascent curve after energy equivalent to the triangular area ADB has been added to the atmosphere, that is, D will be a measure of the surface temperature. The triangular scale is applied as follows. Place AC on the tephigram parallel to the isotherms and AE along the surface pressure isobar. Select the required value for BD and slide the scale laterally maintaining AE on the surface isobar until the area PAQ is equal to BQR. Then D is the required temperature. The method of using the rectangular scale is similar. Select the diagonal AD corresponding to the required area and place it along the surface pressure isobar. Adjust the scale always maintaining this diagonal on the surface pressure isobar and AC will lie along the isotherm. Adjust the scale so that the area PAQ equals BQR. Then D is the required temperature.

It is readily apparent that Gold's method is not a rigorous solution of the heat balance in the lowest layers of the atmosphere on clear days during daytime heating. Yet experience shows that his values yield results which are practical, useful and reasonably consistent in day-to-day forecasting. Some qualitative allowance can be made when the ground is unusually wet or very dry leading to a greater or smaller loss of sensible heat in the evaporation of water. In a pro-

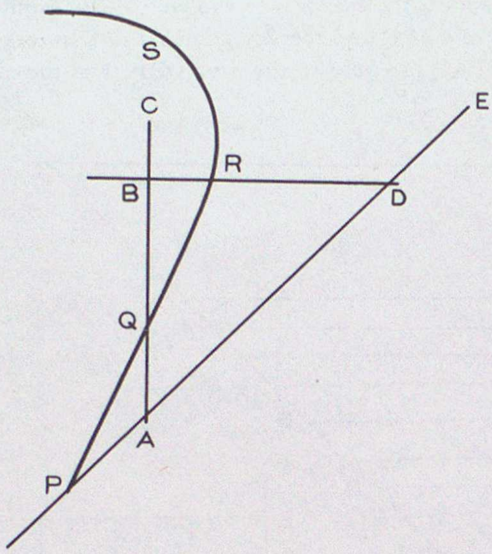

FIGURE 3 *The application of scales for assessing areas on the tephigram*

longed dry spell maximum temperatures may exceed the calculated maximum by some 2° or 3°F.

Another widely used adaptation of Gold's method was suggested by Johnston/ and modified as a result of comments by Jefferson*. In his paper Gold[20] gave assessments of the monthly values of the thickness of a surface layer, expressed in millibars, which would be changed from an isothermal to an adiabatic state as a result of solar radiation. His values are as follows:

Thickness of surface layer changed from an isothermal to an adiabatic state

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.
mb											
60	80	95	110	120	125	120	110	100	85	60	50

Johnston's suggested method of using these figures is:

(i) Modify the lowest part of the tephigram to approximate to conditions near dawn.

(ii) Draw in the surface pressure line and the pressure line at a thickness corresponding to the required month.

/ *Met. Mag., London,* 87, 1958, p.265.

* *Met. Mag., London,* 88, 1959, p.151.

(iii) Draw an isothermal between the two pressure lines cutting the modified environment curve so that the areas between the isothermal and the environment curve are equal.

(iv) Draw a dry adiabatic downwards from the point where the isothermal cuts the upper pressure line till it cuts the surface pressure line. The temperature at this point is the forecast maximum temperature for the day.

14.7.1.2. Temperature rise on clear mornings. The application of values from Table 3 to the relevant morning upper air ascent will enable an estimate to be made of the rate of rise of screen temperatures on a clear morning. Jefferson[22] also extended Gold's method to obtain the rise of temperatures. He found from an examination of actual cases at Northolt the times at which a given amount of energy had been supplied to the lowest layers. The basis of his work is described below.

During a period of two years (January 1945 - December 1946) days which fulfilled the following conditions were selected:

(i) Low cloud amount less than 7/10 between sunrise and noon or only increases above this value for one hourly observation.

(ii) No medium or high cloud or not more than 3/10 for more than one hour.

(iii) No rain reported during the previous 12 hours.

(iv) Visibility not falling below 1,000 yards for more than 3 hours and no thick fog at any time.

From an examination of the upper air data available for each selected day an estimated ascent for the early morning for Northolt was made and the lowest hourly temperature at Northolt for that morning was plotted as an additional surface reading. Having this estimated ascent, the next step was to find for each day the temperature which would be reached when the energies corresponding to 0.25, 0.5, 1 and 2 square centimetres on the tephigram* have been supplied to the lowest layers. From the actual hourly temperatures observed at Northolt

* Jefferson's values refer to A.M. Form 2810 (1953 edition). Equivalent areas on the three scales of tephigram issued since 1955 have been shown in Table 4.

Jefferson estimated to the nearest tenth of an hour the time in hours from sunrise at which these temperatures were actually attained. From these times the average times for each month for each energy equivalent were computed and used to draw smoothed curves showing monthly averages. From these curves Jefferson obtained for practical use at Northolt the corrected mean monthly values shown in Table 4.

TABLE 4 *Average time in hours from sunrise*

	Equivalent areas on tephigrams cm.²			
~~Form 2810 1955 edition and~~. Large scale insert on Met Forms 2810A 2810B 1956 edition . . .	0.49	0.97	1.95	3.9
Met Form 2810 19~~56~~63 edition . .	~~0.25~~ 0.27	0.54	1.09	~~2.0~~ 2.18
Small scale section on Met Forms 2810A 2810B 1956 edition . . .	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	...

For practical use Jefferson described a scale similar to the triangular scale in Figure 2.

The method of use at Northolt was:

(i) From Table 4 and the morning ascent curve forecast temperatures were computed when the various amounts of energy had been supplied.

(ii) Using these values and the forecast day maximum a graph of the rate of rise of temperature was constructed.

The data in this paragraph were determined for Northolt. The extent to which they need modification for other inland areas must be determined from examination of local data. A somewhat shorter method for determining suitable constants for other areas is suggested by Mr. Jefferson when commenting on some remarks by Mr. Inglis.* For stations on or near the coast considerable care and discretion should be used in attempting to determine mean values or apply the results to practical forecasting when the wind is on shore.

14.7.1.3. Forecasting temperatures under cloudy skies. The presence of moderate or large amounts of cloud materially effects the day maximum temperatures. Even a layer of quite thin and high cirrostratus materially reduces the incoming solar radiation and lower and thicker cloud sheets have a very marked effect. Although numerical values are not yet calculated for application to forecasting it is useful to know the magnitude of these effects. For thin cirrostratus and altostratus clouds there seems to be fairly general agreement that the albedo is of the order of 0.4 and for dense cloud of extensive area and great depth the albedo approaches 0.8. Thus the heat available to raise surface temperatures is

* *Met. Mag., London,* 75, 1950, p. 236.

where, T_{min} is the night minimum temperature, T_1' is the wet-bulb temperature at time of day maximum, D is the dew-point at time of day maximum and C is the correction for wind speed and cloud amount, which varies according to locality.

Boyden obtained some curves for Kew Observatory from which allowances could be made for mean low cloud amounts, wind speeds and occasions of fog.

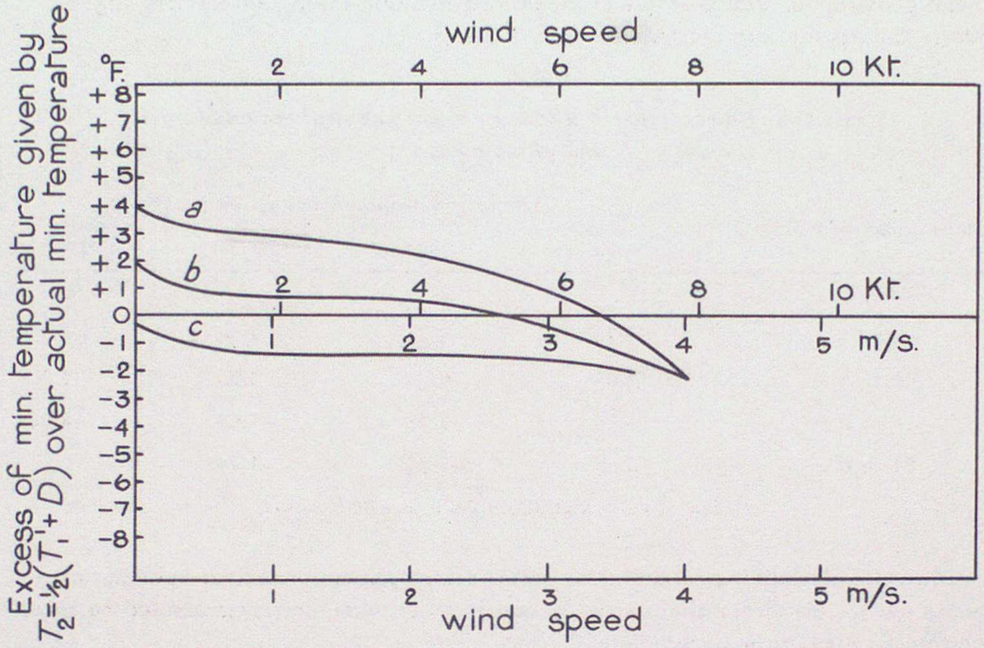

FIGURE 4 *Diagram for predicting minimum temperature allowing for cloud amount and wind speed*

a: no fog; mean low cloud amount < 2 oktas. *b*: no fog; mean low cloud amount 2-6 oktas. *c*: no fog; mean low cloud amount > 6 oktas. To obtain the final estimate for minimum temperature subtract from $\frac{1}{2}(T_1' + D)$ the ordinate of the appropriate curve at the estimated wind speed. If fog is forecast subtract a further 3.5°F.

His curves are reproduced in Figure 4. The method is simple to use as follows:

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

(ii) Determine night minimum T_{min} from formula $\frac{1}{2}(T_1' + D)$ + constant (a correction for wind and cloud).

(iii) If T_{min} is less than D as determined in (ii) take night minimum as T_{min}−3.5°F.

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 Figure 4 apply and to determine the value of the constant by comparison of predictions with sufficient number of actual observations of night minima. Boyden also suggests that the correction for fog may be less than 3.5°F. at most stations as the thermometers at Kew are exposed in a screen on a north wall at a height of 10 feet which differs somewhat from the standard exposure.

(*b*) Craddock and Pritchard's[27] formula was based on a statistical investigation covering 16 stations in England which were not close to the sea. The regression equation is:

$$T_{min} = 0.316T_{12} + 0.548D_{12} + 2.12,$$

where T_{min} is the night minimum temperature, T_{12} is the observed temperature at 1200 hours and D_{12} is the observed dew-point at 1200 hours. An allowance for mean geostrophic wind speed and cloud amounts is made by subtracting algebraically the appropriate corrections from Table 5.

TABLE 5 *Corrections to allow for mean geostrophic wind speed and cloud amounts*

*Mean cloud amount**	*Mean geostrophic wind speed (kt.)**			
	0–12	*13–25*	*26–38*	*39–51*
oktas		°F.		
0–2	4.26	1.74	1.14	–1.85
2+ to 4	2.83	0.20	0.21	–5.0
4+ to 6	1.33	–1.58	–1.00	–4.00
6+ to 8	0.05	–2.31	–1.74	...

*Mean of values at 1800, 2400 and 0600 hours

Craddock's formula is useful for forecasting temperatures over substantial areas but for its direct application to one locality some correction based on local performance might be appropriate.

(*c*) McKenzie's[28] formula is:

$$T_{min} = \tfrac{1}{2}\,(T_{max} + D) + C,$$

where T_{max} is the maximum temperature, D is the average dew-point of the air mass expected during the night, C is the constant for each station for given cloud amount and wind.

From a large number of suitable examples at Dyce, McKenzie found the constants for use with varying cloud amounts and wind speeds. His original table was given for each tenth of cloud but these have been converted in Table 6 to intervals of 2 oktas as this seems unlikely to alter the values by more than a fraction of 1°F. All constants were found to be negative.

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 2° or 3°F. from 1200 G.M.T. one day to 0600 G.M.T. the following day. From the main observations over that period calculate the mean dew-point, the average low 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 6, using average values of the constants found for each combination of cloud amount and wind speed.

TABLE 6 *Values of constants (C) for Dyce*

Average wind speed		*Average low cloud cover (oktas)*				
Force	*Speed*	0	2	4	6	8
	kt.			°F.		
0	0	−15	−11	−11	−8	−8
1	1-3	−13	− 9	− 9	−7	−5
2	4-6	− 8	− 7	− 6	−5	−3
3	7-10	− 7	− 6	− 5	−5	−3
4	11-16	− 6	− 5	− 5	−3	−2
5	17-21	− 4	− 4	− 4	−3	−2
6	22-27	− 4	− 3	− 3	−3	−2

There are several other formulae for forecasting night minima. An important part of the application of these empirical rules is to use them regularly and consistently so that experience is gained in the technique and limitations of the rules. The rules should not be applied blindly but with discretion and judgment. With wide and long experience they will yield good results. The main errors are likely to arise from difficulties in forecasting cloud cover and wind speed.

14.7.2.2. Cooling under clear skies. Saunders has 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[29] work rests on a discontinuity in the rate of cooling at grass level which occurs around 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 United Kingdom on clear evenings at times varying between 1600 and 2230 G.M.T. according to the time of the year. Saunders' earlier work was done at Abingdon and reproduced in Figure 5 are screen thermograms for Abingdon for 24-26 March 1944. The discontinuity at 2000 G.M.T. is very clear on all three evenings.

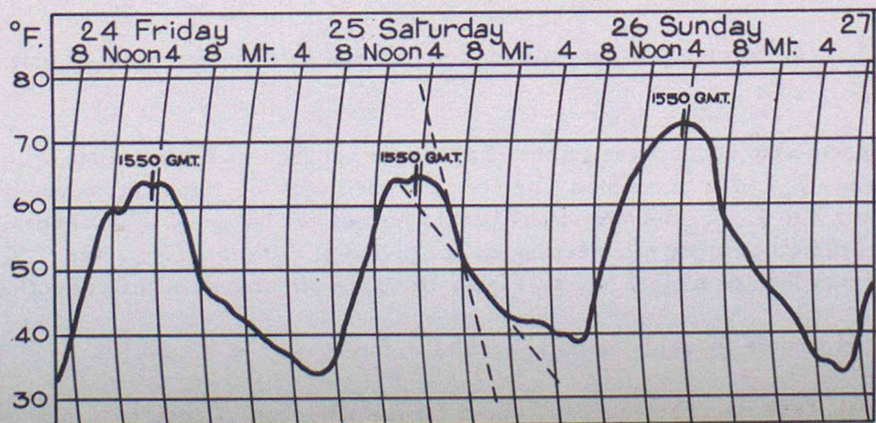

FIGURE 5 *Screen thermograms for Abingdon, 24-26 March 1944*

The investigation was continued at Northolt during the period August 1948 to April 1950. Saunders[30] was able to construct a graph showing the time of discontinuity in the rate of cooling throughout the year. His graph is reproduced in Figure 6. This graph enables an estimate to be made of the time of the dis-

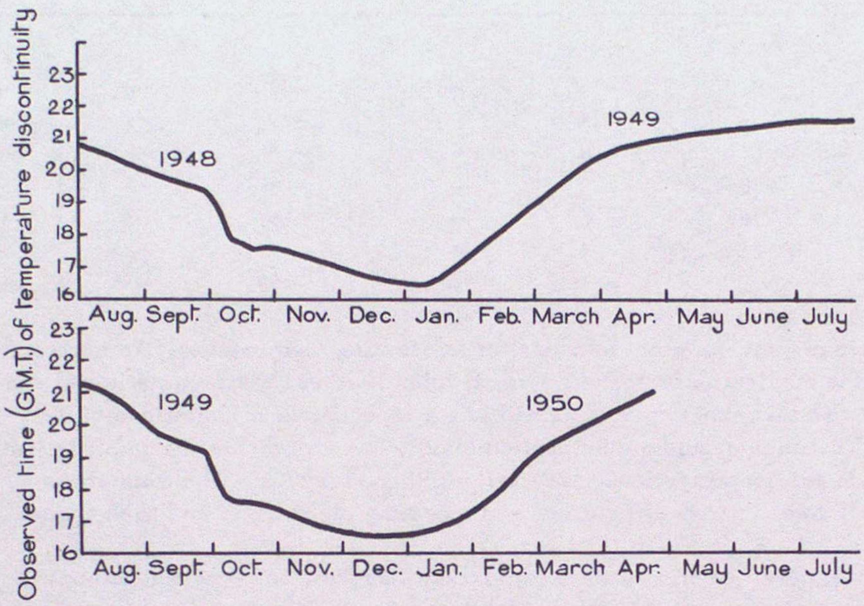

FIGURE 6 *Variation of observed time of discontinuity in rate of fall of temperature with time of year at Northolt*

continuity at Northolt for any day of the year. The sudden change in time in the first week of October is noteworthy. Saunders attributes this to the change from dry to wet soil.

Saunders developed two regression equations which enable the temperature T_r at the time of the discontinuity to be calculated according to whether there was or was not an afternoon inversion with base at or below 900 millibars.

With no inversion $T_r = \frac{1}{2}(T_{max} + D) - 0.6°F$.

With inversion $T_r = \frac{1}{2}(T_{max} + D) - 4.0°F$.

where T_{max} is the maximum dry bulb temperature and D is the dew-point at time of T_{max}.

Saunders also constructed graphs showing the subsequent cooling after temperature T_r had been reached. He compiled the following graphs reproduced in Figures 7 to 9. For the midsummer part of the summer half-year it was necessary at Northolt to apply a correction, as the period of cooling after T_r was reached was less than eight hours. Figure 10 shows the amount of the correction.

The procedure for applying these results to forecasting at Northolt is:

(i) Take the observed T_{max} and D for the afternoon. Consider if they are representative of the air likely to be over the locality during the night. If not, make an estimate of appropriate values from observations upwind.

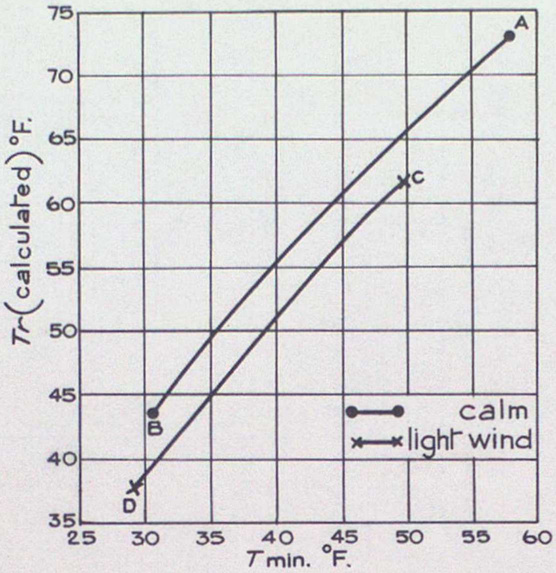

FIGURE 7 *Relation between initial and final temperatures of the period of subsequent cooling during the spring and late summer*

Calm: mean geostrophic wind 0-12 kt.
Light wind: mean geostrophic wind 13-18 kt.

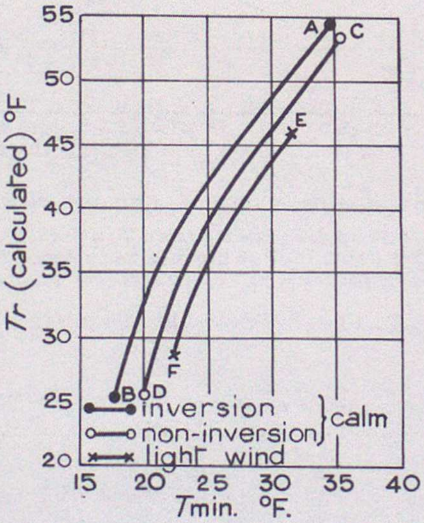

FIGURE 8 *Relation between initial and final temperatures of the period of subsequent cooling during the winter*

Calm: mean geostrophic wind 0-16 kt.
Light wind: mean geostrophic wind 17-21 kt.

(ii) Estimate synoptically, allowing for trend and advection, whether or not there will be an inversion at or below 900 millibars during the afternoon.

(iii) Calculate T_r from the appropriate regression equation.

(iv) Estimate the time of T_r from Figure 6. For wet topsoil subtract one hour in late spring and early summer and half an hour in late summer.

(v) From the forecast T_r and the forecast mean geostrophic wind speed use Table 7 to obtain T_{min}.

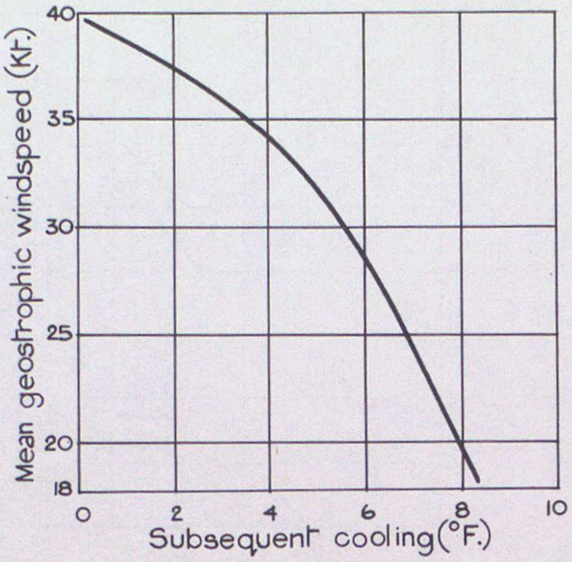

FIGURE 9 *Relation between the number of degrees of subsequent cooling and the mean geostrophic wind speed for occasions of stronger wind*
Lower limit of geostrophic wind speed 19 kt. in summer and 22 kt. in winter.

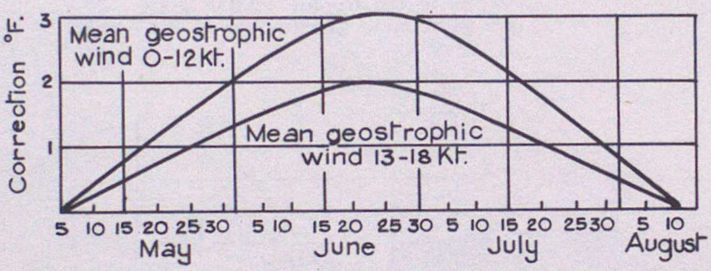

FIGURE 10 *Correction for reduced length of period of subsequent cooling*
The correction is applied when Figure 7 is used during the midsummer period. Note that if T_r is reached early owing to moist topsoil this correction will be partially or wholly offset.

TABLE 7 *Method for obtaining* T_{min}

Season	*Forecast mean geostrophic wind*	*Method for* T_{min}
	kt.	
Summer (late March - early October)	0-12	Fig. 7, curve AB } Add to T_{min} a time correction from Fig. 10 during 5 May to 10 August except after rain.
	13-18	Fig. 7, curve CD }
	over 18	By $T_{min} = T_r$ (calculated) $- \Delta$°F. where Δ is given by Fig. 9.
Winter (early October - late March)	0-16	Fig. 8, AB or CD according to afternoon lapse rate.
	17-21	Fig. 8, EF.
	over 21	As in summer for winds over 18 kt.

This work applies to Northolt with a clay soil and Saunders believes that the times given in Figure 6 may be applicable to clay soils elsewhere. Regarding values for T_{min} at other localities on other types of soil separate corrections must be deduced from a sufficient number of cases for each section of Table 7. Special techniques will be needed at coastal stations except with a calm wind or wind off the land. Saunders' technique has aroused considerable interest and a number of articles in the *Meteorological Magazine* on the application of this method in various localities are listed in the Bibliography.[31-37]

14.7.2.3. *Cooling under cloudy skies.* With appreciable cloud amounts there is a material increase of radiation at night towards the earth as clouds absorb and radiate approximately as black bodies. The radiation from a cloud towards the earth has a relatively large effect on the radiation balance at night. As would be expected cooling due to radiative effects is less under cloudy than under clear skies. The three formulae listed for computing night minimum temperature contain corrections for cloud amount and may be used to forecast minimum temperatures under cloudy skies.

Summersby[38] extended Saunders work on night minima at Northolt by determining the correction to be applied for various cloud amounts. Summersby's work

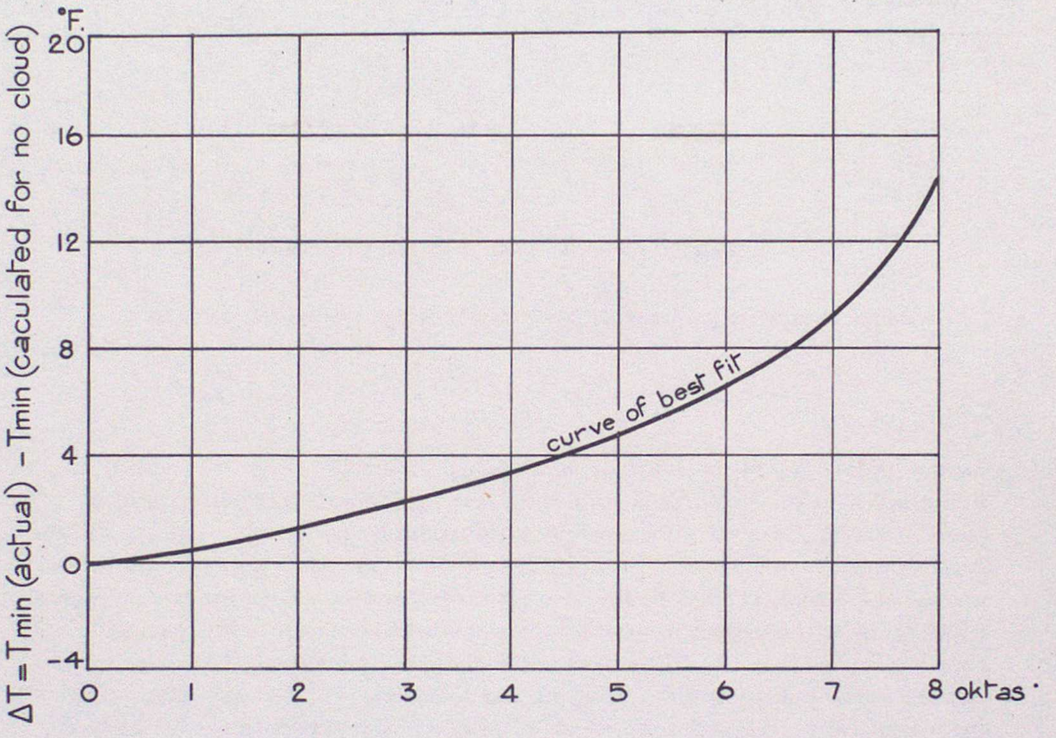

FIGURE 11 *Relation between average cloud amount and decrease in cooling*

excludes nights with only cirrus cloud, nights with any precipitation and nights with frontal passage, however weak, or with an advective change of dew-point. His corrections are shown in Figure 11. He suggests a formula of the form

$$\Delta T = C_1 \left\{ \exp (C_2 A) - 1 \right\},$$

where C_1 and C_2 are constants and A is the arithmetic mean cloud amount in oktas during the night. For Northolt he found

$$\Delta T = 2 \left\{ \exp (A/4) - 1 \right\}$$

and suggests that these constants would probably be satisfactory at stations with similar subsoil and topography.

14.7.3. *Grass minima and ground frost*

Forecasts of minimum air temperatures are important for a number of purposes. Another minimum temperature which is also of great economic value is the grass minimum temperature - particularly when this temperature is likely to fall to 30.4°F. or below, that is, there will be a ground frost. Experience of meteorological observations soon shows that the grass minimum temperature is generally several degrees below the screen minimum temperature. An estimate of this difference may be made from the following set of rules.

Craddock and Pritchard[27] in their statistical investigation used actual values for 38 cases from 16 stations in England for the difference ($T_{min} - G_{min}$) where G_{min} = grass minimum temperature. Their table relating screen and grass minima is reproduced as Table 8.

TABLE 8 *Mean value of ($T_{min} - G_{min}$) under stated conditions*

Mean cloud amount*	*Mean* geostrophic wind speed (knots)* 0 - 12	13 - 25	26 - 38	39 - 51
oktas		°F.		
0 - 2	7.2	7.6	5.9	4.8
2 - 4	8.8	6.7	5.6	4.4
4 - 6	6.7	5.4	4.7	4.2
6 - 8	7.2	3.7	4.5	5

*Mean of values at 1800, 2400 and 0600 hours

In an investigation at Munster, R. Faust[39] has proposed the formula

$$T_{14} + \frac{D_{14}}{2} < 79$$

for the forecasting of ground frost on radiation nights for which it is specified that cloud should be less than two tenths and wind speed less than Beaufort force 2 (that is, less than 4 knots) during the cooling period. T_{14} and D_{14} are the screen temperatures and dew-point (°F.) at 1400 local time. Munster is a lowland station and Jefferson[40] tested the formula at Hullavington for all radiation nights which fulfilled the requirements during the period September 1950 - June 1951. From a plot of ($T_{14} + D_{14}/2$) against grass minimum temperature Jefferson found that the curve intersected the 32°F. isotherm at about 79. James[41] also tested the formula at St. Athan for the whole of 1945 and the period November 1949 to May 1951. He found a constant of 78. St. Athan is two miles from the Welsh coast and 150 feet above mean sea level. Hullavington is situated on the south-east of the Cotswold Hills and is some 340 feet above mean sea level.

From the results it would seem that for radiation nights as defined, the formula could be applied suceessfully to most lowland stations in reasonably level country with fair success. 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 ($T_{14} + D_{14}/2$) as high as 82. Discretion should be used and the forecast should be carefully phrased when ($T_{14} + D_{14}/2$) is only slightly above the critical value.

Saunders[30] investigated the relation between screen minimum (T_{min}) and the grass minimum (G_{min}) on clear nights (cloud $\leqq 1/8$) with no fog at Northolt for August 1948 to April 1950. He compiled Tables 9 and 10. 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. This may affect both grass minimum and screen temperature, particularly the former. On such occasions minima may be several degrees higher than otherwise expected.

Some of the variations in Table 10 probably arise from this cause.

By the application of Saunders' technique to the forecast screen minimum forecasters can make estimates of the grass minimum for radiation nights. The formula

$$T + \frac{D}{2} < 79$$

gives a criterion for estimating the occurrence of night frosts and Table 7 enables an estimate to be made for varying cloud amounts. From the several values obtained a reasonably accurate forecast should be possible.

TABLE 9 *Variation of difference between screen and grass minimum temperatures with wind speed*

Season	*Geostrophic wind speed*	*Mean* $T_{min} - G_{min}$	*Standard deviation*
	kt.	°F.	°F.
Summer	0 - 12	8	1.5
Winter	0 - 16	Vary in accord with Table 10	
Summer	13 - 24	11.1	2.5
Winter	17 - 24		
Whole year	Over 24	7.4	1.3

TABLE 10 *Variation of difference between screen and grass minimum temperatures with temperature for occasion of calm or light winds during winter*

T_{min}	*Mean* $T_{min} - G_{min}$
°F.	°F.
31 - 35	4
26 - 30	7
21 - 25	9
16 - 20	11

Grass minimum temperatures are measured over close-cropped grass and this is the temperature which is forecast. It would be expected that temperatures over other types of surface would be somewhat different. Gloyne[42] examined observations made over short turf and bare soil at Starcross, Devon, during the period January 1949 - December 1950. The site was 29 feet above mean sea level, a few hundred yards from the west bank of the Exe estuary and the soil was sandy loam with a good deal of alluvial silt. On the vast majority of nights it was found that the screen minimum was greater than the bare soil minimum which in turn was greater than the grass minimum. For radiation nights (which were specified as nights on which the grass minimum was 7°F. or more below the screen minimum) Gloyne found the mean values in Table 11. The period over which these values were obtained is quite short but it gives an indication of the order of magnitude of the excess of bare soil minima over grass minima for radiation nights.

TABLE 11 *Values on radiation nights at Starcross, Devon (1949 and 1950) of means of bare soil minimum minus grass minimum*

	Jan.	Feb.	March	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
1949	2.6	3.7	4.4	5.2	4.1	4.9	4.4	3.5	3.9	4.1	2.3	2.0
1950	...	1.5	2.4	3.9	4.1	4.9	3.9	4.5	3.7	4.6	3.6	3.8

14.7.4. *The effect of surface wind*

The existence of a moderate surface wind reduces diurnal temperature changes owing to the fact that mechanical turbulence is always associated with surface winds to a greater or lesser degree. By night, and in the absence of wind, radiative cooling in the lowest layers will tend to establish at least a stable lapse rate even if not an inversion of temperature. Under stable conditions and according to the formula in Section 14.5 mechanical turbulence will transport heat downwards which will offset the radiative cooling to some degree. Contrariwise by day when insolation tends to establish unstable lapse rates mechanical turbulence will then transport heat upwards which effectively distributes the surface heating through a greater vertical depth. The effect of surface wind is incorporated in many of the values given in Section 14.7.2.

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 and the effect of the surrounding seas on air currents reaching the British Isles is dealt with at some length in Section 14.8.1. In régimes 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 Section 14.10.1.

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.

14.7.4.1. *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. This causes pressure to rise at upper levels 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 sea-breeze inland is variable and for a discussion of this the reader is referred to Chapter 13.

The effect of sea-breezes on temperatures is also quite variable. In an investigation at Worthy Down, Peters[43] found that about half the sea-breezes were accompanied by a sudden drop in temperature of 1° - 4°F. at their onset or an abrupt commencement of the usual evening fall of temperature. (Worthy Down is about 25 miles inland from the Solent and Southampton Water and nearly 40 miles from the southern extremity of the Isle of Wight. Sea-breezes attain a maximum frequency between 1600 and 1800 G.M.T. at Worthy Down.) He does 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 indentified at Worthy Down Peters compiled tables showing the differences in maximum temperatures, Worthy Down minus Calshot and Worthy Down minus St. Catherines Point, Isle of

TABLE 12 *Distribution of differences in day maxima at Worthy Down, Calshot and St. Catherine's Point for 58 occasions of sea-breeze*

	Temperature difference (°F.)																	
	11	+11	+10	+9	+8	+7	+6	+5	+4	+3	+2	+1	0	−1	−2	−3	−4	−5
	number of occasions																	
Worthy Down, Calshot	2	1	0	2	2	1	6	2	6	3	3	7	4	7	7	4	1	0
Worthy Down, St. Catherine's Point, Isle of Wight	5	2	5	2	9	6	7	5	6	4	2	3	1	1	0	0	0	0

Wight. Table 12, showing these values, has been included to indicate the order of magnitude and the variability of the effect of sea-breezes on day maximum temperatures. Calshot is sheltered to some extent by the Isle of Wight and this probably accounts for the appreciable number of occasions on which Calshot had a higher maximum than Worthy Down. St. Catherine's Point is well exposed to the English Channel and on 56 out of 58 occasions of sea-breeze it had a lower day maximum temperature than Worthy Down. The greatest positive differences in day maximum temperatures between Worthy Down and Calshot and St. Catherine's Point were respectively 18° and 21°F. The extent and nature of differences for other localities must be determined from an examination of records on suitable occasions of sea-breeze.

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.

14.7.4.2. The 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 that 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[44] and from some of the references therein. One of these is a book by Geiger (available in English) entitled *The climate near the ground*. This is a full and easily readable account of climate in the lowest two metres of the atmosphere.

14.7.5. 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 régimes 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 few hundred feet. 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 outgoing 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 the air just above the ground. The air in immediate contact with the ground is cooled and Best[15] 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 100 feet 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 Section 14.7.2.2). 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[45] 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 several hundred feet 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 régime 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[46] observed that the temperature at 12.4 metres began to fall rapidly (at 0045 G.M.T.) as the fog thickened. The fall amounted to 4°F. in 25 minutes and by 0130 G.M.T. the fall had been sufficient to convert the inversion into a lapse of temperature below 12.4 metres. At 0055 G.M.T. the temperature at the 30-metre level also began to drop rapidly and by 0130 G.M.T. it had fallen by about the same amount (4°F.). 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°F.

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[45] states that in deeper fogs the lapse rate usually seems close to the saturated adiabatic value (3.3°F. per 1,000 feet). 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 45, 46 and 47 in the Bibliography on page 85.

14.7.6. *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 100 miles 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 15 miles. 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 10°F. 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.[48]

14.8. DISTRIBUTION OF TEMPERATURE IN THE FREE ATMOSPHERE IN THE NEIGHBOURHOOD OF THE BRITISH ISLES.

A knowledge of synoptic climatology and of atmospheric models is a valuable part of a forecaster's equipment in assessing current data and making short-period forecasts. Many forecasters gain such a background by long experience of day-to-day analysis and forecasting. Such a background has essentially a personal bias. It seemed desirable therefore to survey published literature on synoptic climatology and extract such data, tables or figures which should be of value when forecasting temperatures. The information in the following sections should be of value to the experienced forecasters and particularly so to those with little experience.

Broadly speaking, knowledge of distribution of temperature in the troposphere and lower stratosphere becomes less detailed with increasing height. The amount of detail in the following sections therefore varies considerably. Belasco[49] has examined much upper air data and made an extensive study of the variations in temperature and humidity of air masses affecting the British Isles. The following section is based entirely on his work.

14.8.1. *Temperatures in air masses over the British Isles*

Belasco divided the air masses affecting the British Isles into nineteen classes and details of his classification are given in Table 13. In some of the succeeding tables, figures and diagrams details are included for tropical maritime (T_A) and polar (P_{IC}) air masses when in their source regions (namely, south-west of the Azores and north and north-east of Iceland) so that modifications to the structure of the air during its journey to the British Isles can be assessed. Figures 12, 13 and 14 show the generalized tracks of air from the different sources.

TABLE 13 *Classification of air masses affecting the British Isles*

Air mass	*Sub-division symbol*	*Source region*	*Main curvature of path to British Isles*	*Nature of path surface*	*Direction of approach to British Isles*
Tropical maritime	T_A*	South-west of Azores	Anticyclonic	...	...
Tropical maritime	T_1	South-west of Azores	Straight	Oceanic	SW.
Tropical maritime	T_2	South-west of Azores	Anticyclonic	Oceanic	S. or SW.
Tropical continental	T_3	Spain, Mediterranean or north-west Africa	Straight	Land and English Channel	SE. or S.
Tropical continental	T_4**	Southern Europe in summer	Anticyclonic	Land and English Channel	E. or SE.
Quasi tropical maritime	T_Q	43-50°N., 15-25°W.	Straight or anticyclonic	Oceanic	SW., W. or NW.
Indeterminate anticyclonic air in and near the central region of an anticyclone	H_O	...	Anticyclonic	Mainly land	Indefinite
	H_{NE}**	...	Anticyclonic	Land and North Sea	NE. - E.
	H_{SE}**	...	Anticyclonic	Land and eastern English Channel	SE. - S.
	H_{SW}**	...	Anticyclonic	Land and western English Channel	SW. - W.
	H_{NW}**	...	Anticyclonic	British Isles and oceanic	NW. - N.

* These classes of air mass, at their source region, are included for comparison with corresponding air masses reaching the British Isles.

** Surface classes only.

TABLE 13 (*Continued*)

Air mass		*Sub-division symbol*	*Source region*	*Main curvature of path to British Isles*	*Nature of path surface*	*Direction of approach to British Isles*
Polar	continental	A_1	North of 50°N. and east of 25°E.	Cyclonic	Land and North Sea	NE., E. or SE.
		A_2		Anticyclonic		NE., E. or SE.
		P_{IC}*	North and north-east of Iceland	Cyclonic	...	...
	maritime	P_1	North and north-east of Iceland	Cyclonic	Oceanic	N. or NE.
		P_2		Anticyclonic		N. or NE.
		P_3	North-west and west of Iceland	Cyclonic		NW.
		P_4		Anticyclonic		NW.
		P_5		Cyclonic		W.
		P_6		Anticyclonic		W.
		P_7		Cyclonic		S. or SW.

* These classes of air mass, at their source region, are included for comparison with corresponding air masses reaching the British Isles.

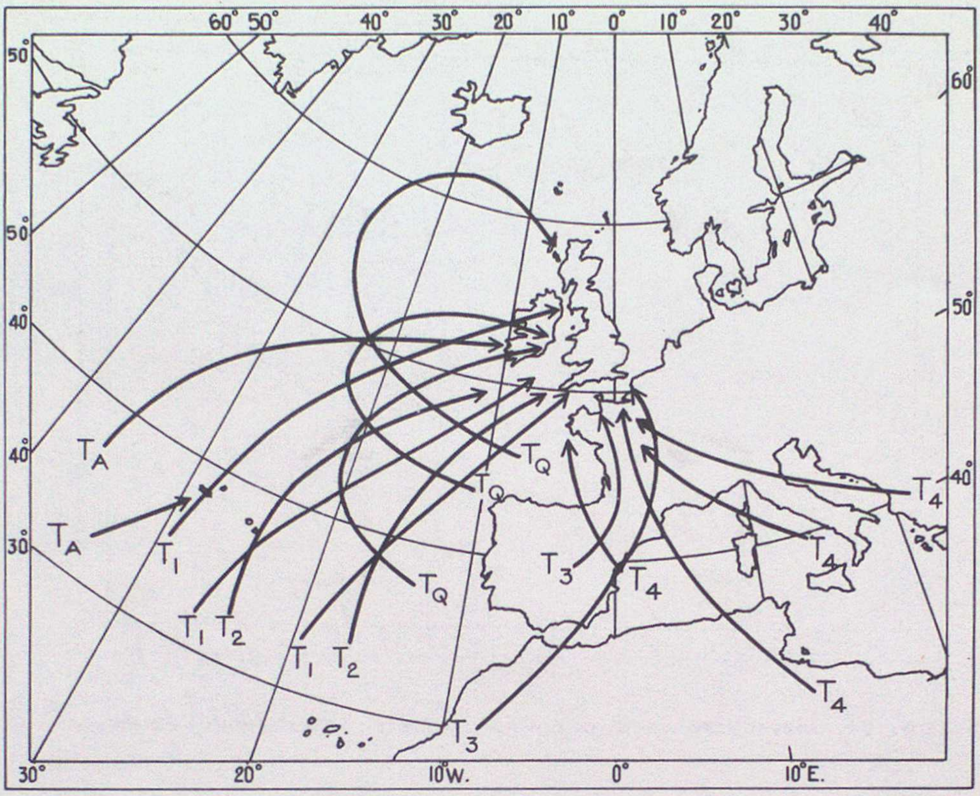

FIGURE 12 *Generalized tracks of the tropical and quasi-tropical air masses*

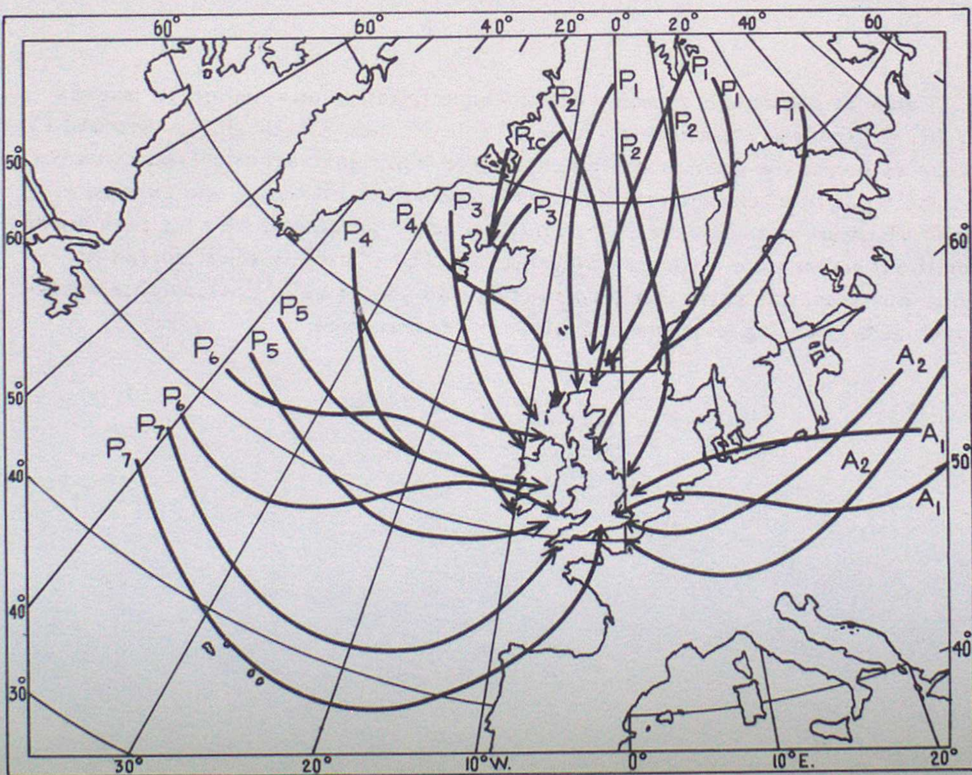

FIGURE 13 *Generalized tracks of the polar air masses*

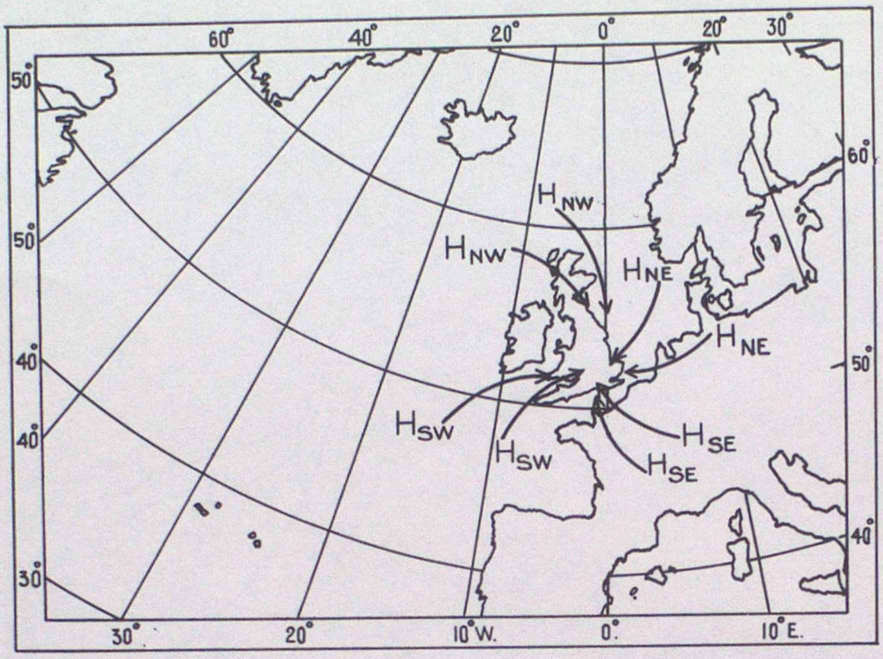

FIGURE 14 *Generalized tracks of the anticyclonic (indeterminate) air masses*

Table 14 contains a summary of the characteristics of selected air masses at 700 millibars in summer and winter. Table 15 shows for 16 air masses and 11 pressure levels the mean monthly temperature, the mean vapour pressure and the wet-bulb potential temperature for summer and winter. Average and extreme daily maximum and minimum temperatures near the surface at Kew for each month for 18 air masses are contained in Table 16. The relatively small spread for some air masses is particularly noteworthy and should be of considerable assistance in forecasting maximum and minimum temperatures.

TABLE 14 *Summary of the characteristics of the selected air masses at 700 millibars in summer and winter*

Summer

Air mass	Temperature Mean	Temperature Extreme Max.	Temperature Extreme Min.	Mean vapour pressure	Mean wet-bulb potential temperature	Mean partial thickness 1000-700mb.
	°F.	°F.	°F.	mb.	°F.	m.
T_A	46	50	42	4.9	63	2,981
T_1	41	45	37	5.0	61	2,963
T_2	44	49	41	4.3	61	2,972
T_3	43	49	40	5.2	62	2,996
T_Q	37	42	31	5.0	60	2,944
H_O	41	44	36	3.8	58	2,941
A_1	...	...	...	...	...	...
A_2	...	...	...	...	...	...
P_{IC}	11	14	6	1.9	42	2,816
P_1	19	22	14	2.8	47	2,868
P_2	19	23	17	2.6	47	2,971
P_3	21	26	18	3.0	49	2,983
P_4	24	27	20	3.3	51	2,902
P_5	25	28	21	3.6	51	2,908
P_6	27	30	25	4.0	53	2,917
P_7	29	32	26	4.3	55	2,923

Winter

Air mass	Temperature Mean	Temperature Extreme Max.	Temperature Extreme Min.	Mean vapour pressure	Mean wet-bulb potential temperature	Mean partial thickness 1000-700mb.
T_A	34	38	29	3.6	56	2,926
T_1	29	34	25	3.9	55	2,880
T_2	33	37	30	3.3	55	2,899
T_3	25	30	21	3.8	53	2,874
T_Q	24	29	19	3.2	51	2,868
H_O	26	31	21	2.9	51	2,865
A_1	− 8	−4	−13	0.7	28	2,731
A_2	− 3	−1	− 4	0.8	31	2,752
P_{IC}	−14	−7	−22	0.5	23	2,658
P_1	− 4	1	−10	0.8	30	2,749
P_2	0	2	− 5	1.0	32	2,771
P_3	3	7	0	1.3	37	2,786
P_4	5	10	− 1	1.4	37	2,795
P_5	8	12	4	1.5	40	2,816
P_6	10	14	8	1.6	40	2,825
P_7	16	20	12	2.6	46	2,847

TABLE 15 *The mean monthly temperature and the mean vapour pressure and wet-bulb potential temperature for summer and winter for 16 air masses and 11 pressure levels*

Pressure	Monthly means of temperature												Summer (July and August)		Winter (January and February)	
	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Mean vapour pressure*	Mean wet-bulb potential temp.	Mean vapour pressure*	Mean wet-bulb potential temp.
mb.	°F.												mb.	°F.	mb.	°F.
T_A	Tropical maritime air in the Azores region															
450	−5	−6	−6	−3	0	5	9	10	8	4	1	−3	0.7	65	0.5	58
500	6	3	4	6	9	15	18	19	17	13	10	8	1.2	64	0.8	57
550	14	12	13	15	17	22	26	27	24	22	19	17	2.1	64	1.2	57
600	21	20	21	22	25	30	33	35	32	29	27	24	2.8	64	1.7	56
650	29	27	27	29	32	36	39	41	38	36	34	31	3.8	63	2.7	56
700	35	34	33	36	39	43	45	47	45	43	40	37	4.9	63	3.6	56
750	41	39	39	43	45	48	50	52	50	48	46	43	5.9	62	4.7	56
800	45	43	44	48	50	53	54	56	55	52	50	48	8.0	62	6.1	56
850	48	47	49	52	54	56	57	59	57	56	54	51	10.5	63	8.3	56
900	51	50	52	56	57	58	59	61	60	59	57	53	13.0	63	10.1	55
950	55	54	55	57	59	61	62	64	62	61	59	56	15.9	63	10.6	54
T_1	Tropical maritime air with a straight path from south-west of the Azores															
450	−9	−11	−11	−8	−5	−1	3	5	3	−1	−4	−7	1.2	64	0.6	57
500	1	−1	−1	1	5	9	12	14	12	9	5	2	1.9	64	1.0	57
550	9	8	7	9	13	17	20	22	20	17	13	11	2.7	64	1.5	56
600	16	16	15	17	20	25	27	30	27	24	20	18	3.6	63	2.0	56
650	23	22	22	23	27	31	34	36	33	30	27	25	4.3	62	3.0	55

*Below the freezing point the vapour pressure is that over ice.

TABLE 15 (*Continued*)

Pressure	Monthly means of temperature												Summer (July and August) Mean vapour pressure*	Summer (July and August) Mean wet-bulb potential temp.	Winter (January and February) Mean vapour pressure*	Winter (January and February) Mean wet-bulb potential temp.
	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.				
mb.	°F.												mb.	°F.	mb.	°F.
T_1	Tropical maritime air with a straight path from south-west of the Azores (Continued)															
700	29	28	28	30	33	37	40	42	39	36	33	31	5.0	61	3.9	55
750	34	34	34	36	39	42	45	47	44	41	39	36	6.2	60	4.9	54
800	39	39	39	41	44	47	50	52	49	46	44	41	8.3	61	6.2	54
850	42	44	44	45	47	50	54	55	53	50	48	45	11.0	62	7.7	53
900	45	47	47	48	51	54	57	58	56	53	51	48	13.5	63	9.3	53
950	48	48	49	50	52	57	60	59	58	55	52	49	14.9	61	10.4	51
T_2	Tropical maritime air with an anticyclonic path from south-west of the Azores															
450	− 7	− 7	− 8	− 5	− 1	4	7	9	7	2	− 1	− 5	1.0	65	0.6	58
500	4	3	3	5	8	14	16	18	16	12	8	6	1.6	65	1.0	57
550	13	12	12	13	16	22	24	26	24	22	17	14	2.3	64	1.4	57
600	20	20	19	21	24	29	31	33	31	27	25	22	3.1	63	1.9	56
650	27	27	26	28	30	35	38	39	37	34	32	30	3.5	62	2.6	56
700	33	34	33	35	37	41	44	45	43	40	38	37	4.3	61	3.3	55
750	39	39	38	40	42	47	49	50	48	45	43	42	5.5	61	3.8	54
800	43	42	43	44	47	51	53	54	51	49	48	46	7.6	61	4.8	53
850	46	45	47	48	50	53	55	56	54	52	51	49	9.6	61	6.1	52
900	46	46	48	48	52	55	58	58	57	54	52	48	12.0	60	8.3	51
950	46	47	47	48	53	59	61	61	59	54	51	48	14.1	60	9.6	50

*Below the freezing point the vapour pressure is that over ice.

TABLE 15 (Continued)

Pressure	Monthly means of temperature												Summer (July and August)		Winter (January and February)	
	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Mean vapour pressure*	Mean wet-bulb potential temp.	Mean vapour pressure*	Mean wet-bulb potential temp.
mb.	°F.												mb.	°F.	mb.	°F.
T_3	Tropical continental air with a straight path from Spain, Mediterranean, north-west Africa or southern Europe in summer															
450	−13	−15	−14	−11	− 7	0	3	4	1	− 3	− 7	−10	1.1	64	0.5	55
500	− 3	− 5	− 3	− 1	3	11	13	14	12	8	3	0	1.8	64	0.9	55
550	6	4	6	8	12	20	21	22	20	16	11	9	3.0	64	1.4	54
600	13	12	13	15	19	28	29	29	27	23	19	16	4.0	63	2.0	54
650	19	19	20	22	26	35	36	36	34	30	26	23	4.4	63	2.8	53
700	25	26	27	29	33	42	43	43	40	36	33	30	5.2	62	3.8	53
750	31	32	33	35	39	48	50	49	45	42	38	35	6.4	62	4.7	52
800	36	37	39	42	46	54	56	56	51	47	43	40	7.6	62	5.2	50
850	40	41	43	47	52	59	61	61	56	53	48	45	10.0	63	5.3	48
900	42	42	47	51	58	64	65	66	61	57	51	46	11.6	64	5.5	46
950	39	42	48	56	62	69	70	70	65	58	52	47	13.1	64	5.8	40
T_Q	Quasi-tropical air from 43-50°N., 15-25°W.															
450	−17	−18	−18	−14	−10	− 7	− 3	− 2	− 5	− 9	−13	−15	0.8	61	0.4	53
500	− 6	− 7	− 7	− 3	0	3	7	8	5	2	− 2	− 4	1.5	61	0.8	53
550	2	2	2	6	9	12	16	17	14	10	7	5	2.3	61	1.3	53
600	11	10	10	14	16	19	23	24	21	18	14	13	3.1	61	2.0	52
650	17	18	17	21	23	26	30	31	28	24	21	19	4.3	61	2.6	51

*Below the freezing point the vapour pressure is that over ice.

TABLE 15 (Continued)

Pressure	Monthly means of temperature												Summer (July and August)		Winter (January and February)	
	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Mean vapour pressure*	Mean wet-bulb potential temp.	Mean vapour pressure*	Mean wet-bulb potential temp.
mb.	°F.												mb.	°F.	mb.	°F.
T_Q	Quasi-tropical air from 43-50°N., 15-25°W. (Continued)															
700	23	24	23	27	30	32	36	38	34	31	27	26	5.0	60	3.2	50
750	30	29	29	32	35	37	43	43	39	36	33	32	6.0	60	3.8	49
800	32	33	34	37	40	43	47	48	44	41	38	36	8.0	61	4.6	48
850	35	36	39	41	44	47	51	52	49	44	41	39	9.9	60	5.5	46
900	39	38	43	45	48	51	54	55	51	47	44	42	11.7	59	6.5	44
950	40	39	44	48	51	55	59	60	56	50	46	43	13.3	58	7.3	42
H_O	Anticyclonic air															
450	−16	−17	−13	− 9	− 3	1	4	5	3	− 2	− 8	−13	1.0	64	0.3	53
500	− 5	− 6	− 3	2	7	11	14	15	13	10	2	− 3	1.6	63	0.4	53
550	4	4	7	11	16	19	22	23	21	18	11	6	2.4	63	0.6	53
600	11	12	15	19	23	27	29	30	29	26	19	13	3.2	62	1.3	53
650	18	19	22	25	30	33	35	36	36	33	26	20	3.4	60	2.3	53
700	24	26	28	31	36	40	41	41	42	39	32	27	3.8	58	2.9	51
750	30	32	33	36	41	45	46	46	47	43	37	32	3.9	56	3.0	49
800	34	36	38	41	44	48	49	50	50	46	40	35	4.4	55	3.1	46
850	39	39	40	44	47	51	53	54	53	48	41	37	7.0	56	3.7	44
900	39	40	41	46	48	54	57	57	54	48	41	39	9.9	57	3.7	41
950	38	39	42	47	49	58	61	60	56	49	41	38	11.1	57	5.5	39

*Below the freezing point the vapour pressure is that over ice.

TABLE 15 (Continued)

Pressure	Monthly means of temperature												Summer (July and August)		Winter (January and February)	
	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Mean vapour pressure*	Mean wet-bulb potential temp.	Mean vapour pressure*	Mean wet-bulb potential temp.
mb.	°F.												mb.	°F.	mb.	°F.
A_1	Polar continental air with a cyclonic track from north of 50°N. and east of 25°E.															
450	−52	−53**	..	..	..	..	..	..	..	..	−40	−44	..	..	..	..
500	−42	−43	..	..	..	..	..	..	..	..	−29	−33	..	..	0.1	31
550	−32	−34	−35	..	..	..	..	..	..	..	−19	−24	..	..	0.2	31
600	−23	−25	−26	..	..	..	..	..	..	..	−10	−17	..	..	0.4	30
650	−15	−17	−18	..	..	..	..	..	..	..	− 2	−10	..	..	0.5	29
700	− 7	−10	−11	..	..	..	..	..	..	..	6	− 5	..	..	0.7	28
750	− 1	− 3	− 3	..	..	..	..	..	..	..	12	1	..	..	0.8	27
800	5	5	5	..	..	..	..	..	..	..	17	7	..	..	1.5	27
850	11	10	11	..	..	..	..	..	..	..	23	13	..	..	2.0	26
900	18	16	17	..	..	..	..	..	..	..	29	19	..	..	2.7	27
950	23	21	24	..	..	..	..	..	..	..	34	23	..	..	3.5	27
A_2	Polar continental air with an anticyclonic track from north of 50°N. and east of 25°E.															
450	−45	−46	..	..	..	..	..	..	..	..	..	..	..	..	..	..
500	−35	−38	..	..	..	..	..	..	..	..	..	..	..	..	..	..
550	−25	−28	..	..	..	..	..	..	..	..	..	..	..	..	0.3	..
600	−15	−18	..	..	..	..	..	..	..	..	..	..	..	..	0.5	32
650	− 8	− 9	..	..	..	..	..	..	..	..	..	..	..	..	0.7	32

*Below the freezing point the vapour pressure is that over ice.
**Estimated value.

TABLE 15 (Continued)

Pressure	Monthly means of temperature												Summer (July and August)		Winter (January and February)	
	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Mean vapour pressure*	Mean wet-bulb potential temp.	Mean vapour pressure*	Mean wet-bulb potential temp.
mb.	°F.												mb.	°F.	mb.	°F.
A_2	Polar continental air with an anticyclonic track from north of 50°N. and east of 25°E. (Continued)															
700	− 2	− 3	..	..	..	..	..	..	..	..	..	..	..	..	0.8	31
750	3	3	..	..	..	..	..	..	..	..	11	6	..	..	1.0	30
800	8	8	12	..	..	..	..	..	..	..	17	12	..	..	1.6	29
850	15	14	19	..	..	..	..	..	..	..	23	18	..	..	2.1	27
900	19	19	23	..	..	..	..	..	..	..	26	23	..	..	3.1	28
950	23	24	28	..	..	..	..	..	..	..	34	27	..	..	4.1	29
P_{IC}	Polar maritime air in the Icelandic region															
450	−55	−56	−54	−50	−46	−38	−30	−30	−36	−45	−51	−54	0.1	46	..	..
500	−46	−47	−44	−39	−36	−28	−20	−19	−26	−34	−41	−43	0.3	45	..	..
550	−37	−38	−35	−31	−27	−19	−11	−10	−18	−25	−32	−33	0.6	45	0.1	27
600	−29	−29	−28	−23	−19	−12	− 2	− 2	−11	−18	−23	−25	0.9	44	0.2	26
650	−21	−22	−22	−16	−11	− 5	6	5	− 4	−11	−16	−18	1.5	43	0.3	24
700	−15	−15	−15	−10	− 5	2	11	9	2	− 4	− 9	−11	1.9	42	0.5	23
750	−10	− 9	− 8	− 4	1	8	17	15	8	3	− 2	− 5	2.4	41	0.7	21
800	− 4	− 2	− 1	2	8	14	23	20	13	9	4	1	3.4	40	1.0	20
850	2	2	5	9	14	20	28	26	19	15	10	7	4.4	40	1.2	19
900	8	9	11	15	20	26	35	31	25	21	16	13	5.5	41	1.5	18
950	14	15	17	20	26	33	41	37	32	27	22	19	7.3	42	2.1	17

*Below the freezing point the vapour pressure is that over ice.

TABLE 15 (*Continued*)

Pressure	Monthly means of temperature												Summer (July and August)		Winter (January and February)	
	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Mean vapour pressure*	Mean wet-bulb potential temp.	Mean vapour pressure*	Mean wet-bulb potential temp.
mb.	°F.												mb.	°F.	mb.	°F.
P_1	Polar maritime air with a cyclonic track from north and north-east of Iceland															
450	−50	−51	−49	−43	−39	−30	−24	−26	−30	−38	−44	−48	0.2	50	..	..
500	−40	−41	−38	−32	−28	−19	−13	−15	−19	−27	−33	−37	0.5	49	0.1	33
550	−30	−30	−28	−22	−18	− 9	− 4	− 5	−10	−17	−23	−27	0.9	48	0.2	32
600	−21	−21	−19	−14	− 9	− 1	5	3	− 1	− 8	−14	−18	1.4	47	0.3	31
650	−12	−12	−10	− 6	− 1	6	12	11	7	0	− 6	− 9	1.9	47	0.5	31
700	− 4	− 5	− 3	2	6	13	19	17	13	7	2	− 1	2.8	47	0.8	30
750	3	3	5	9	12	20	25	24	19	13	9	6	3.8	47	1.1	30
800	9	9	11	15	18	25	31	29	25	20	16	12	4.8	46	1.6	30
850	16	15	17	21	24	31	36	35	32	26	22	19	5.9	47	2.3	30
900	22	21	23	28	30	36	42	41	37	32	28	25	7.4	48	3.2	29
950	26	27	29	33	36	42	47	46	43	38	34	30	8.6	48	3.9	30
P_2	Polar maritime air with an anticyclonic track from north and north-east of Iceland															
450	−46	−47	−42	−39	−34	−26	−20	−22	−28	−32	−36	−40	0.1	49	..	..
500	−35	−37	−32	−28	−24	−16	− 9	−12	−17	−21	−26	−30	0.3	48	0.1	34
550	−25	−27	−23	−19	−15	− 7	0	− 3	− 8	−13	−16	−21	0.7	48	0.3	33
600	−17	−18	−15	−11	− 7	1	9	5	0	− 5	− 8	−12	1.2	47	0.5	33
650	− 9	− 9	− 7	− 3	1	9	16	12	8	3	− 1	− 5	1.8	47	0.7	33

*Below the freezing point the vapour pressure is that over ice.

TABLE 15 (Continued)

Pressure	Monthly means of temperature												Summer (July and August)		Winter (January and February)	
	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Mean vapour pressure*	Mean wet-bulb potential temp.	Mean vapour pressure*	Mean wet-bulb potential temp.
mb.	°F.												mb.	°F.	mb.	°F.
P_2	Polar maritime air with an anticyclonic track from north and north-east of Iceland (Continued)															
700	−1	−1	1	1	4	14	22	19	15	10	6	2	2.5	47	1.0	32
750	6	5	7	10	15	19	27	25	21	16	13	9	3.1	46	1.3	32
800	12	10	13	16	20	25	32	30	27	23	19	15	4.5	46	1.7	31
850	17	16	19	22	26	31	36	35	33	28	24	20	5.3	47	2.5	31
900	22	22	25	27	31	37	43	41	38	33	29	25	7.6	47	3.4	30
950	27	27	30	33	36	42	47	46	43	39	35	31	8.8	48	3.9	29
P_3	Polar maritime air with a cyclonic track from north-west of Iceland, approaching the British Isles from the north-west.															
450	−43	−45	−42	−39	−34	−27	−22	−22	−26	−31	−36	−40	0.3	51	0.1	38
500	−32	−33	−30	−27	−23	−16	−11	−12	−15	−20	−25	−28	0.6	50	0.2	38
550	−21	−23	−20	−17	−13	−7	−1	−2	−6	−11	−15	−18	1.0	50	0.4	37
600	−12	−13	−11	−9	−5	1	7	6	3	−3	−7	−10	1.5	49	0.6	37
650	−4	−5	−3	−1	3	9	15	14	11	5	1	−1	2.2	49	0.9	37
700	4	3	5	7	10	17	22	21	18	13	9	6	3.0	49	1.3	37
750	11	11	12	14	17	23	28	27	24	19	15	13	4.1	49	1.7	36
800	17	17	18	20	23	28	33	32	30	25	22	20	5.3	49	2.5	36
850	23	23	24	26	29	33	39	38	35	31	27	25	6.7	50	3.3	36
900	29	29	30	31	35	39	45	44	41	37	33	31	8.2	51	4.5	37
950	33	34	35	36	40	45	50	49	46	42	38	35	10.1	52	5.3	36

*Below the freezing point the vapour pressure is that over ice.

TABLE 15 (Continued)

Pressure	Monthly means of temperature												Summer (July and August)		Winter (January and February)	
	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Mean vapour pressure*	Mean wet-bulb potential temp.	Mean vapour pressure*	Mean wet-bulb potential temp.
mb.	°F.												mb.	°F.	mb.	°F.
P_4	Polar maritime air with an anticyclonic track from north-west of Iceland, approaching the British Isles from the north-west															
450	−36	−37	−35	−31	−25	−21	−15	−17	−21	−27	−31	−35	0.5	54	0.1	41
500	−26	−26	−24	−20	−15	−10	− 4	− 7	−11	−16	−20	−24	0.9	54	0.2	40
550	−16	−17	−15	−11	− 5	− 1	4	2	− 2	− 8	−11	−15	1.2	53	0.4	39
600	− 8	− 9	− 7	− 3	2	6	12	10	5	1	− 3	− 7	1.7	53	0.6	38
650	0	− 2	1	5	9	13	19	17	12	8	4	1	2.5	52	1.0	37
700	6	5	7	11	16	20	25	23	18	14	10	8	3.3	51	1.4	37
750	12	11	13	17	22	26	30	28	25	20	17	15	4.2	50	1.8	37
800	18	17	19	22	28	31	35	33	30	25	22	20	5.2	49	2.6	37
850	24	22	24	28	33	37	41	38	36	31	28	26	6.7	50	3.4	36
900	29	28	30	33	39	42	46	45	41	37	33	31	8.2	51	4.5	37
950	33	33	35	39	43	47	51	50	47	41	38	36	9.8	52	5.6	36
P_5	Polar maritime air with a cyclonic track from north-west of Iceland, approaching the British Isles from the west															
450	−37	−37	−35	−32	−28	−23	−18	−19	−22	−26	−30	−34	0.5	52	0.1	42
500	−26	−26	−24	−21	−17	−12	− 7	− 8	−11	−15	−19	−23	0.8	52	0.3	41
550	−16	−16	−15	−12	− 8	− 3	2	1	− 2	− 6	−10	−13	1.3	52	0.4	41
600	− 7	− 7	− 6	− 3	0	5	11	10	7	2	− 2	− 5	1.8	52	0.7	40
650	0	1	2	5	8	13	18	17	14	10	7	4	2.5	52	1.0	40

*Below the freezing point the vapour pressure is that over ice.

TABLE 15 (*Continued*)

Pressure	Monthly means of temperature												Summer (July and August)		Winter (January and February)	
	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Mean vapour pressure*	Mean wet-bulb potential temp.	Mean vapour pressure*	Mean wet-bulb potential temp.
mb.	°F.												mb.	°F.	mb.	°F.
P_5	Polar maritime air with a cyclonic track from north-west of Iceland, approaching the British Isles from the west (Continued)															
700	8	8	10	13	16	20	25	24	21	18	14	10	3.6	51	1.5	40
750	14	15	16	19	22	26	31	30	27	24	20	17	4.5	51	2.3	39
800	21	21	22	25	28	32	37	35	33	30	26	24	5.9	51	3.1	39
850	27	27	28	31	34	38	43	41	39	35	32	29	7.3	52	4.3	40
900	33	33	34	37	39	43	48	47	44	41	37	35	9.1	53	5.1	40
950	37	37	39	41	43	48	53	52	50	46	42	39	11.0	54	6.2	39
P_6	Polar maritime air with an anticyclonic track from north-west of Iceland, approaching the British Isles from the west															
450	−31	−31	−30	−27	−21	−17	−13	−13	−18	−23	−26	−29	0.5	55	0.1	44
500	−20	−21	−20	−16	−11	− 7	− 2	− 3	− 8	−13	−16	−18	0.9	55	0.3	43
550	−11	−12	−11	− 7	− 2	2	7	7	1	− 4	− 7	− 9	1.4	55	0.4	42
600	− 3	− 4	− 3	1	6	10	15	14	10	4	1	− 2	2.1	55	0.7	41
650	3	4	5	9	14	18	22	21	17	12	8	6	2.8	54	1.0	41
700	10	11	12	15	20	24	28	26	23	18	15	12	4.0	53	1.6	40
750	16	16	18	21	26	30	33	31	29	24	21	19	4.6	52	2.1	40
800	22	22	24	27	32	35	39	37	35	30	27	24	5.6	52	3.0	39
850	27	27	29	32	37	40	45	43	40	35	32	29	7.3	53	4.0	39
900	32	33	34	37	42	46	51	49	45	40	37	35	9.4	53	5.1	39
950	37	38	39	42	47	51	56	54	50	45	41	39	11.7	54	6.0	38

*Below the freezing point the vapour pressure is that over ice.

TABLE 15 (Continued)

Pressure	Monthly means of temperature												Summer (July and August)		Winter (January and February)	
	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Mean vapour pressure*	Mean wet-bulb potential temp.	Mean vapour pressure*	Mean wet-bulb potential temp.
mb.	°F.												mb.	°F.	mb.	°F.
P_7	Polar maritime air with a cyclonic track from north-west of Iceland, approaching the British Isles from south of 45°N.															
450	−28	−29	−29	−26	−21	−17	−14	−13	−16	−19	−23	−26	0.6	56	0.3	47
500	−17	−18	−17	−14	−10	− 6	− 2	− 1	− 4	− 8	−11	−15	1.0	56	0.5	47
550	− 8	− 9	− 8	− 5	− 1	3	7	7	5	1	− 2	− 5	1.6	56	0.8	47
600	1	0	1	4	8	12	15	16	14	10	7	4	2.2	56	1.1	47
650	9	8	9	12	15	19	22	23	21	17	15	12	2.9	56	1.7	46
700	16	15	16	19	22	26	29	29	28	24	22	19	4.3	55	2.4	46
750	23	22	22	25	28	32	35	36	34	31	27	25	5.2	55	3.0	45
800	29	28	27	31	34	38	40	42	40	36	33	31	6.6	55	4.2	45
850	34	33	33	37	39	43	46	47	45	42	39	37	8.7	56	5.4	45
900	39	39	39	42	45	48	51	52	50	47	44	42	11.0	57	7.1	46
950	44	43	44	47	50	54	57	56	55	52	49	46	12.9	57	8.4	46

*Below the freezing point the vapour pressure is that over ice.

TABLE 16 *Average and extreme daily maximum and minimum temperatures near the surface at Kew for each month for 18 air masses*

Air mass	*January*			*February*			*March*			*April*			*May*			*June*		
	Average	*Highest*	*Lowest*	*Average*	*Highest*	*Lowest*	*Average*	*Highest*	*Lowest*	*Average*	*Highest*	*Lowest*	*Average*	*Highest*	*Lowest*	*Average*	*Highest*	*Lowest*
	°F.																	
	Daily maximum temperatures																	
T_1	53	56	50	53	55	50	57	60	54	59	61	57	67	71	62	70	75	68
T_2	52	56	50	55	58	50	58	63	55	65	70	61	72	75	69	74	77	71
T_3	49	53	43	50	55	46	59	66	52	68	76	56	76	87	67	78	87	68
T_Q	48	51	42	49	51	44	52	55	48	55	58	53	61	66	58	67	75	63
H_O	41	47	33	46	55	35	55	68	42	62	73	50	68	80	53	75	86	68
H_{NE}	37	41	31	37	44	33	46	50	40	52	61	45	60	68	49	70	77	60
H_{SE}	41	49	35	45	55	38	53	59	48	64	73	59	70	79	63	79	86	74
H_{SW}	49	52	47	52	56	49	57	62	53	62	67	59	67	69	65	71	76	67
H_{NW}	41	46	38	44	49	41	50	53	47	61	65	58	65	68	61	72	80	65
A_1	31	34	28	34	40	30	39	46	32	43	44	41	..	..	..	..	..	..
A_2	32	36	27	34	39	31	39	45	31	47	48	45	..	..	..	..	..	..
P_1	36	39	32	38	41	32	41	46	37	47	51	41	52	58	46	59	62	53
P_2	35	39	29	39	41	36	44	48	39	50	53	47	54	57	50	61	66	58
P_3	41	44	39	44	48	42	47	49	45	53	58	51	57	60	54	63	67	59
P_4	41	44	38	43	47	39	49	53	44	55	59	49	60	66	54	66	74	61
P_5	46	49	44	47	48	43	51	54	49	57	61	54	62	63	59	65	68	60
P_6	48	51	44	49	54	45	52	57	47	61	70	56	67	74	61	71	78	63
P_7	50	53	48	51	54	48	54	57	52	59	65	54	65	68	63	69	71	64
M*	44	..	..	45	..	..	49	..	..	55	..	..	62	..	..	68	..	..
	Daily minimum temperatures																	
T_1	49	51	47	48	51	46	49	52	46	51	55	49	53	57	51	58	63	54
T_2	46	53	41	47	52	43	48	51	43	50	55	45	54	57	48	58	62	54
T_3	40	47	36	40	44	38	44	51	39	49	53	45	55	63	50	59	65	53
T_Q	42	47	38	40	46	34	44	48	39	46	49	41	48	53	44	55	56	52
H_O	29	35	21	30	39	23	34	41	24	38	49	31	44	52	33	51	59	44
H_{NE}	29	39	23	30	34	23	34	41	26	39	46	31	44	51	35	51	58	44
H_{SE}	33	38	27	32	39	28	36	41	31	44	49	38	48	55	44	55	60	49
H_{SW}	42	45	41	43	43	40	43	47	41	47	50	43	50	52	47	55	60	49
H_{NW}	33	39	29	34	38	30	37	41	34	42	48	38	45	50	40	52	58	45
A_1	25	30	19	28	32	24	32	37	25	37	39	33	..	..	..	..	..	..
A_2	25	30	18	27	30	21	29	35	19	32	35	31	..	..	..	..	..	..
P_1	31	35	27	31	35	27	34	39	31	37	41	32	41	46	35	48	53	44
P_2	25	30	18	28	29	24	30	34	26	36	39	31	38	41	30	46	50	42
P_3	34	36	32	36	39	32	37	38	34	42	46	38	46	49	41	51	53	47
P_4	33	41	27	34	39	28	36	41	31	41	43	35	44	50	39	51	57	46
P_5	38	42	35	38	41	35	41	43	37	46	50	43	49	53	46	53	56	49
P_6	37	40	34	37	41	34	41	45	36	44	48	41	48	53	44	52	59	48
P_7	44	46	40	43	45	41	44	46	41	47	51	44	51	53	49	55	58	54
M*	35	..	..	35	..	..	36	..	..	40	..	..	45	..	..	51	..	..

* M = all air masses combined, 1871-1940

TABLE 16 *(Continued)*

Air mass	*July*			*August*			*September*			*October*			*November*			*December*		
	Average	*Highest*	*Lowest*	*Average*	*Highest*	*Lowest*	*Average*	*Highest*	*Lowest*	*Average*	*Highest*	*Lowest*	*Average*	*Highest*	*Lowest*	*Average*	*Highest*	*Lowest*
	°F.																	
	Daily maximum temperatures																	
T_1	72	76	67	72	77	67	70	72	67	62	67	59	57	61	54	54	59	52
T_2	76	78	74	76	78	72	71	75	65	62	65	59	57	61	55	53	55	51
T_3	82	91	72	82	92	74	74	81	67	64	71	58	57	66	53	52	55	49
T_Q	68	71	63	68	71	63	65	71	62	58	62	52	53	58	49	50	53	45
H_O	78	85	72	73	82	65	68	79	60	56	66	47	43	52	39	38	45	30
H_{NE}	72	79	65	72	77	64	65	72	57	54	61	47	49	56	43	40	43	33
H_{SE}	80	85	77	80	84	75	71	79	65	58	63	52	49	55	43	42	49	36
H_{SW}	76	80	74	75	79	71	70	73	64	59	63	54	51	55	48	49	51	47
H_{NW}	75	79	71	74	77	69	66	74	57	57	61	51	47	52	43	42	45	39
A_1	..	..	..	..	..	..	..	..	..	..	..	..	41	41	40	34	39	27
A_2	..	..	..	..	..	..	..	..	..	..	..	..	41	42	39	36	40	29
P_1	64	66	58	62	64	57	57	61	53	47	53	43	43	48	40	40	44	36
P_2	67	70	65	65	68	61	59	64	55	47	49	45	43	46	37	46	42	31
P_3	66	68	63	66	68	63	61	67	59	51	54	47	48	51	46	44	47	42
P_4	69	74	66	69	72	64	62	68	55	54	60	48	47	51	41	43	48	36
P_5	69	72	67	69	72	66	65	68	61	56	58	53	51	53	48	49	51	47
P_6	74	83	69	76	82	68	69	76	63	59	65	55	53	56	49	49	54	44
P_7	72	75	69	70	74	64	69	73	63	59	63	56	54	56	52	52	54	49
M*	71	..	..	70	..	..	65	..	..	56	..	..	49	..	..	45	..	..
	Daily minimum temperatures																	
T_1	60	63	57	60	64	57	59	61	56	55	61	51	53	58	49	50	53	47
T_2	60	63	57	61	64	57	58	62	53	52	55	49	49	54	46	46	49	41
T_3	61	66	54	61	66	55	58	64	48	51	57	45	48	53	45	45	48	40
T_Q	57	61	56	57	60	56	55	59	50	50	55	45	45	49	41	43	47	39
H_O	54	61	49	51	57	43	46	54	40	37	48	28	32	39	25	29	39	33
H_{NE}	56	60	52	57	61	51	53	58	43	44	52	33	39	48	30	33	41	27
H_{SE}	57	61	53	56	61	53	52	59	45	43	51	36	40	44	34	34	42	26
H_{SW}	59	63	56	59	62	53	57	61	53	52	56	49	47	49	44	45	47	43
H_{NW}	57	61	52	54	57	50	52	56	46	41	47	36	36	43	30	34	37	30
A_1	..	..	..	..	..	..	..	..	..	..	..	..	37	39	35	27	31	21
A_2	..	..	..	..	..	..	..	..	..	..	..	..	37	38	34	29	34	23
P_1	53	57	48	52	55	45	47	54	43	36	39	33	34	41	30	33	38	29
P_2	49	52	46	49	53	44	44	49	38	35	39	29	30	33	25	26	32	19
P_3	54	58	48	54	58	48	52	57	46	42	48	37	39	43	35	36	39	33
P_4	53	58	48	51	57	47	48	54	42	41	47	36	37	43	29	33	39	27
P_5	57	59	53	56	60	52	54	58	49	46	50	42	43	47	39	41	44	37
P_6	56	62	53	56	62	50	52	58	46	46	48	40	42	47	37	40	45	35
P_7	59	62	57	58	61	57	56	59	54	49	53	43	46	50	44	45	49	41
M*	55	..	..	54	..	..	50	..	..	43	..	..	39	..	..	36	..	..

* M = all air masses combined, 1871-1940

Tephigrams of average conditions in summer and winter of eight of the air masses included in Table 13 are reproduced in Figures 15, 16 and 17.

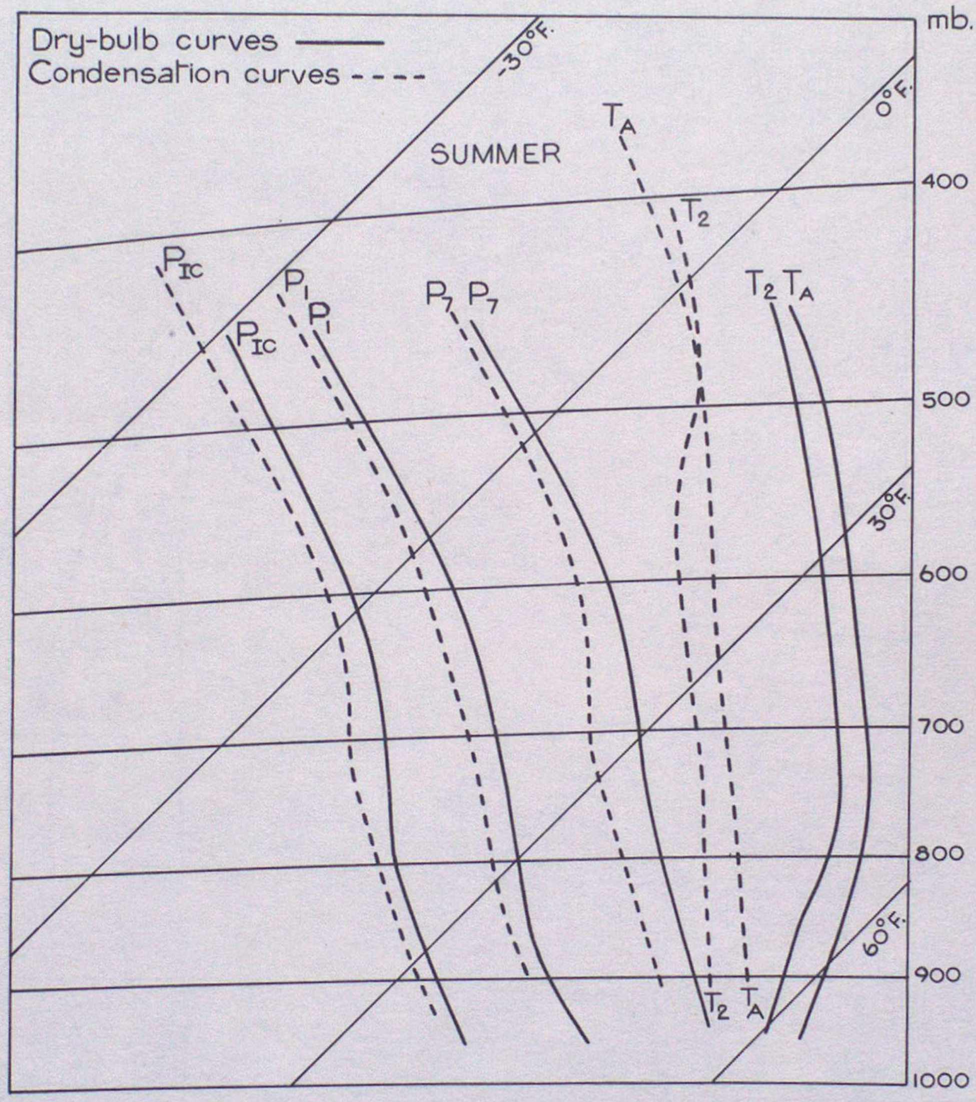

FIGURE 15 *Tephigrams of average conditions in air masses T_A, T_2, P_{IC}, P_1 and P_7 in summer*

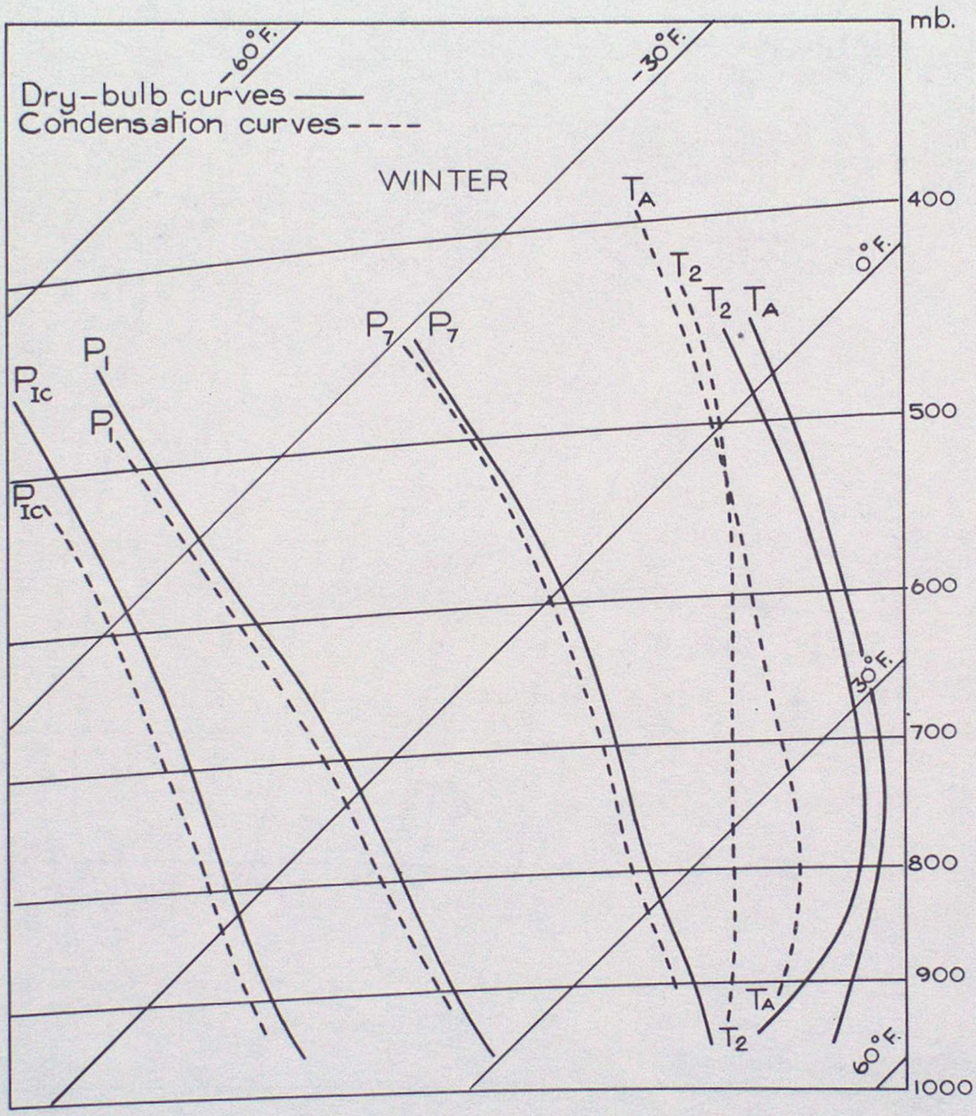

FIGURE 16 *Tephigram of average conditions in air masses T_A, T_2, P_{IC}, P_1 and P_7 in winter*

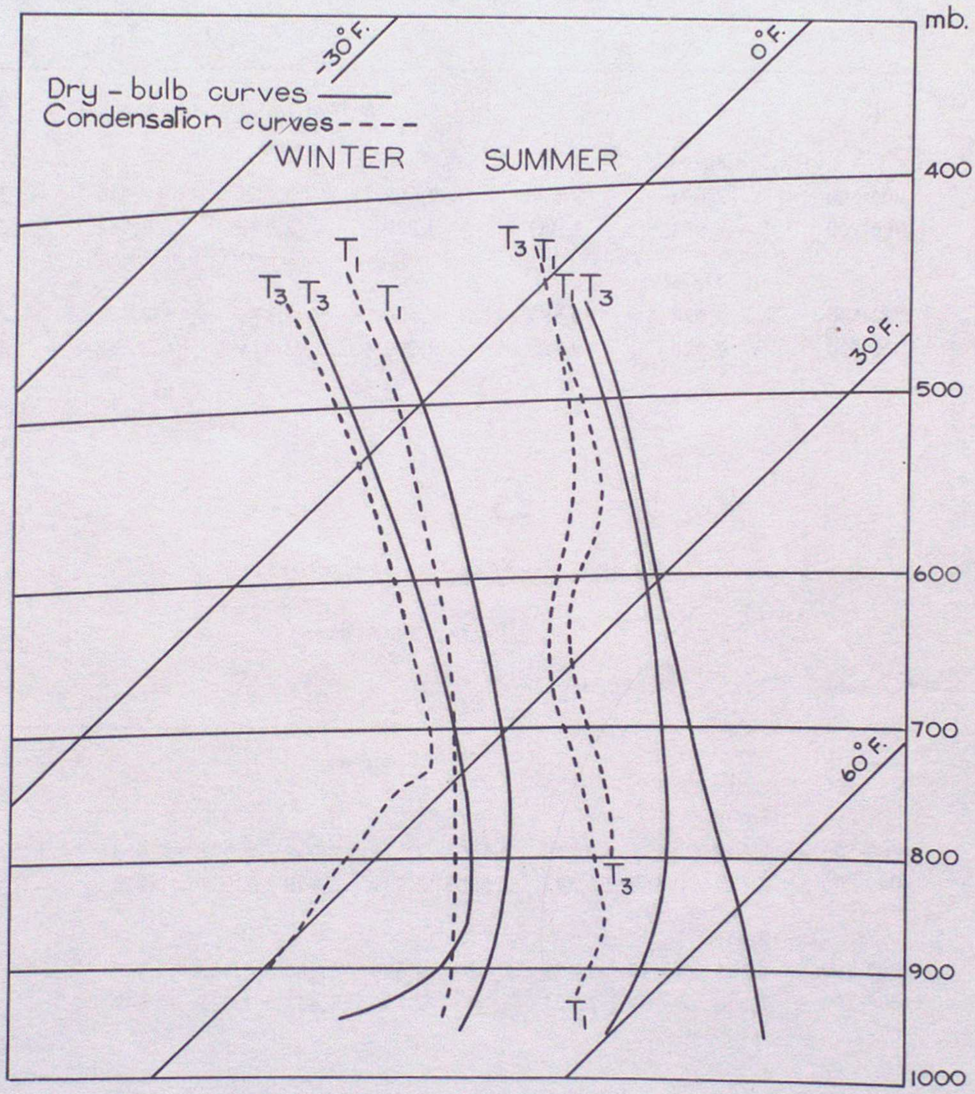

FIGURE 17 ***Tephigrams of average conditions in tropical maritime (T_1) and tropical continental (T_3) air in summer and winter***

Mean partial thicknesses for 1000-700 and 700-500 millibars for most of the air masses included in Table 13 are given in Tables 17 and 18.

TABLE 17 *Mean partial thicknesses*

Thickness	T_A	T_1	T_2	T_3	T_Q	H_O
			Air mass			
mb.			*metres*			
	Summer					
700-500	2,694	2,676	2,688	2,676	2,630	2,679
1000-700	2,981	2,963	2,972	2,996	2,944	2,941
	Winter					
700-500	2,624	2,582	2,604	2,576	2,569	2,588
1000-700	2,926	2,880	2,899	2,874	2,868	2,865

TABLE 18 *Mean partial thicknesses*

Thickness	P_{IC}	P_1	P_2	P_3	P_4	P_5	P_6	P_7	A_1	A_2
					Air mass					
mb.					*metres*					
	Summer									
700-500	2,508	2,539	2,539	2,560	2,573	2,569	2,597	2,603	...	...
1000-700	2,816	2,868	2,871	2,883	2,902	2,908	2,917	2,923	...	...
	Winter									
700-500	2,362	2,410	2,475	2,450	2,478	2,484	2,512	2,527	2,390	2,438
1000-700	2,658	2,749	2,771	2,790	2,795	2,816	2,825	2,847	2,731	2,752

Table 19 gives for the period 1871-1948, the absolute highest and lowest maximum and minimum screen temperatures recorded at Kew together with the dates and the air masses in which these extremes occurred. This table is included for the guidance it can give to forecasters in forecasting surface temperatures in synoptic situations when extremes of temperatures seem likely.

Although Belasco's data are restricted generally to levels at or below 500 millibars and mean values should not be used in place of direct upper air observations in the appropriate air mass, they can nevertheless materially assist in interpreting in terms of temperature a forecast chart for 24 hours or more ahead.

Temperature

TABLE 19 *Extremes of temperature at Kew, 1871-1948*

Month	Highest maximum				Lowest maximum			
	Temp.	Air mass	Date Day	Date Year	Temp.	Air mass	Date Day	Date Year
	°F.				°F.			
Jan.	57	T_2	2	1922	22	A_1*	5	1894
Feb.	62	T_2*	10	1899	23	A_2*	9	1895
March	69	H_O	23	1945	31	A_1	9	1933
Apr.	80	T_3*	20	1893	35	P_1	5	1911
May	87	T_3	{24, 29}	{1922, 1944}	46	P_2*	1	1877
June	91	T_3	3	1947	50	P_1	19	1903
July	91	T_3	31	1943	54	P_1	5	1920
Aug.	94	T_4	9	1911	55	P_2*	7	1898
Sept.	92	T_4	1	1906	48	P_1	29	1918
Oct.	82	T_3	5	1921	38	P_2*	29	1873
Nov.	66	T_2	5	1938	25	P_1*	28	1890
Dec.	59	T_1	4	1931	22	A_2*	14	1890

Month	Highest minimum				Lowest minimum			
	Temp.	Air mass	Date Day	Date Year	Temp.	Air mass	Date Day	Date Year
	°F.				°F.			
Jan.	53	T_2	3	1922	9	P_2*	17	1881
Feb.	52	H_{SW}	10	1939	11	A_1*	7	1895
March	52	T_2	25	1912	17	P_1	5	1909
Apr.	55	T_3	15	1945	26	P_2	2	1922
May	63	H_{SE}	30	1944	30	H_{NE}*	5	1877
June	65	T_3	25	1935	37	{P_2, P_2}	5 3	1880 1923
July	68	T_3	29	1948	43	{P_2, H_{NW}}	1 11	1882 1907
Aug.	67	H_{SE}	13	1911	41	H_{NW}*	31	1890
Sept.	64	T_2	2	1932	31	H_{SE}	30	1919
Oct.	61	T_1	{6, 10}	{1916, 1933}	25	P_1*	28	1895
Nov.	58	T_2	5	1938	20	H_O*	19	1871
Dec.	53	T_1	4	1934	11	H_O*	22	1890

* Estimated.

14.8.2. Some other features of the temperature distribution in the free atmosphere

In addition to the routine upper air observations flights by specially equipped aircraft have been made for a number of years to explore particular features of the atmosphere. Using data from such flights and other available routine upper air observations a number of workers have published papers on several features of the atmosphere and some general account of these was given in Chapter 7. Forecasters should refer to Chapter 7 as necessary but the following points should be borne in mind when analysing and forecasting the distribution of temperature.

14.8.2.1. Temperatures in the vicinity of fronts. Sawyer[50] drew a distinction between the broader frontal regions and the narrower frontal zones. He defined a frontal region as the distance over which a significant temperature gradient extended and a frontal zone as that distance over which the steepest gradient of temperature extended. The average width of the frontal region was about 600 miles and the average temperature difference across it about 15°F. (8°C.). Frontal zones had an average width of 130 miles (that is, about one quarter of the total width of the frontal region) and a temperature difference of 9°F. (5°C.) (a little more than half the total air-mass contrast). Variations from front to front and the uncertainties of the analysis were such that these figures should be regarded as giving little more than the broad indication of the usual structure of a frontal zone. Nevertheless Sawyer felt justified in concluding that about half the total air-mass contrast of temperature was concentrated in a frontal zone with a width usually of 100-200 miles. These figures should provide a useful background when analysing and forecasting the distribution of temperature in day-to-day operational work.

14.8.2.2. Temperatures at the tropopause. Examination of published literature indicates that it is not yet possible to present a comprehensive set of models related to surface or upper features recognizable on synoptic charts which forecasters could use as routine in preparing forecasts of the height and temperatures at the tropopause. Reference to the treatment of the tropopause in Chapter 7 shows that there are discontinuities and folds in the surface of the tropopause. However, from the synoptic and analytical viewpoint, at most times the tropopause may be regarded as a material surface moving with the air. At infrequent intervals the tropopause dissipates at one level and reforms at another.

One type of tropopause discontinuity which is often clear cut and is widely recognized is the tropopause funnel. In the central part of the funnel the tropopause is low - usually well below 300 millibars and sometimes below 400 millibars. Although the potential temperature at this low tropopause is usually between 27 and 35°C. (80 and 95°F.) this leads to a very wide range of actual temperature (−36 to −60°C.) if this tropopause is assumed to form between 400 and 300 millibars. Some narrowing of the range is normally possible from a consideration of details of the temperature distribution on any one day. Reference to Chapter 7 will indicate the types of synoptic situation which are associated with the formation or existence of tropopause funnels.

Tropopause funnels with a closed wind circulation appear to be dynamically stable and are often associated with a cold pool in the troposphere. Once formed these tend to persist for a few days. Their collapse is often associated with the overrunning of a higher tropopause associated with the approach of a vigorous warm frontal system and when this does occur their decay is rapid and they may disappear from the chart in 12 to 24 hours.

There seems to be a close association between the folded tropopause and the jet stream and also with major warm-front systems. A more detailed account is given in Chapter 7. The available evidence does not seem sufficiently firm to indicate that a folded tropopause should be forecast whenever these conditions seem likely to exist but it does imply that forecasters should always be on the alert for such a development and examine critically all available data in the appropriate region for the first signs of such a development.

14.8.2.3. Temperatures and the jet stream. Some thermal features of the jet stream were described in Chapter 7 and these should assist forecasters when attempting to fill in detail between the widely spaced upper air observations or when ascents fail to reach the level of the jet stream.

14.8.2.4. Temperatures in the stratosphere (below 20 kilometres). Routine radio-sonde ascents from stations in north-west Europe reach heights of the order of 15 to 20 kilometres (about 100 to 50 millibars) with fair regularity. On any one occasion sufficient ascents normally reach these heights to enable charts and tephigrams to be plotted and the horizontal (or isobaric) and vertical distribution of temperature to be determined with fair accuracy. Examination of these data from the tropopause to about 50 millibars shows that day-to-day variations of temperature exist and that it is possible to distinguish areas of relatively warm and cold stratospheric air. These areas of warm and cold air behave quite differently from those in the troposphere where synoptic experience soon shows that these tropospheric areas of cold and warm air tend to move with the wind. In the stratosphere the areas of warm and cold air seem quite slow-moving. Sawyer[51] finds that these areas tend to retain a constant position in relation to the troughs and ridges of the flow as the wind blows through them and considers that the vertical motion of the air at these levels is probably dominant in controlling the temperature.

When a jet stream exists in the upper troposphere there are often strong temperature gradients in the lower stratosphere. The information contained in the treatment of the jet stream in Chapter 7 should assist both in analysis and forecasting.

14.8.2.5. Temperatures in the stratosphere (from 20 to 30 kilometres). Day-to-day observations of temperature at these levels are not available as routine. If forecasts of temperatures at these heights are required they must be based on the statistical information given in Chapter 7 and modified according to any recent observations which may be available.

14.9. FORECASTING TEMPERATURES IN THE UPPER AIR

Forecasting of temperatures in the upper air for short periods (less than or equal to about 24-36 hours) as practised in the United Kingdom rests on analyses of the horizontal (or isobaric) and vertical distributions of actual temperatures (compiled from radio-sonde and aircraft reports) by means of upper air charts and tephigrams. From the analyses of surface and upper air contour charts forecast charts can be prepared as described in earlier chapters and these will include forecast values of thicknesses which will indicate the vertically averaged temperature for the several layers of the atmosphere. Except in the lowest few thousand feet and near well marked inversions it is usually possible to infer the forecast temperature from the forecast thickness for an appropriate layer, when this is available, by sketching an appropriate lapse rate on the tephigram through the forecast thickness value. The following paragraphs discuss some of the physical processes, consideration of which will supplement any estimates which can be made from forecast thickness values or provide assistance in the absence of such forecasts.

Consideration should first be given to advection so that an assessment can be made of the movement of the air at the level(s) required for the period in question. This follows without appreciable difficulty from the co-ordinated analyses of a time series of surface and upper air charts. From this analysis and prognosis the forecaster will be able to determine with fair accuracy the location of the air for which he must then determine from the available upper air ascents the actual vertical distribution of temperature as accurately and in as much detail as is required and possible. Having determined this the next step is to modify the vertical distribution to allow for the physical processes which are expected to affect that air and so modify the temperatures. An important process is transfer of heat between air and the surface of the earth of different temperatures and the subsequent redistribution of heat within the air by convection or turbulence. Other important processes are vertical motion, cooling by evaporation of precipitation and radiation.

14.9.1. Modification by contact with the earth's surface

14.9.1.1. Warmer air across a cooler surface. Contact with the cooler, underlying surface soon cools the lower layers of the atmosphere and establishes a stable lapse rate or perhaps an inversion. Turbulent transfer of heat then transports heat from the overlying layers downwards and this is one of the processes by which the layers above the surface are cooled. This cooling of upper layers is not very rapid and does not normally extend above about 850 millibars. Cooling of air over a cool surface therefore primarily affects the layer near the surface and some rules for forecasting these low-level temperatures are given in Section 14.10. For somewhat higher levels quantitative rules do not seem to be available but Belasco[49] has established some figures relating to cooling of air as it moves from the region of the Azores to the United Kingdom 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 20. 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 United Kingdom.

TABLE 20 *Average decrease of temperature in tropical maritime air between the Azores and the coasts of south-west England*

Season	*Average decrease*			
	Sea temperature	*Air temperature*		
		Surface	*950 mb.*	*900 mb.*
	°F.	°F.	°F.	°F.
Summer	11	9	3	2
Winter	10	8	5	3

14.9.1.2. Colder air across a warmer surface. Air moving from a colder to a warmer surface is characterized by an unstable lapse rate throughout a layer of considerable depth from the surface upwards. As air is advected across a warmer sea and the surface air is warmed, convection takes place and this distributes the heat gained by the surface layers throughout the layer which is penetrated by convection. This redistribution of heat usually continues in a vigorous fashion. Some values for surface temperatures can be computed by the rules in Section 14.10 and an assessment of the vertical temperature distribution over the United Kingdom may be made by examination of actual tephigrams and a comparison with the averages contained in Table 15. It should be remembered that the values in Table 15 are not those due solely to warming by the sea and convective processes, but are the result of all processes which are taking place. For cold airstreams moving south warming by subsidence, particularly on the western parts of the northerly flow, is very common and the effect of vertical motion will now be described.

14.9.2. Effect of vertical motion

When cloud-free air subsides it warms at the dry adiabatic rate. Forecasters with only limited experience soon recognize the large modification to vertical temperature distribution and the vertical limit of convection which subsidence can bring about in a few hours. The crux of the problem for forecasting its effect on temperatures is to determine when and where subsidence will start, and the levels through which it will be effective and the period for which it will continue. For those regions for which upper air ascents are available, tephigrams should be carefully analysed for the first signs of modification due to subsidence. Use should be made of the conservation of wet-bulb potential temperature to identify the parcels of air and so determine the magnitude of any descent which has occurred. The extent to which any such subsidence will be continued or extended in area must be estimated from a critical examination of synoptic surface and upper air patterns. From well marked thickness patterns it is possible to indicate areas where general descending motion may be taking place. When this is coupled with a detailed examination of the tendency field, of the vertical extent of convective clouds and of any decrease in shower activity some estimates can be made of the amount and extent of subsidence. Where stabilization of a cold air mass is not due to the overrunning of warmer air much valuable information can often be extracted from aircraft reports about tops of convective clouds or from a verbal debriefing of aircrew. In the clear air from a height well above cloud tops aircrew can observe the horizontal variation in vertical development of convective clouds and a pertinent question during debriefing will often either confirm or deny the existence of suspected subsidence. When aircraft are flying several thousand feet above cloud tops, estimates by aircrew of the actual heights of the tops may be somewhat in error but the relative variation of heights and the flattened appearance of the cloud tops are usually a good indication of the existence of a stable layer which, under synoptic conditions where there had previously been a deeper layer of unstable air, may be regarded as the existence of a subsiding layer. When successive radiosonde ascents clearly indicate that subsidence has just commenced and the synoptic situation seems favourable for continued subsidence, some reliance may be placed on continued subsidence at about the same rate for a further 12 to 24 hours. Much judgment must be used in applying this simple extrapolation. When cold air bursts southwards and there is a broad stream of cold air, subsidence is often noticeable in the more southerly parts towards the western edge of the polar outburst - particularly where the isobars become curved anticyclonically.

Estimates of subsidence must be consistent with a chart of the forecast thickness pattern when this is available. In the preparation of such charts the existence of subsidence often appears necessary if the thickness pattern is to be consistent with the orderly development and extrapolation of the 1,000-millibar and 500-millibar contour patterns.

Subsidence or ascent in the atmosphere, extending over areas of 100 miles radius or more, is believed to extend through most of the troposphere reaching its maximum in the middle troposphere, that is, around or somewhat below 500 millibars. Ascent tends to be concentrated in frontal regions or other areas of precipitation and 10 centimetres per second (17 millibars per hour) would be a typical velocity here. Subsidence tends to occur over larger areas with somewhat lower speeds, for example, 3 centimetres per second (5 millibars per hour)[52] and ascent in unsaturated air masses has similar characteristics. The "so-called" subsidence inversion may not represent the lower limit of descent as air may pass downwards through an inversion which may be maintained by radiative and turbulent processes which are likely to be vigorous at a cloud top.

When air rises bodily in a mass at a frontal surface or a range of hills, the air cools at the dry adiabatic lapse rate (5.4°F. per 1,000 feet) until the dew-point is reached and thereafter at the saturated adiabatic lapse rate.

Because active ascent usually leads to condensation and the cooling is partially offset by the release of latent heat, and also because it is usually associated with the advection of warmer air, the effect of ascent on the plotted upper air soundings is not so immediately obvious as that of subsidence.

14.9.3. Cooling by precipitation

As precipitation falls through air which is not saturated (with respect to water and ice for liquid and frozen precipitation respectively) evaporation takes place. In order to transform part of the precipitation to vapour, latent heat of evaporation must be supplied and it is extracted mainly from the air through which the precipitation is falling. No simple formula which could be applied for practical day-to-day forecasting is known for the rate of cooling or the extent of the cooling. There is a clearly defined lower limit to which air can be cooled by the evaporation of water into it and this is the wet-bulb temperature. Some assessment of the rate of cooling is desirable for practical forecasting. Some rates of cooling for three uniform rates of rainfall of uniform drop size were calculated in *Note on lowering of cloud base during rain.*[53] The following values from that paper indicate the approximate times required for the initial difference between the dry- and wet-bulb temperatures to be reduced to one tenth of its value.

Very heavy rain (50 millimetres per hour, uniform drop size 2 millimetres diameter) time about ¾ hour.

Heavy rain (5 millimetres per hour, uniform drop size 1 millimetre diameter) time about 3¼ hours.

Light rain (1 millimetre per hour, uniform drop size 0.2 millimetre diameter) time about 1¼ hours.

Dolezel[54] has investigated this problem and has made rather more elaborate calculations but has restricted all his calculations to rainfall at the rate of 0.1418 inches per hour (approximately 3.6 millimetres per hour). From a preliminary investigation of the amounts of evaporation from raindrops of various sizes for two very different distributions of sizes of raindrops he further confined his calculations to uniform drops of 1.2 millimetres in diameter and to the evaporation taking place as the drops fell through a column of air 1,200 metres thick. He investigated three types of vertical temperature distribution: isothermal, a lapse of 0.5°C. per 100 metres and an inversion of 1.0°C. per 100 metres. He chose his values of temperature so that the initial mean temperature of the 1,200-metre layer with base at 1,000 millibars was 10°C. and the relative humidity was 50 per cent, corresponding to a wet-bulb temperature of 5.4°C. For all three types of vertical temperature distribution cooling is rapid in the first hour and is approximately 3°C., that is, the difference between dry- and wet-bulb temperatures is reduced to ⅓ in the first hour. From inspection of Dolezel's graphs the initial differences between dry- and wet-bulb temperatures would be reduced to one tenth, (that is, by nine tenths) in about 1 hour 25 minutes with an inversion of 1°C. per 100 metres, in about 1 hour 50 minutes with isothermal conditions and in about 2 hours 10 minutes with a lapse of 0.5°C. per 100 metres. These times are materially shorter than those for broadly comparable rainfall (5 millimetres per hour, uniform drop 1 millimetre diameter) in *Note on lowering of cloud base during rain.*[53]

A reasonable working rule would seem to be that the initial difference between dry- and wet-bulb temperatures will be reduced to one tenth of its value (that is, the dry-bulb temperature will approximate to the wet-bulb temperature) in about 1¼ to 2 hours after the commencement of all but the heaviest rain and that the greater the stability of the air the more rapid will be the cooling. For very heavy rain (as in a violent convective shower) cooling down to the wet bulb may occur in about half an hour. The preceding discussion deals only with the ideal case of rain falling continuously through the same air column - the times being measured from the start of the rain. When a mature shower drifts across a station the temperature drops rapidly and the period for which rain has fallen may not be known. However the preceding considerations give some guidance on the nature of the temperature changes to be expected. On those occasions when precipitation is not continuous or evaporates before reaching the ground the extent of cooling must be estimated.

Another factor which should be considered is the cooling of air as frozen precipitation falls through it. This process can cool air below its wet-bulb temperature. It is a very important factor in modifying the thermal structure of the lower air and should always be considered when deciding, in marginal conditions, whether precipitation will reach the ground in the frozen state as hail or snow or as liquid. For a hail shower which lasts 10 minutes, produces the equivalent of 10 millimetres of rainfall and contains hailstones of a fairly dense structure (density 0.9 grammes per cubic centimetre) and of diameter 4 millimetres, Gold[55] calculated that the hailstones would fall about 2,000 metres below the freezing level through air with a dry-adiabatic lapse rate before all the hailstones would be melted. For less dense hail (graupel: density 0.5 grammes per cubic centimetre) melting would be complete in about half the distance. Gold concluded that in such a shower cooling at about 1,000 metres below the freezing level would be some 1° to 2°C. Some later calculations by Mason[56] on the melting of hailstones confirm the order of Gold's figures for the distances through which hail will fall before being completely melted.

Snow falls much more slowly than hail and the amount of cooling at a given distance below the freezing level is greater than for hail. For a snow-fall equivalent to 10 millimetres of rain Gold calculated that air with a dry-adiabatic lapse rate could be cooled by 5°C. at about 1,000 metres below the freezing level. During this cooling process a dry-adiabatic lapse rate would be reduced to about half its value. If snow is also lying on the ground to a depth of 1 centimetre and is melted entirely by the overlying air and not by the ground, Gold found that it could cool a layer of air 30 metres thick by some 10°C. From these figures it will be recognized that snow-fall can substantially modify temperatures in the vertical below the 0°C. isotherm. On those occasions when continuous snow-fall seems likely (height at which the wet-bulb temperature reaches 0°C. is usually less than about 1,000 metres) and snow may lie, cooling by the falling and lying snow tends to produce an almost isothermal layer from the ground to the freezing level with temperatures close to 0°C.

14.9.4. Effect of radiation

Although the effect of radiation on the physical state of the atmosphere seems now fairly well understood and calculation of the various radiative fluxes can be made with reasonable accuracy, there has been little positive progress towards the achievement of techniques which can be successfully applied to day-to-day forecasting. No practical technique is offered but an indication of the magnitude of the cooling for the extreme cases of no cloud and a complete cover of cloud is given. When the atmosphere is completely cloudless it is estimated that a cooling of some 2° to 3°F. at all levels of the atmosphere in the latitude of the United Kingdom is likely during a period of 24 hours. When there is a complete cover of low cloud, the total loss of heat is not greatly altered but is concentrated at the cloud tops. Between the ground and the cloud layer, radiation modifies the temperatures only slightly. The cloud top radiates as a black body and the loss of radiation is considerable and causes a marked drop in temperature at the cloud top. Cooling due to radiation alone could reduce temperatures at the cloud top by 10°F. in 12 hours - but this cooling will then require an unstable lapse of temperature in the upper regions of the cloud. Turbulent mixing within the cloud then modifies the distribution of heat and temperature through the unstable layer. Where broken cloud or cloud at medium and high levels is present calculation of radiative balances is more difficult. Variation in humidity also substantially alters the flux of radiation. An analysis of some measurements of the flux of long-wave radiation by Houghton and Brewer[57] has confirmed that a rate of cooling of about 2° to 3°F. per 24 hours at all levels for a cloud free atmosphere is about the best estimate for forecasting that can be expected with our present techniques.

14.10. EFFECT OF SURROUNDING SEAS ON AIR TEMPERATURES OVER THE UNITED KINGDOM

The general theory and moderating effect of surrounding seas on the temperatures (and general climate) of a land area are well known and need no repetition in this handbook. Air arriving in the British Isles from directions between south-south-west and west has usually had a long sea track over the ocean where sea temperature gradients are not very strong. As a consequence the rate of modification of temperature in the latter part of the sea track is relatively slow. Reports and upper air ascents, although sparse, are usually adequate to estimate the temperature in the air mass on arrival at the coast. Values in Table 15 can usually be assumed typical of unstable air masses from north or north-west according to their track. Flow from a westerly direction is more common than that from the east and this contributes to the forecaster's greater experience in westerly types. Air which has arrived from a north-east through east to south-south-east direction has generally had some form of continental land track in its fairly recent history and modifications to its structure by a relatively short sea track are most important for accurate forecasting in such situations. Owing to the comparative infrequency of easterly types forecasters do not rapidly gain experience and some rules for calculating the effect of the North Sea on temperatures, which were formulated several years ago and have been probably little used, are included. In winter the North Sea is relatively warm to cold easterly streams and consequently warms them before they reach our shores and the extent of warming will now be considered.

14.10.1. The influence of the North Sea in winter on temperature and dew-point

Frost[58] has propounded 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

(i) the variation of wind direction with height may be neglected
(ii) the air in contact with the surface of the sea is saturated at the temperature of the sea surface and
(iii) 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 x_1 and T_1 are the humidity mixing ratio and temperature of the original air, and x_2 and T_2 are the values corresponding to the sea surface temperature then the humidity mixing ratios and temperature x_3 and T_3 of the air, after a trajectory over the North Sea in excess of about 60 miles, are given by the following formulae:

$$T_3 = T_1 + 0{\cdot}6\ (T_2 - T_1) \quad \ldots\ldots \quad (1)$$

$$x_3 = x_1 + 0{\cdot}6\ (x_2 - x_1)\ . \quad \ldots\ldots \quad (2)$$

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 x_1 and x_2 from the tephigram and calculate T_3 and x_3. The dew-point corresponding to x_3 is obtained from the tephigram. Frost quotes a number of examples - two of which are given here.

(i) 26 March 1939.

The temperatures of the air at Skagen and Blaavands Huk in Denmark were 32°F. and 31°F. respectively. The dew-point at both stations was 26°F. and the temperature of the North Sea was 43°F. From a tephigram a dew-point of 26°F. corresponds with a humidity mixing ratio of 2.9 grammes of water vapour per kilogram of air and a sea temperature of 43°F. with a humidity mixing ratio of 5.9. The temperatures of the air on arrival in England should be:

$$\begin{aligned} T_3 &= 32 + 0{\cdot}6\ (43 - 32) \\ &= 38{\cdot}7^\circ\mathrm{F.}, \\ \text{or}\quad T_3 &= 31 + 0{\cdot}6\ (43 - 31) \\ &= 38{\cdot}2^\circ\mathrm{F.}, \end{aligned}$$

or say between 38° and 39°F. The humidity ratio should be:

$$\begin{aligned} x_3 &= 2{\cdot}9 + 0{\cdot}6\ (5{\cdot}9 - 2{\cdot}9) \\ &= 4{\cdot}7 \ \text{(corresponding to a dew-point of 37°F.).} \end{aligned}$$

The temperature of the air on arrival at the English coast was 38°F.

(ii) 18-20 December 1938.

The conditions are of special interest. On 18 and 19 December a cold and almost dry easterly current extended from Russia across the continent to England. The temperature of the North Sea was 42°F. to which corresponds a humidity mixing ratio of 5.7. If it is assumed that the vapour content of the air over the land was nearly zero then by application of equation (2) the humidity mixing ratio which will occur after a passage across the North Sea will be (0.6×5.7) that is, 3.4 corresponding to a dew-point of 29°F. (Note that this dew-point is independent of the initial temperature of the air which is assumed dry on leaving the continent.) The dew-point actually observed at Spurn was 30°F. On the 19th the weather at Spurn Head was cloudy with occasional slight snow showers which became heavy and almost continuous by the 20th. Over the continent from the 17th to the 20th the temperatures were becoming steadily colder. On the 17th and 18th the temperatures were between 12°F. and 20°F. over north-west Germany and Holland and by the 19th temperatures had fallen to under 7°F.

While temperatures on the continent had been between 12°F. and 20°F. only slight snow showers had occurred and, whilst forecasters would probably anticipate increasing snow showers with increasing cold air on the continent, it is only possible to indicate this in a vague manner. Without bothering about the thermodynamical complications that might be anticipated when the temperature of the air fell below the calculated dew-point (which is independent of the temperature of the air leaving the continent as this was assumed dry) any snow which is formed as a result of vertical convection would arrive at the ground without any opportunity of evaporating and as a result the intensity of the snow showers should increase rather rapidly.

A value of the critical air temperature on the continent appropriate to a trajectory to Spurn Head where the calculated dew-point was 29°F. can be found as follows. Let it be T_c. Then from equation (1) the temperature of this air on arrival at Spurn would be $(0.4\ T_c + 25.2)$. If this is to be less than 29°F. (the calculated dew-point) then

$$0.4\ T_c < 3.8, \text{ i.e., } T_c < 10°\text{F.}$$

Hence 10°F. appears to be an approximation to the critical value and therefore the heavy and almost continuous snow might have been forecast on this occasion from the fall of temperature below this value on the continent.

14.10.2. The cooling of warm air over the North Sea

In an investigation of *Haars or North Sea fogs on the coast of Great Britain* Frost[48] has 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_1 is the initial temperature of the warm air as it leaves a land surface and T_2 is the temperature of the sea (assumed constant) then the temperature T_3 of the air after traversing various distances across the sea surface is given by:

$$T_3 - T_2 = (T_1 - T_2)\ f(xz),$$

where x is the distance traversed and z is the height above the surface. Taking z as 4 feet (that is, screen height) the figures in Table 21 are the values of $f(xz)$ for trajectories up to 1,000 kilometres in steps of 100 kilometres.

TABLE 21 *Values of f (xz) for trajectories up to 1,000 kilometres*

x (kilometres)	100	200	300	400	500	600	700	800	900	1,000
$f(xz)$ for z = 4 feet	.175	.152	.141	.133	.127	.123	.119	.116	.113	.110

This means that with an initial difference of 10°F. between the air and sea temperatures the difference is reduced quite rapidly to 1¾°F. after 100 kilometres track across the sea and thereafter slowly but steadily to about 1°F. with increasing distance. These figures were calculated for a wind speed of 4 metres per second (about 9 knots) but Frost considers that they are valid for wind speeds of less than Beaufort force 4 to 5 (about 12-20 knots). It is thus seen what a powerful and rapid control a cool sea surface has on temperatures at screen height. For those isobaric situations where warm air leaves a land mass, these coefficients will enable a numerical estimate of its screen temperature to be made after traversing a cooler sea surface provided a representative sea temperature is available. Where initial differences are only about 10°F. temperatures on the coasts will be within 1° or 2°F. of the sea temperatures. In fairly restricted situations where the initial temperature differences may be large (for example, a summer situation where hot continental air reaches south-east England after a relatively short sea crossing) temperature differences will be reduced to about one fifth after 100 kilometres of sea track.

Frost also gives an approximate method for estimating the height of the temperature inversion based on a modification of the formula due to G.I. Taylor: namely,

$$UZ^2 = 4\,Kx,$$

where Z is the height of the inversion, U is the wind speed, x is the fetch over the sea and K is the eddy diffusivity (roughly 3×10^3 C.G.S. units for moderate winds).

Frost calculates the following inversion heights for a wind of 8-10 miles per hour but they are probably typical of most light or moderate winds:

Inversion height	*Fetch over sea*
feet	*miles*
200	10
400	30
850	100
1,600	300
2,200	500

These values for the heights of inversions should be used only when observed values are not available.

14.11. EFFECT OF FLOW OVER HILLS

The great complexity of air flow over hills was described in Chapter 13 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 pilots 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.

14.11.1. *Föhn effect*

When a moist airstream flows over a hill barrier it rises and cools at the dry-adiabatic rate until the dew-point is reached. Thereafter cooling proceeds at the wet-adiabatic rate and cloud is usually formed. If the air is forced by the hills to ascend well above its condensation level it tends to lose some of the moisture by precipitation on the windward slopes. Assuming that in descent on the lee side the air flow more or less follows the ground contour, heating will take place at the wet-adiabatic rate until all cloud droplets have evaporated and thence at the dry-adiabatic rate. If some of the moisture content was precipitated on the windward slopes temperatures below the level at which the cloud was dispersed on the lee side will be greater than those at the same level on the windward slopes. This is the well known Föhn effect and it should be realized that the effect depends primarily on the precipitation of moisture on the windward side. Calculation of the modification to temperature requires an estimate of the amount of moisture removed on the windward slope. There is no known general rule for this and the best that can be done is to form a judgment from the topography and the stability and moisture content of the air mass to windward. An upper limit is given by assuming that, after saturation has been reached, all moisture condensed by further lifting falls as precipitation. The more spectacular Föhn effects are produced by extensive mountain ranges (for example, the Rockies or the Alps) but the Scottish and Welsh hills and the Pennines produce noticeable Föhn effects in suitable situations. Smaller masses of high ground (for example, Dartmoor) may also cause a Föhn of a local character. McCaffery[59] has described a situation with a south-south-east flow across the Cairngorms on 23 January 1952, when the temperature at Kinloss was 38°F. compared with 33°F. at Leuchars. Lawrence[60] 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 and was flowing across Scotland on the morning of 18 February. Air which had passed near Aldergrove with a screen temperature of about 52°F. reached Dallachy, Morayshire (57°39' N., 3°04' W., 37 feet) around 0300 G.M.T. 18 February when a temperature of 60°F. was reported.

A secondary effect which may accentuate temperature differences in Föhn conditions should be borne in mind. It may be that the drying out of the air on the windward slopes results in predominantly cloudy or overcast conditions to windward and to partly cloudy conditions to leeward. By day, insolation will then cause temperatures on the lee side to show a much more marked diurnal rise than those to windward.

14.11.2. *Effect of forced ascent of air*

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[61] 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 2,000 feet above mean sea level it may be assumed that the air may at times over the high ground be lifted at all levels 2,000 feet 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 many thousands of feet 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 1½). 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.

14.11.3. *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 few hundred feet 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. Judgment 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.

14.12. SELECTED SYNOPTIC EXAMPLES

The selection of examples has been biased towards the unusual rather than usual types to highlight a number of effects for example, the effect of the existence and absence of cloud cover on maximum and minimum temperatures and high temperatures associated with strongly subsided air reaching the British Isles from the continent. Other examples show an unusually cold day in spring, the reversal of the normal diurnal changes of temperature and the marked lowering of day maxima in persistently foggy areas in winter.

As a large number of examples might tend to confuse, the number has been intentionally restricted. The few cases illustrated should indicate, in a broad way, the magnitude of variations in temperature due to a few parameters and provide a valuable framework against which forecasts of temperatures can be considered.

In order to keep the synoptic illustrations as clear as possible the detail plotted for each station has been restricted on most charts to those elements which have a direct relevance to the point being illustrated. The codes and symbols used are those in force in January 1958, and a list of some of the specifications is given in Appendix I.

14.12.1. Cold type in spring with fresh easterly winds and almost complete cover of low cloud - 28 March 1952

Plate I shows the synoptic situation at 1200 G.M.T. on 28 March. To the north and west of the British Isles pressure was high; central pressures being about 1030 millibars. Very low central pressures (970 - 975 millibars) existed in two depressions near Spain and north-east of the Azores, and an elongated belt of low pressure extended from the Azores through western continental Europe to western Russia. With this distribution of pressure there existed a strong easterly gradient across almost the whole of the British Isles. During the night 26-27 March an occlusion had moved west-south-west across the British Isles and had cleared all areas except Devon and Cornwall by 0600 G.M.T. on 27 March.

During the day on the 27th surface winds were generally between north-north-east and east-north-east, force 2 to 4. In the morning there were substantial breaks in the clouds over southern, central and western England. In these areas there was a noticeable diurnal rise of temperature to about the mid forties. There were some clear skies, particularly in central and western districts of England during the night 27-28 March but during the morning of the 28th almost all stations in England were reporting a cover of ¾ or more of low cloud with bases between 1,500 and 2,500 feet. Surface winds had veered to between north-east and east and were generally stronger, being between force 4 and 6.

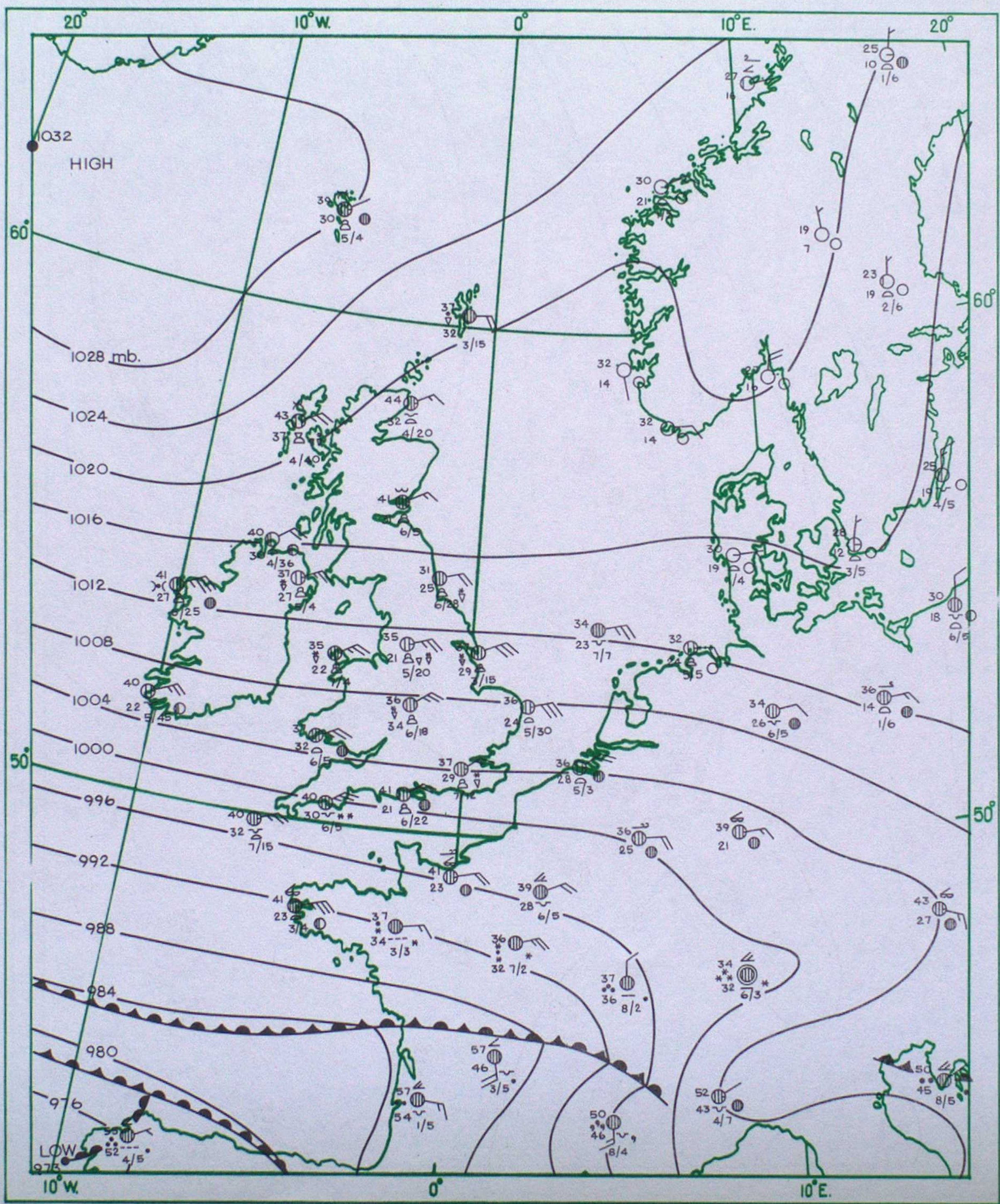

PLATE I Synoptic situation, 1200 GMT, 28 March 1952

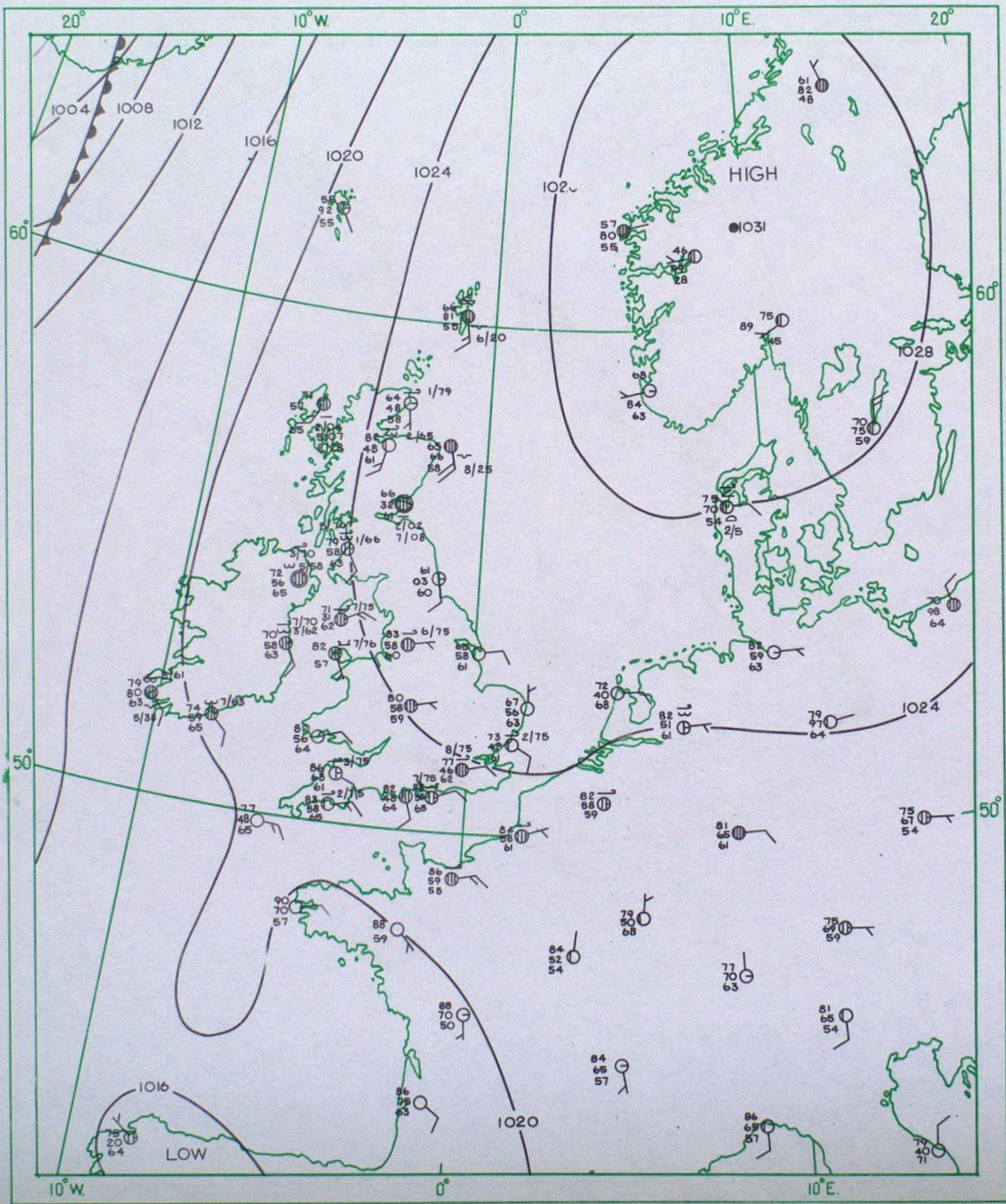

PLATE II Synoptic situation, 1200 G.M.T., 23 August 1955

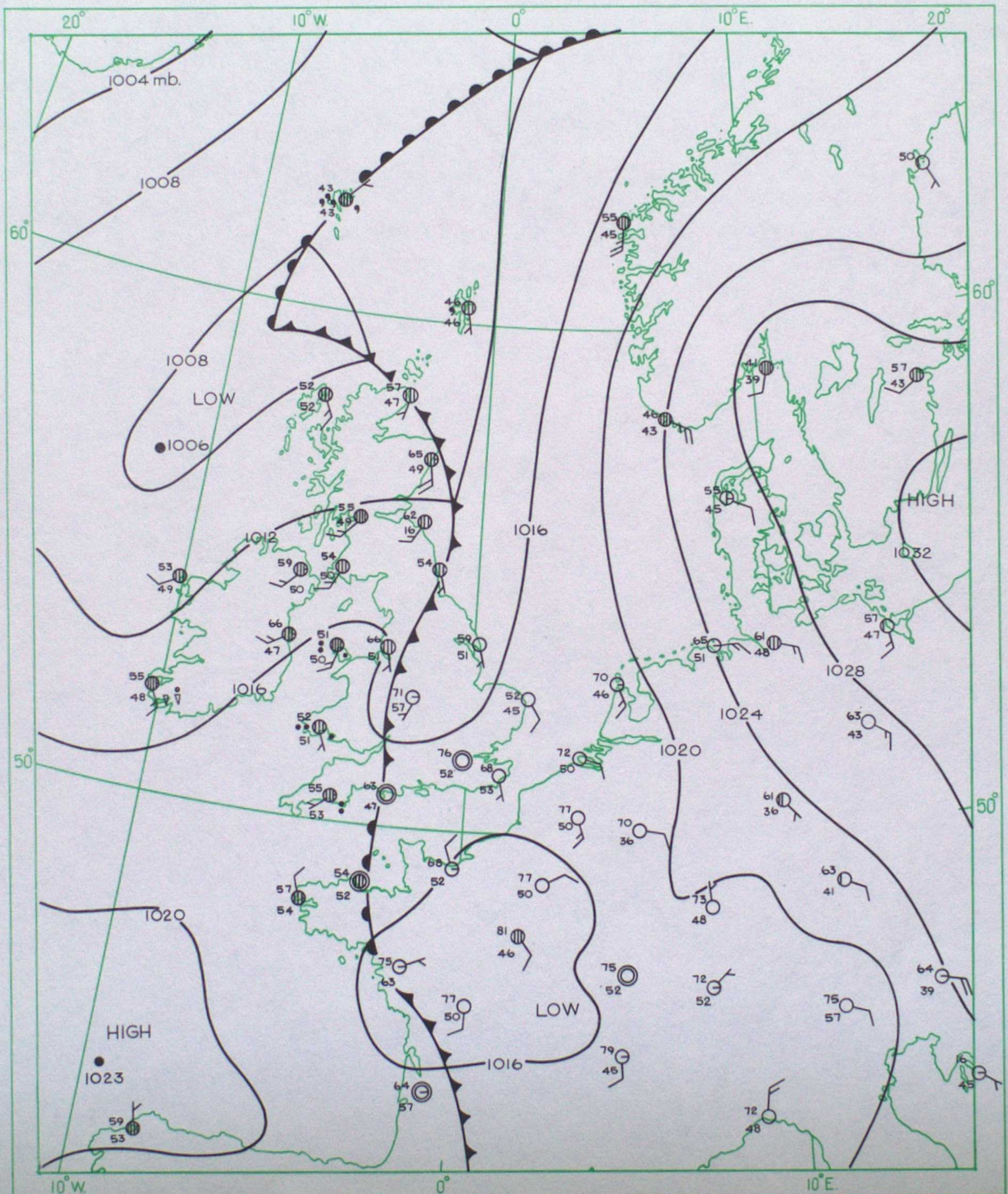

PLATE III Synoptic situation, 1200 G.M.T., 16 April 1949

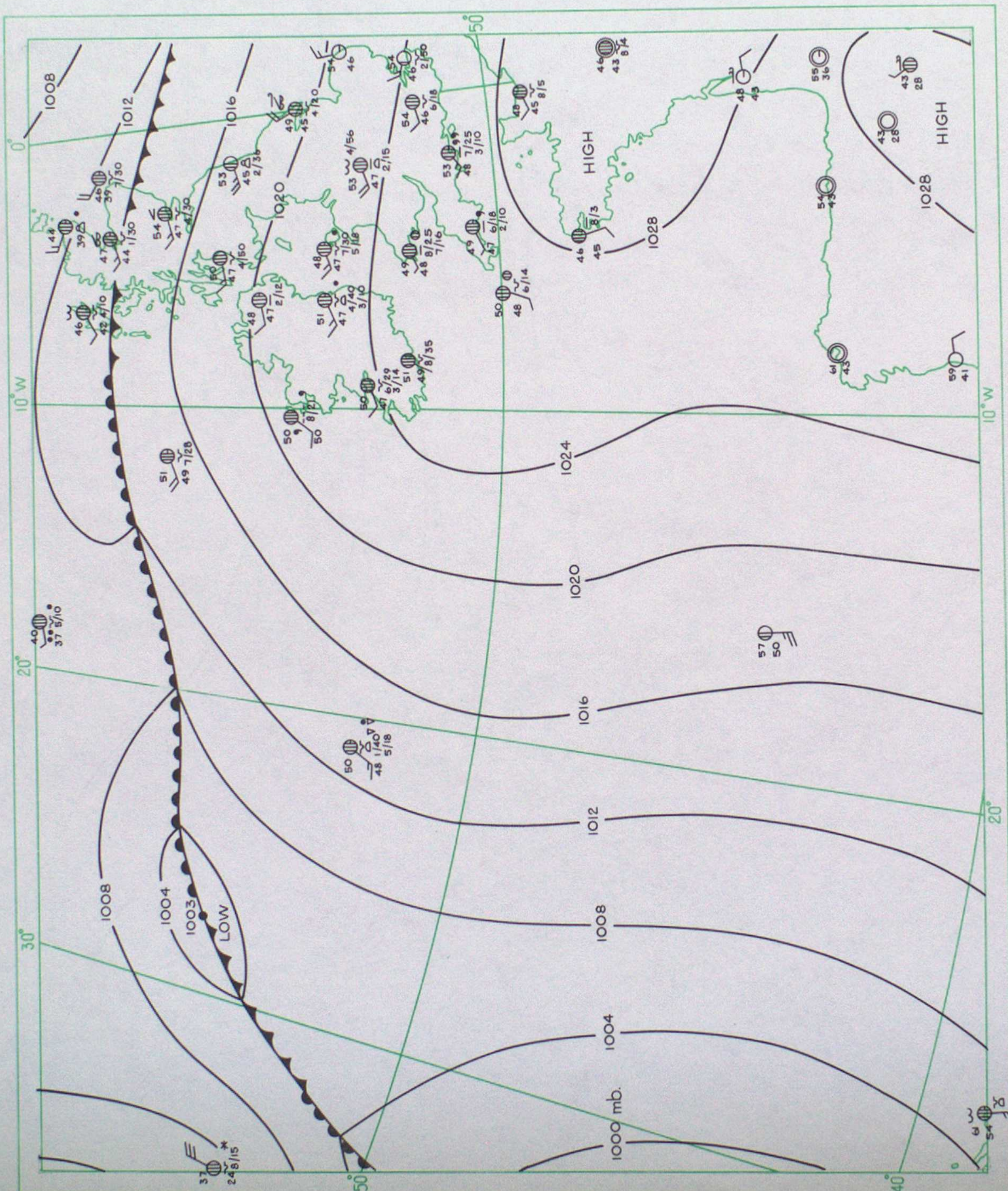

PLATE IV Synoptic situation, 1200 GMT, 29 January 1953

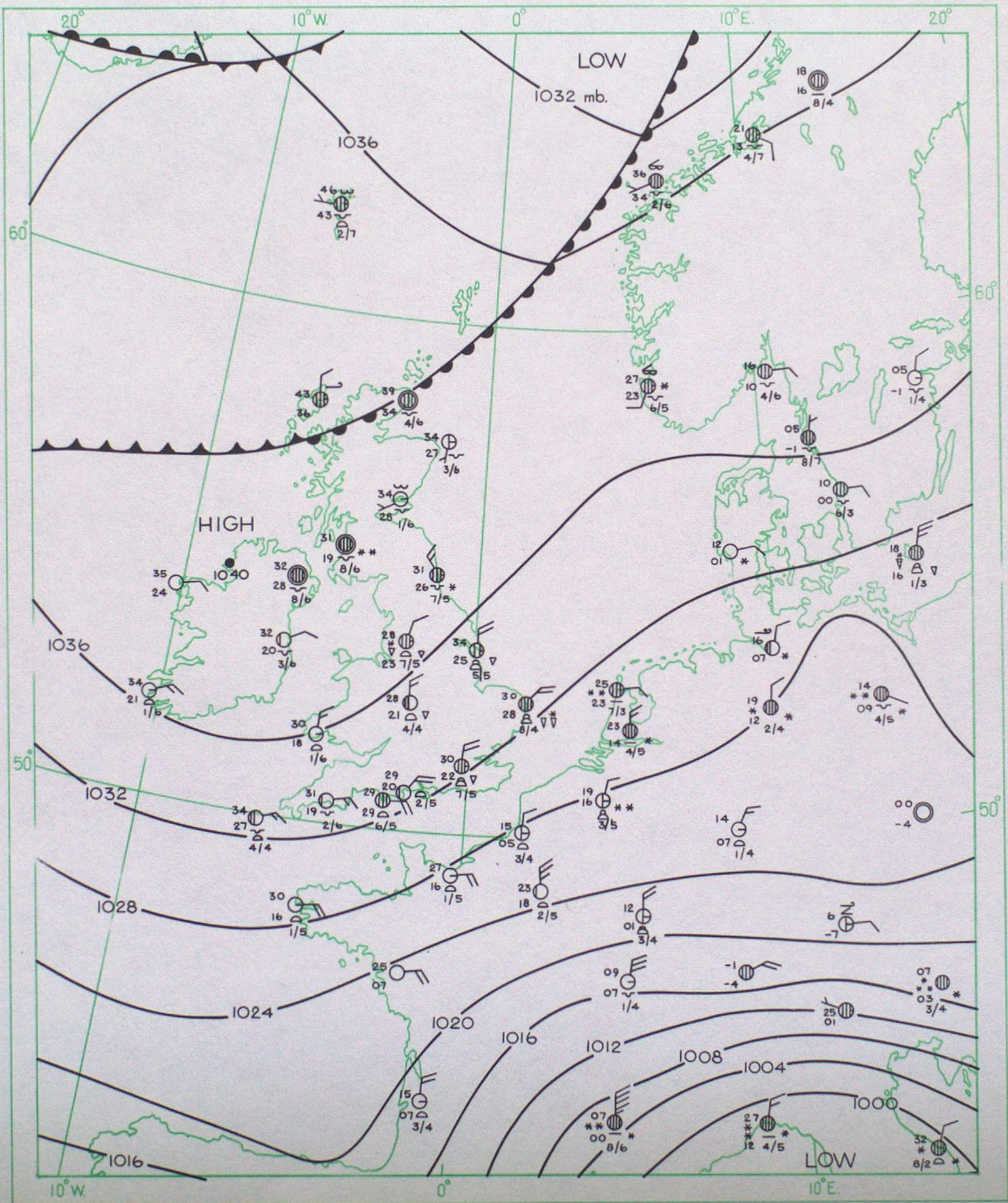

PLATE V Synoptic situation, 1200 G.M.T., 10 February 1956

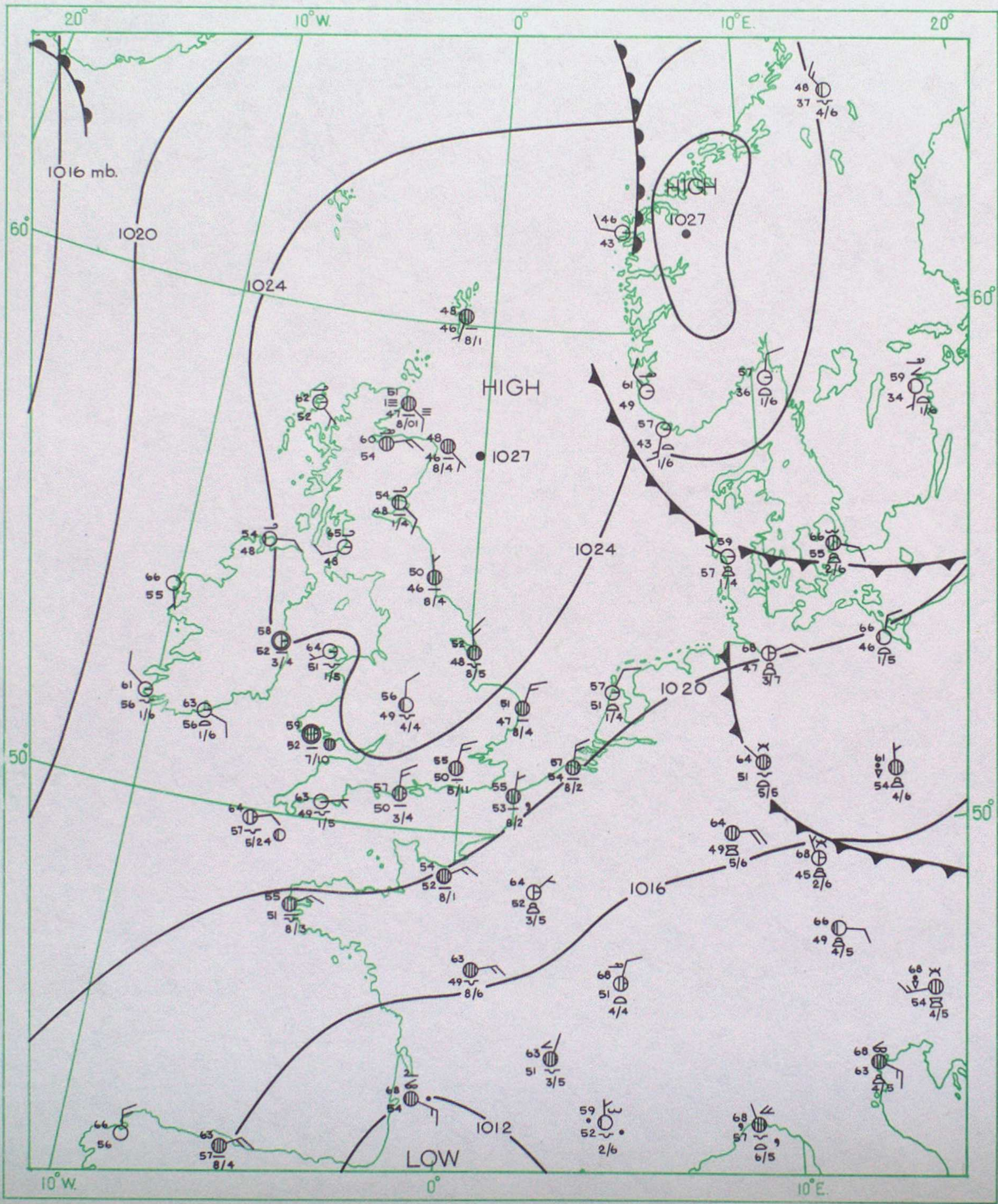

PLATE VI. Synoptic situation, 1200 G.M.T., 3 June 1954

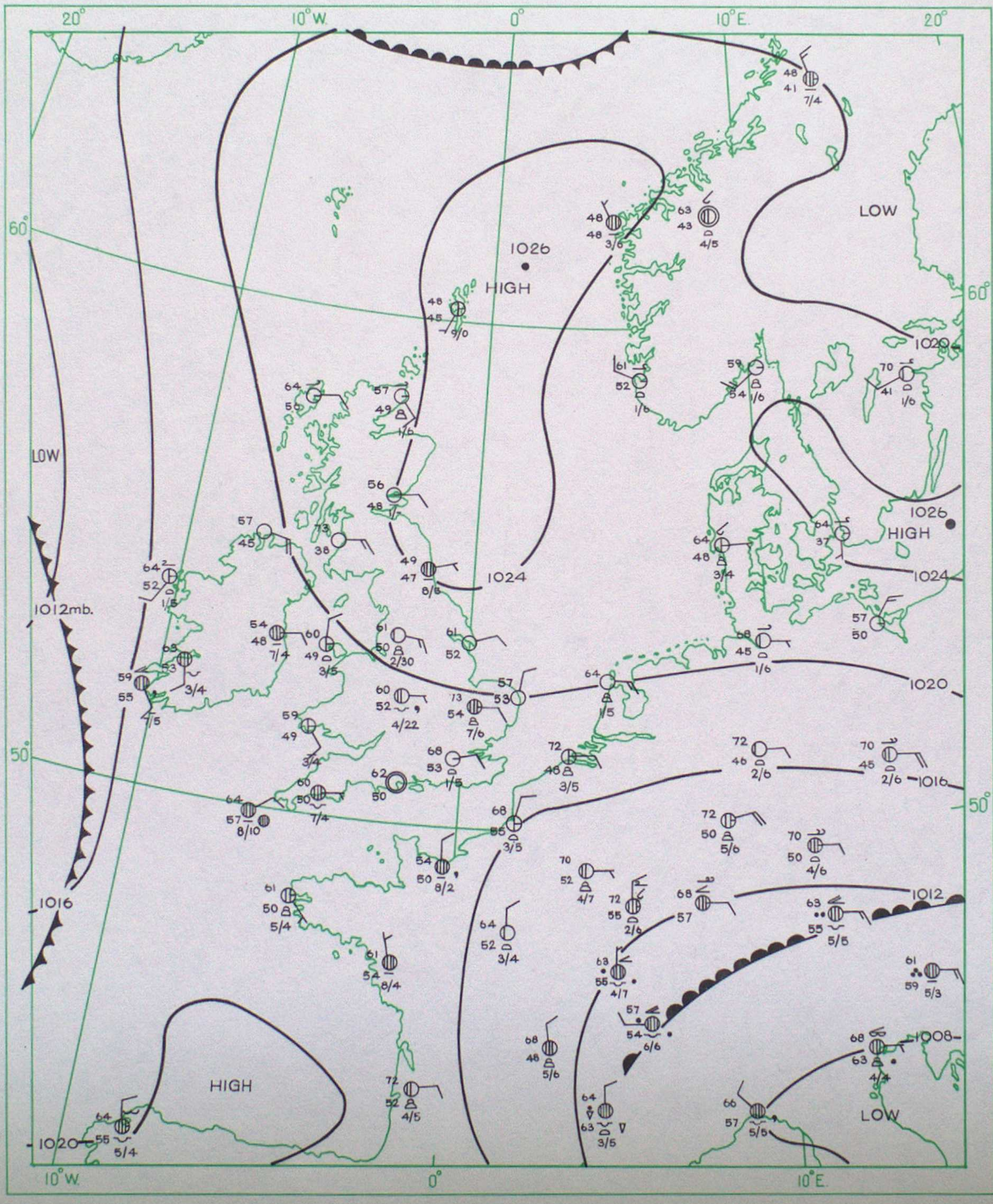

Plate VII Synoptic situation, 1200 G.M.T., 4 June 1954

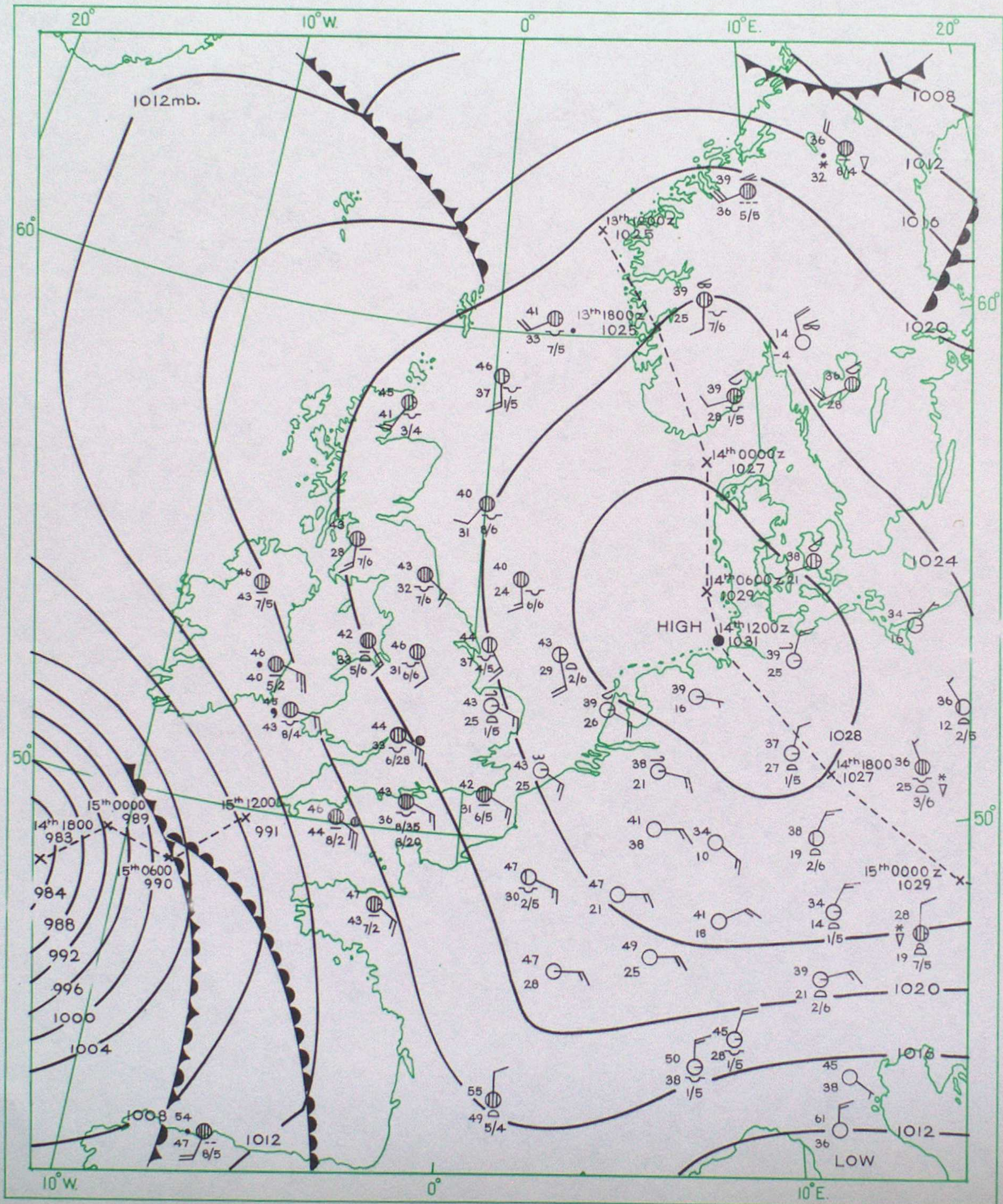

PLATE VIII Synoptic situation, 1200 G.M.T., 14 March 1952

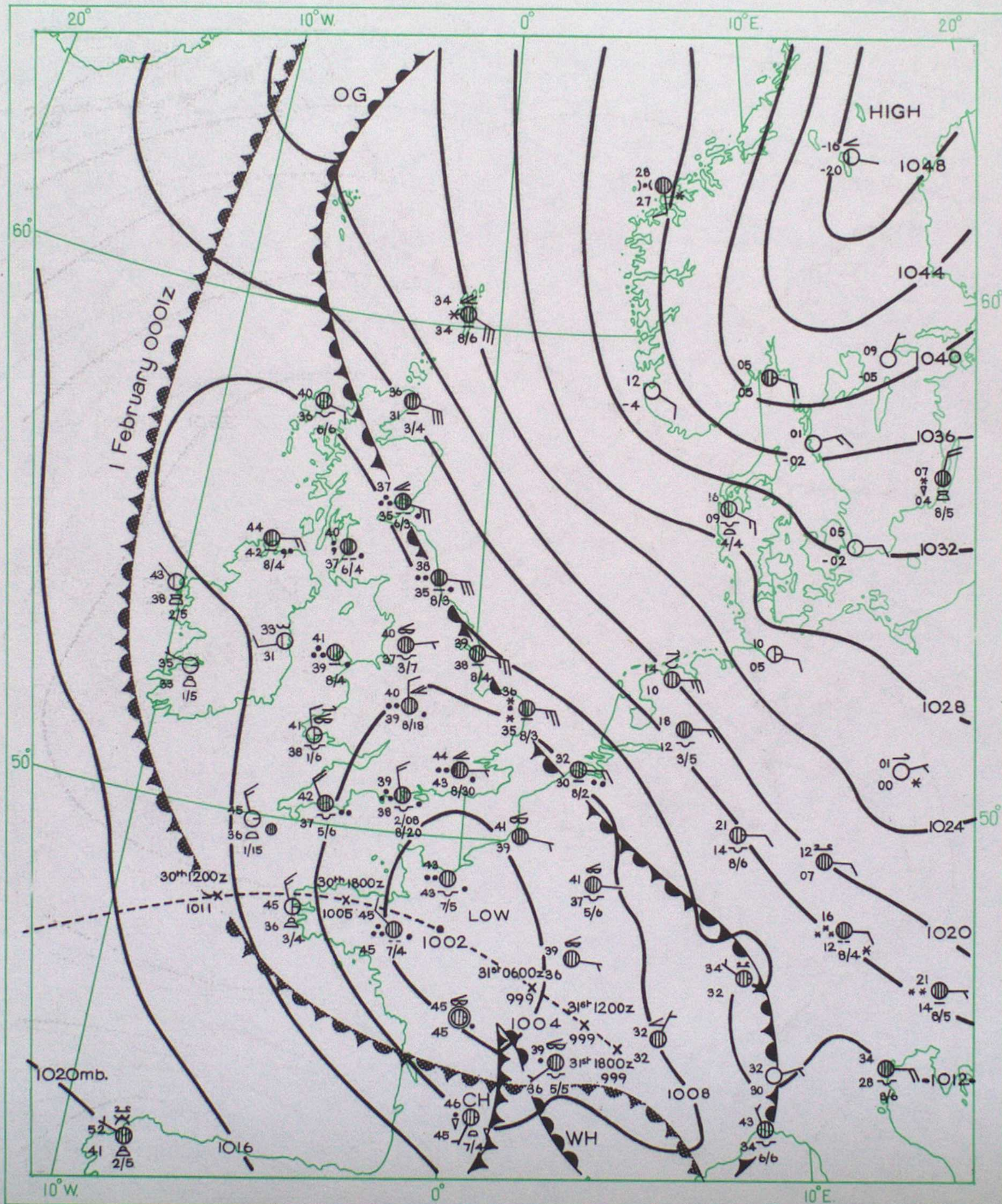

PLATE IX Synoptic situation, 0001 G.M.T., 31 January 1956
and frontal positions, 0001 G.M.T., 1 February 1956

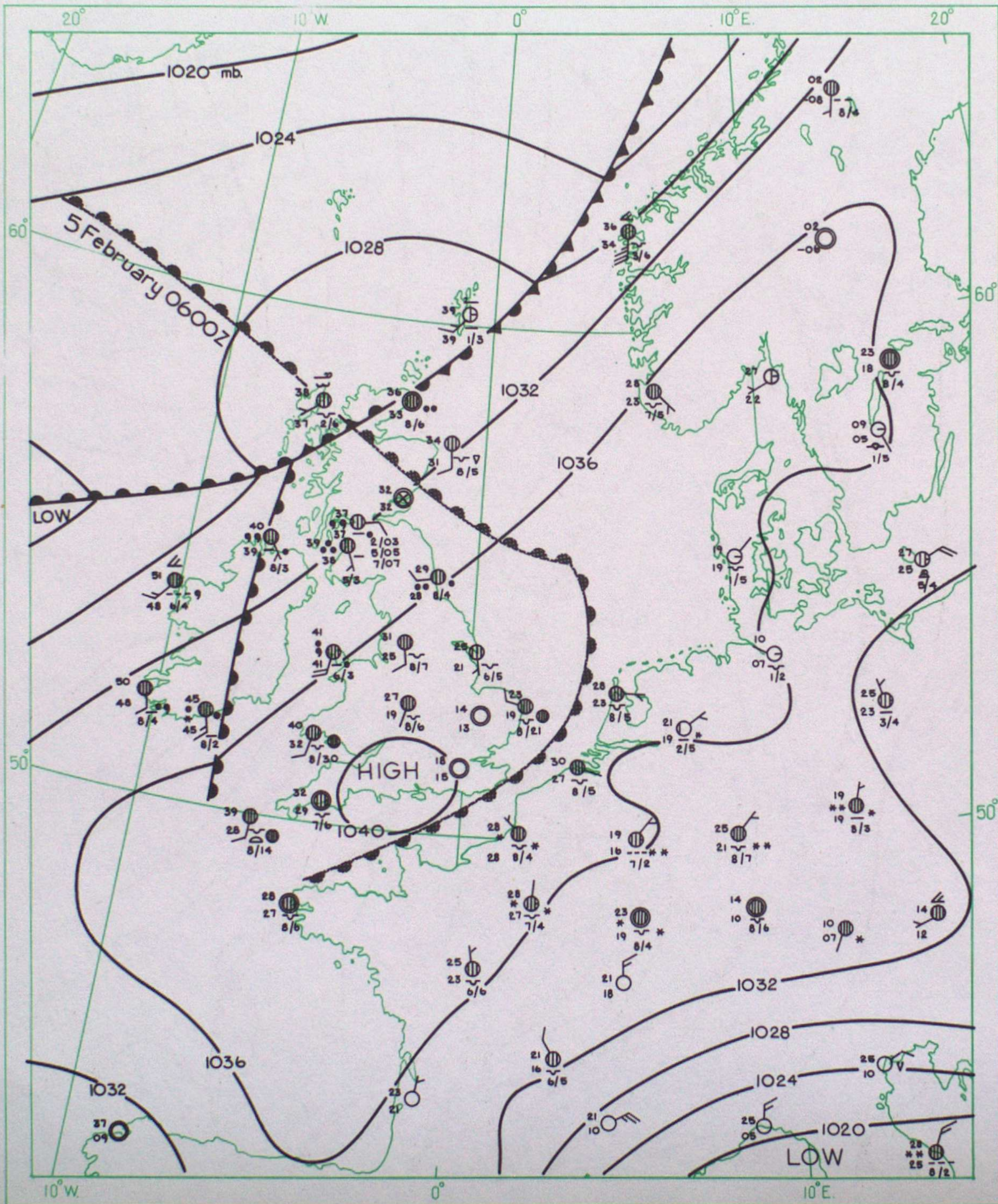

PLATE X Synoptic situation, 0600 G.M.T., 4 February 1956 and frontal positions, 0600 G.M.T., 5 February 1956

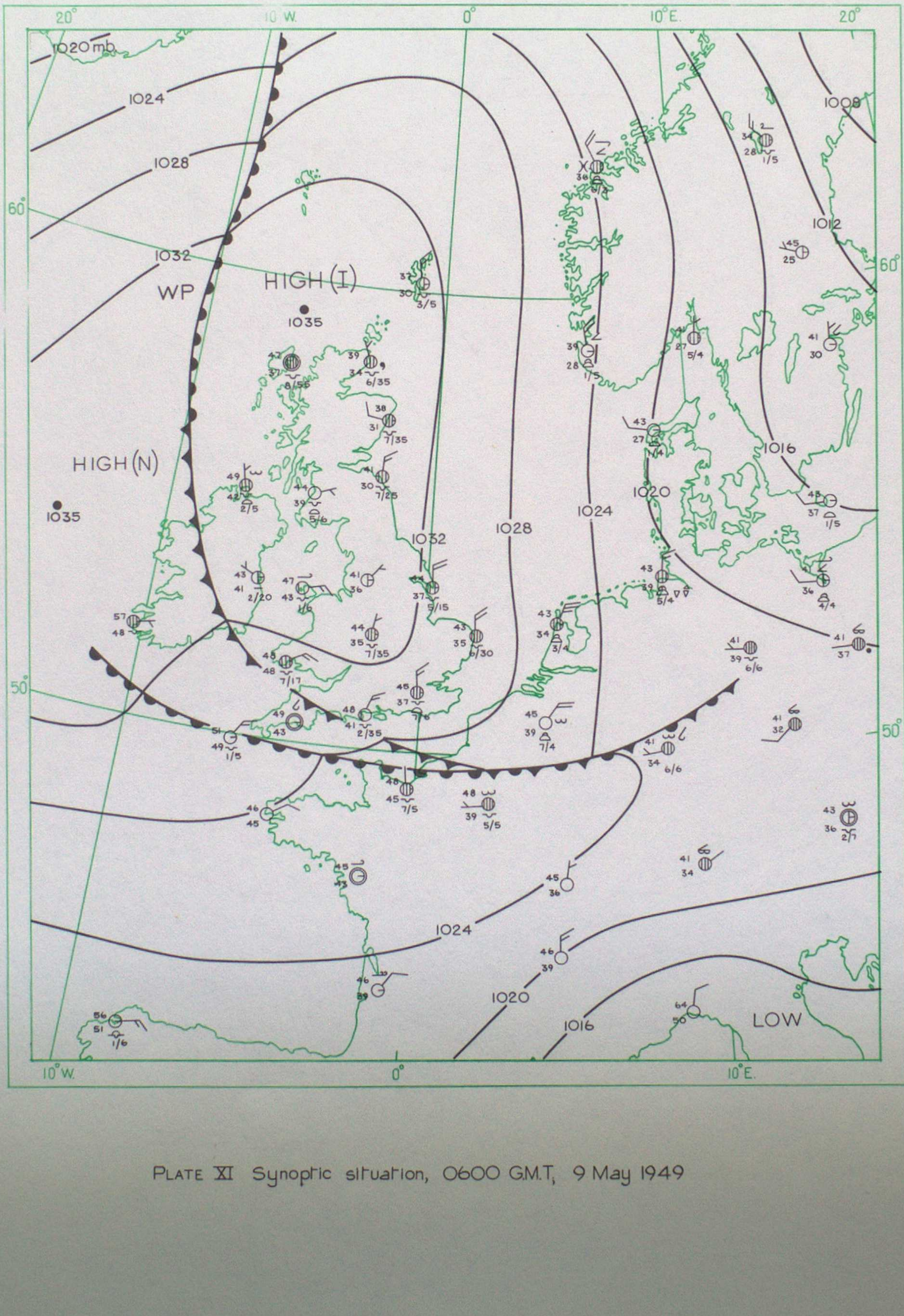

PLATE XI Synoptic situation, 0600 G.M.T., 9 May 1949

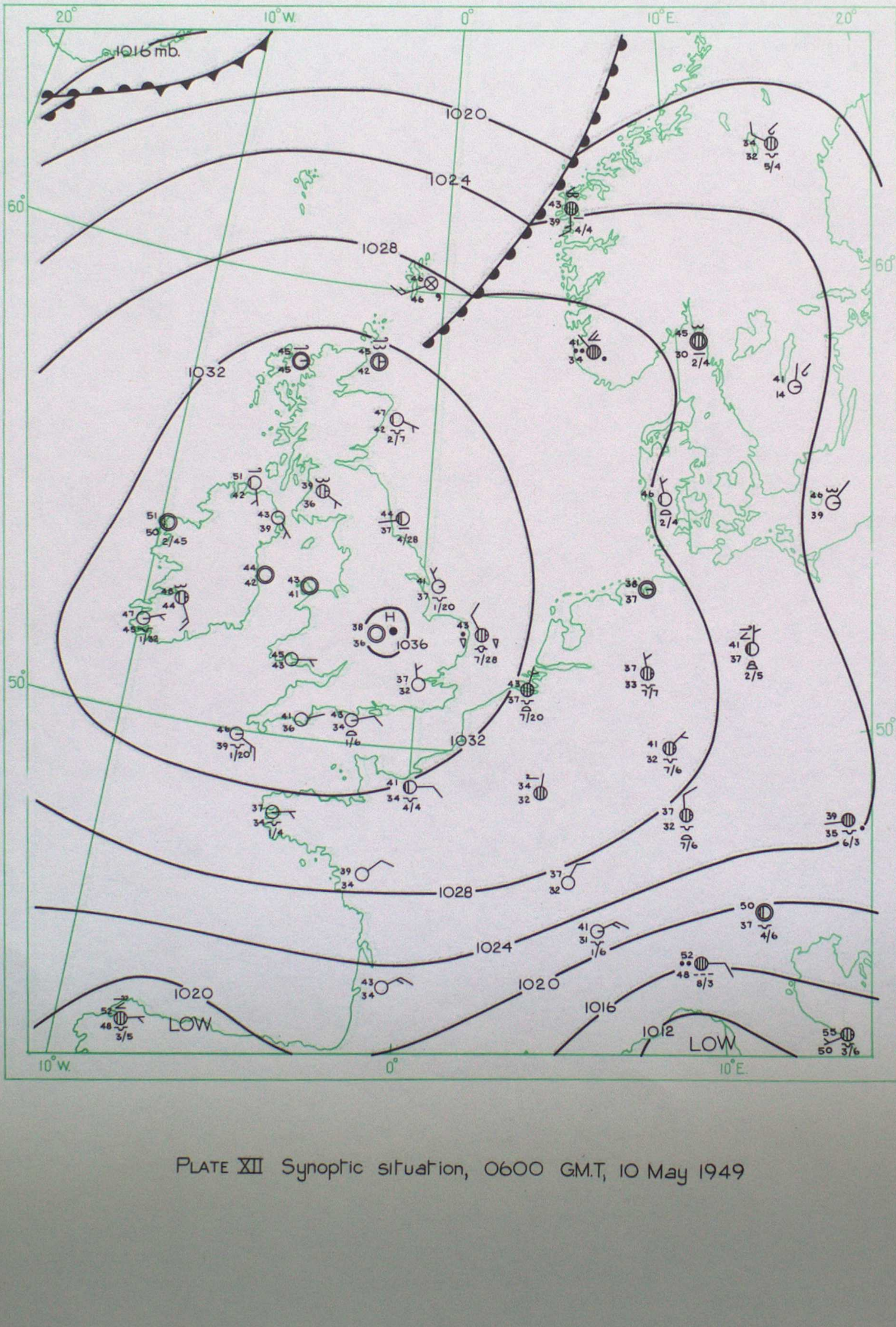

PLATE XII Synoptic situation, 0600 G.M.T., 10 May 1949

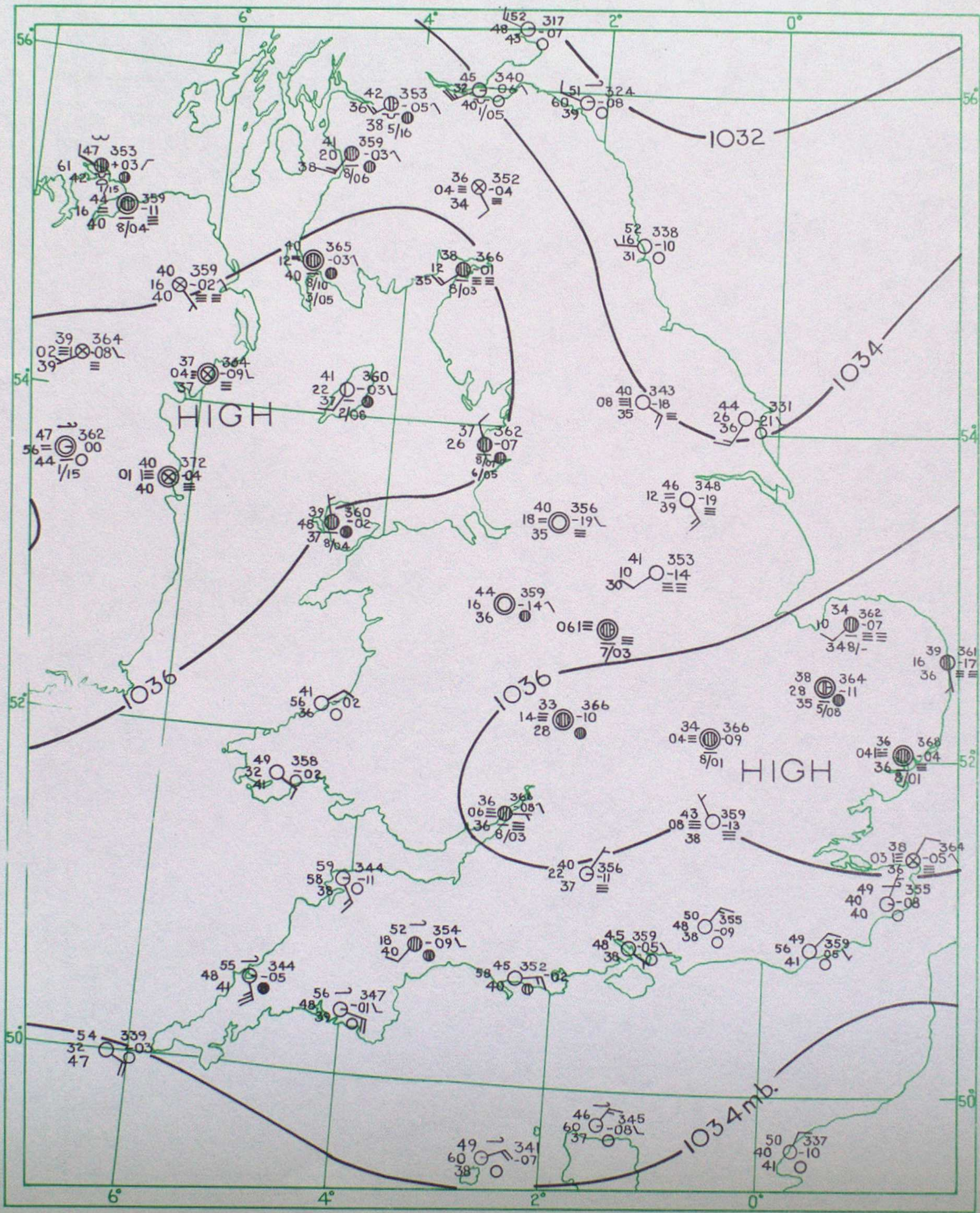

PLATE XIII Synoptic situation, 1500 G.M.T., 2 March 1953

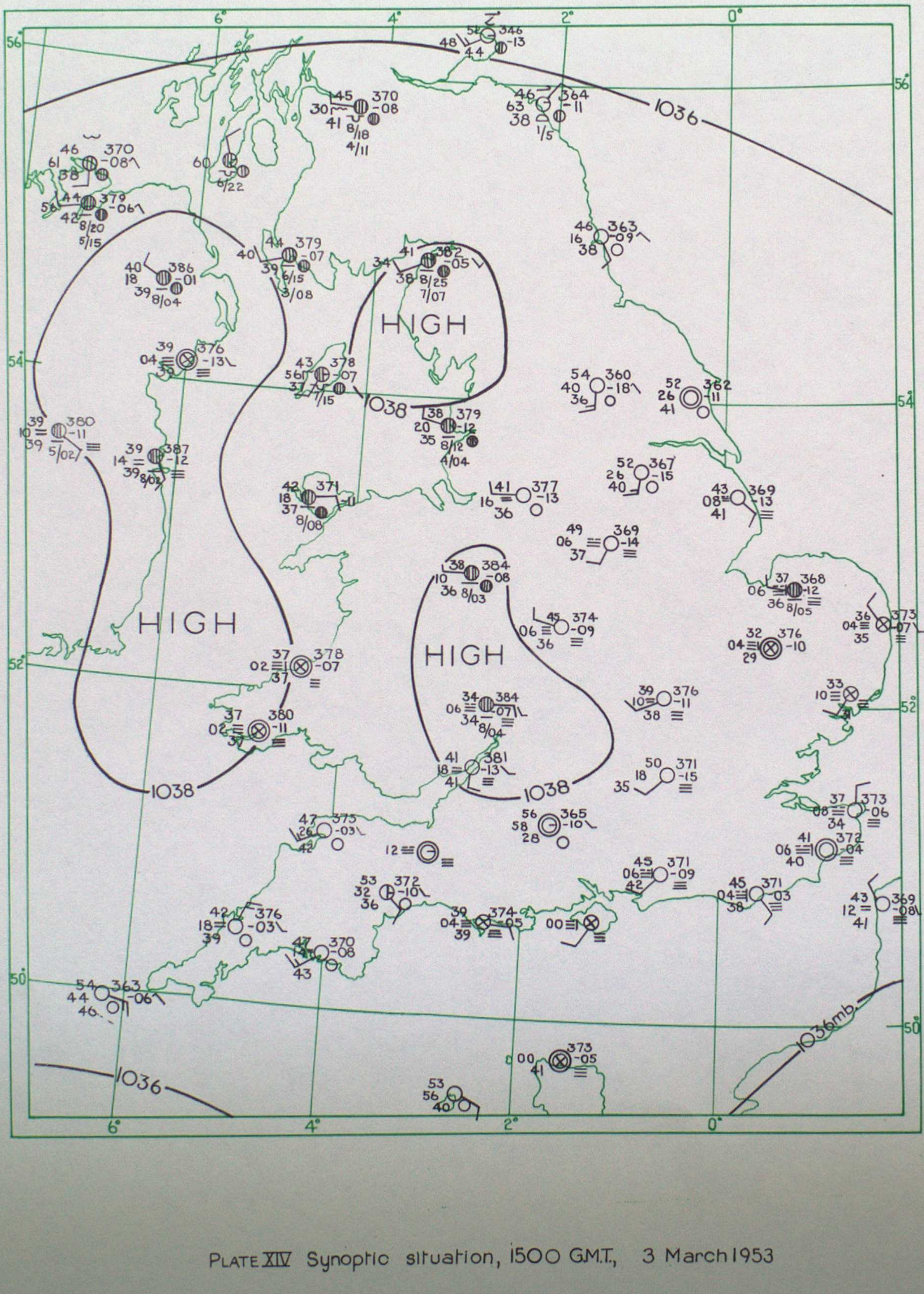

PLATE XIV Synoptic situation, 1500 G.M.T., 3 March 1953

Scattered snow showers had been reported on the 27th but by the 28th snow showers were very frequent and widespread. The combined effect of increased cloud cover, stronger winds and cooling by precipitation restricted day maxima on the 28th to the mid thirties over extensive areas in England. In extreme western districts of England and Wales day maxima just exceeded 40°F. and in north and west Scotland day maxima on the 28th were not significantly different from those on the 27th.

The night 28-29 March continued cloudy with snow showers but during the 29th continuous snow associated with a depression over north-west France spread northwards to affect most of England south and east of a line from the Wash to the Bristol Channel. Winds increased to gale force. In south-east England the night 28th-29th had been less cold generally than the 27th-28th but some very low day maxima were recorded on the 29th. 32°F. was the maximum at London Airport and the 33°F. recorded at Kew was the lowest value in late March since records began in 1871. Table 22 shows maximum and minimum temperatures recorded at a selection of stations in the United Kingdom.

TABLE 22 *Maximum and minimum temperatures 26-28 March 1952*

	Minimum temperature		*Maximum temperature*	
	night 26th-27th	*night 27th-28th*	*day 27th*	*day 28th*
	°F.	°F.	°F.	°F.
London Airport	30	30	43	37
Gorleston	31	32	42	38
Elmdon	25	27	42	37
Plymouth	35	30	44	41
Pembroke Dock	35	31	46	40
Valley	31	32	45	39
Squires Gate	27	27	49	41
Leuchars	33	31	47	42
Renfrew	35	32	46	42
Aldergrove	29	33	44	40
Wick	32	33	45	44
Stornoway	37	35	44	44

Figure 18 is a reproduction of the thermogram for London Airport for 27-29 March 1952 and illustrates the effect of increased cloud cover, stronger winds and snow on the amplitude of the diurnal range of temperature. On 27 March, a

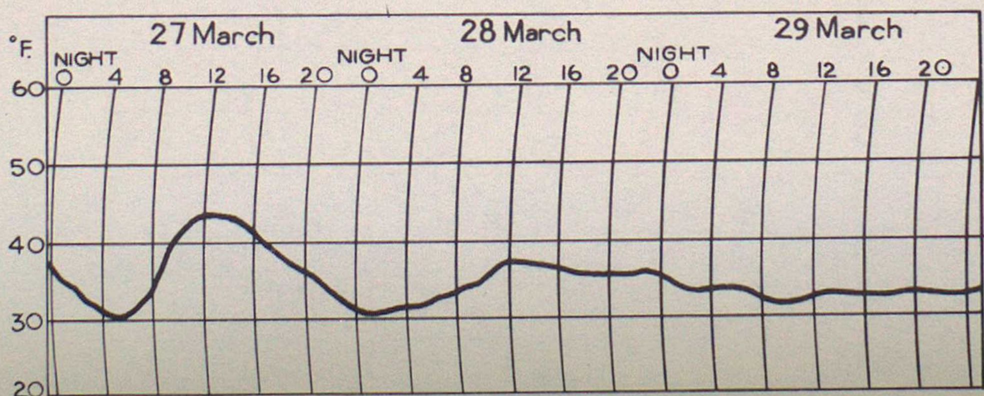

FIGURE 18 *Thermogram for London Airport, 27-29 March 1952*

day of broken cloud and moderate winds, the diurnal variation was well marked. Stronger winds, increased cloud cover and some snow showers restricted the diurnal variations on the 28th. During the 29th winds increased to gale force, there was continuous snow and, after midday, temperatures were almost uniform (around 32°F.).

14.12.2. Very warm type with light winds from continent - 23 August 1955

Plate II illustrates the synoptic situation at 1200 G.M.T., 23 August 1955. Since 21 August pressure had been high generally in an extensive belt from the British Isles across the German Plains and Scandinavia to the northern parts of Russia in Asia. On the 23rd the anticyclonic cell centred over Scandinavia was almost stationary. Air which was reaching the British Isles on the 23rd had a long continental land track due to the extensive east-south-east gradient which had existed over much of continental Europe on the south side of the anticyclonic belt. The tephigrams for Crawley and Hemsby for 0200 G.M.T., 23 August are shown in Figure 19 and are noteworthy for the general dryness of the air up to at least 500 millibars. Application of Gold squares indicates maxima in the upper eighties.

Although there was rather a lot of high cloud throughout much of the day, temperatures had reached or exceeded 80°F. by noon almost everywhere in England and Wales except at a few places near the coasts, notably the east coast. Day maxima inland were generally between 80° and 90°F.

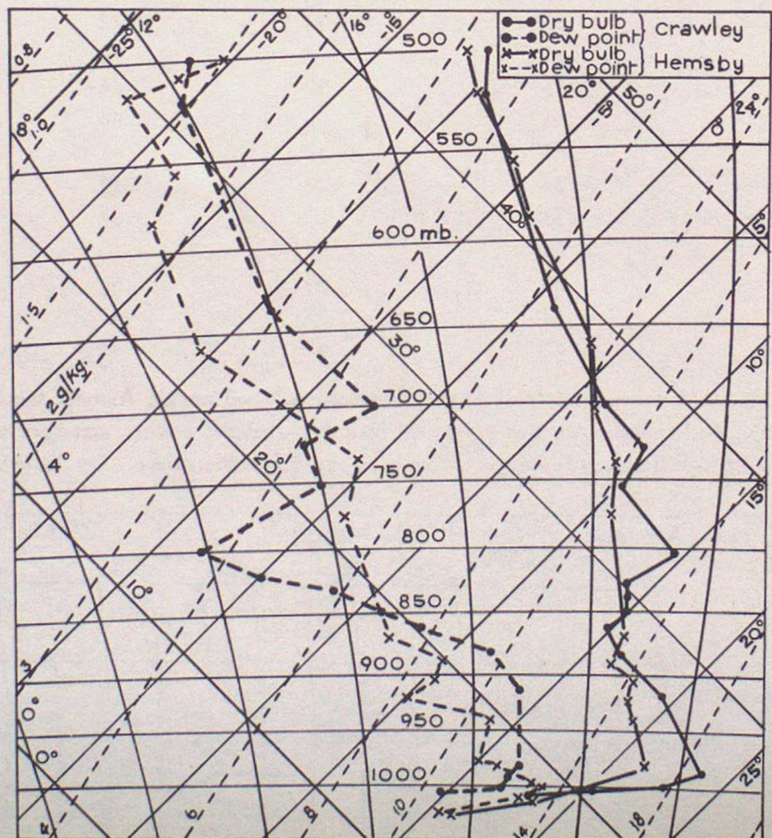

FIGURE 19 *Tephigrams for Crawley and Hemsby, 0200 G.M.T., 23 August 1955*

Temperature

A selection of thermograph traces for 23 August 1955 is shown in Figure 20. London Airport and Elmdon are typical of inland stations (*a*). It will be seen from Plate II that the gradient wind is onshore at Felixstowe and offshore at Chivenor. The marked difference in the performance of temperature at both places is clearly shown in Figure 20(*b*). At Tangmere where the gradient wind is partly onshore temperatures were mainly intermediate between Felixstowe and Chivenor. Figure 20(*b*) shows that there should be no lowering of daytime maxima at coastal stations on a "rule of thumb" basis. Each synoptic situation and locality must be considered individually. It is noteworthy that, on 23 August, the highest maximum temperature (90°F.) was recorded at Chivenor.

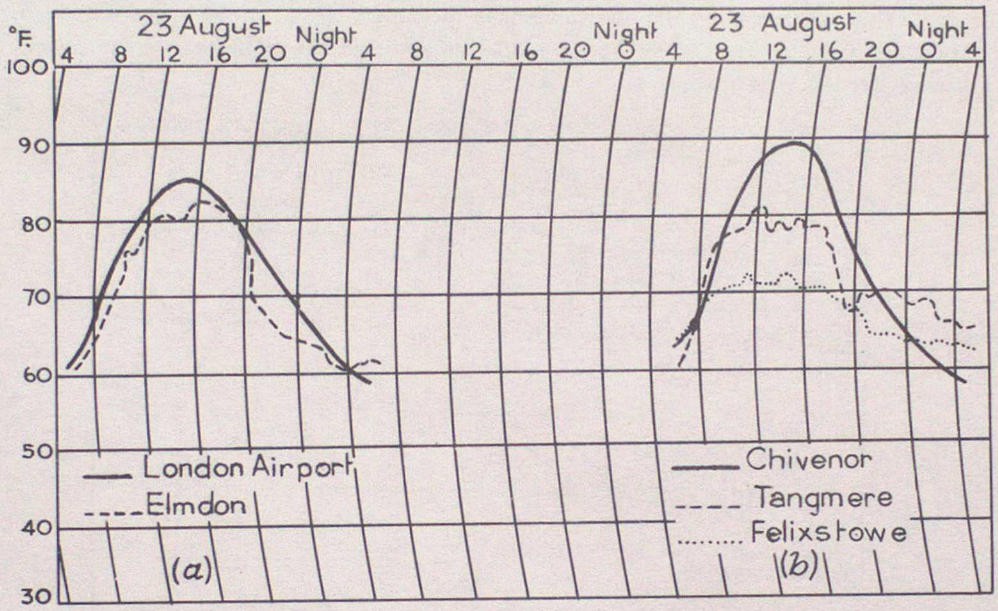

FIGURE 20 *A selection of thermograms for 23 August 1955*
(*a*) Inland stations (*b*) Coastal stations

In late spring, summer and early autumn light south-easterly drifts from the continent and strongly subsided air (implying an absence of low and medium cloud and of convection through any appreciable depth) can usually be relied upon to produce high maxima over substantial areas of the British Isles. On some occasions this type of situation is accompanied by a cold frontal trough often extending in a north - south direction from western areas of the British Isles well into France. The trough is often moving only slowly in a general eastward direction and its movement is frequently difficult to estimate. On some occasions there is an outbreak of thunderstorms some distance ahead of the trough and in areas where thunderstorms occur day maxima are usually moderated.

Some other occasions of high temperature with light winds from the continent were:

(i) 12 August 1953 when some maxima in England reached 92°F.

(ii) A four day spell: 26-29 July 1948. The duration of this spell was noteworthy as well as the very high maxima (90°-94°F.) at several stations in south-east England on the 28th.

(iii) 16 April 1949, when maxima in England reached the eighties. This example was unusually early in the year and it also illustrates a trough extending from the western United Kingdom to France. The 1200 G.M.T. chart for 16 April is shown in Plate III.

14.12.3. Very mild type in winter with some breaks in the cloud - 29 January 1953

Plate IV shows the synoptic situation at 1200 G.M.T. on 29 January. Pressure was high to the south of the British Isles and low in mid-Atlantic. The frontal system separating the cold polar air from the warm tropical air was situated in the extreme north of Scotland and was slow-moving. Isobars over the United Kingdom showed noticeable anticyclonic curvature. During the night 28th-29th minima were generally in the mid or upper forties. On the 29th there were some breaks in the low cloud in some districts and a few stations recorded more than 5.0 hours of sunshine. Maxima were generally in excess of 50°F. and several stations recorded 56°-57°F.

Tephigrams for Larkhill and Camborne for 0200 G.M.T. on 29 January are shown in Figure 21. The marked subsidence and dryness of the air in the lower troposphere is noteworthy.

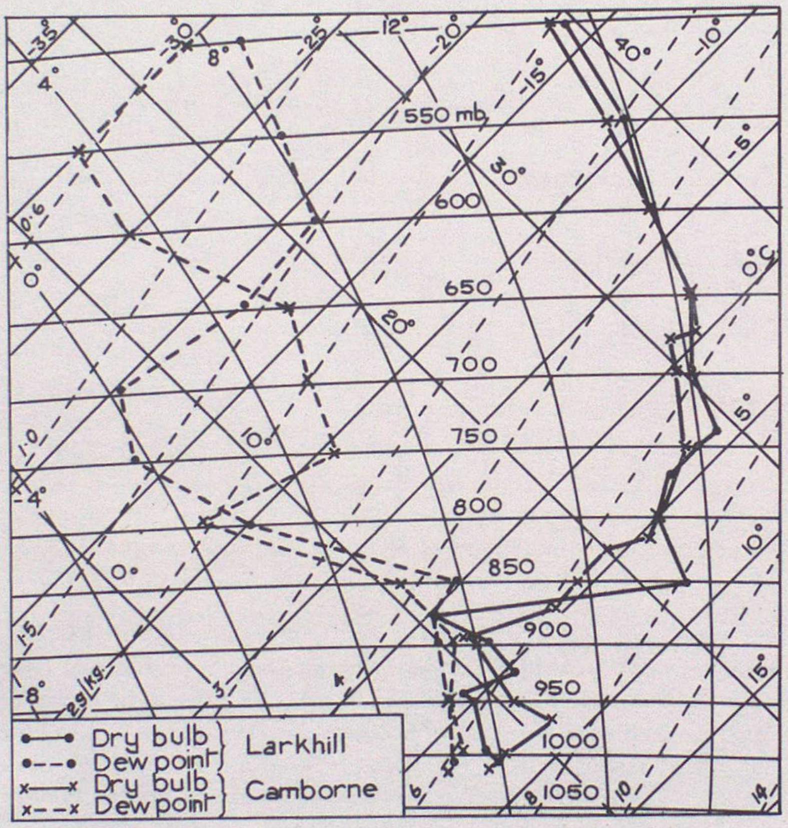

FIGURE 21 *Tephigrams for Larkhill and Camborne, 0200 G.M.T., 29 January 1953*

14.12.4. Effect of the North Sea on warming cold continental air in winter - 10 February 1956

Plate V shows the synoptic situation at 1200 G.M.T. on 10 February. Pressure had been low in the western Mediterranean and high in an extensive belt from the British Isles to Scandinavia for several days. An east-north-east gradient covered east and south-east England and the air reaching England was leaving the continent with temperatures between about 10° and 25°F. Application of the rules given in Section 14.10.1. indicate that temperatures on the east coast should be between about 29° and 35°F. and at 0600, 1200 and 1800 G.M.T. recorded temperatures at Gorleston and Felixstowe were between 28° and 30°F.

These temperatures were reasonably representative of day-time temperatures in central and southern England and indicate the degree of warming by the North Sea. Day-time temperatures in neighbouring continental stations uninfluenced by maritime effects were several degrees lower in spite of prolonged almost clear skies at some stations. At night continental minima were several degrees below values in England. Table 23 shows some values for selected stations.

TABLE 23 *Minimum and maximum temperatures on 10 February 1956 at a selection of stations in north-west Europe*

	Screen minimum, 9th-10th	*Screen maximum, 10th*
	°F.	°F.
London Airport	24	31
Gorleston	27	33
Elmdon	19	31
Spurn Head	26	35
Tynemouth	28	35
Paris (Le Bourget)	12	27
Brussels	10	23
Amsterdam	16	25
Dusseldorf	07	21
Hanover	01	19

14.12.5. Effect of variable cover of low stratus cloud on consecutive days in an east-north-easterly situation 3-4 June 1954

This example is selected to illustrate the degree of control exerted on day-time temperatures by North Sea stratus cloud. It also illustrates the importance of accurate timing of cloud dispersal by insolation - usually a difficult problem on the forecast bench.

Plates VI and VII show the synoptic situation at 1200 G.M.T. on 3 and 4 June 1954. The anticyclone over the North Sea dominated the weather over the United Kingdom during the 3rd and 4th. There was very little change in the isobaric pattern except a suggestion that some air reaching extreme south-east England on the 4th might have had a continental track from the Low Countries. From the point of view of forecasting temperatures the significant difference between the two days was the persistence of extensive low stratus at many stations in eastern and central England until after midday on the 3rd and its early clearance on the 4th. The late clearance on the 3rd restricted sunshine and the amount of heat reaching the ground, which was available for warming the lower layers of the atmosphere, was correspondingly restricted. During the night 3rd-4th low stratus reformed quite rapidly, generally by 2100 G.M.T., in many places where the late afternoon and evening had been clear. In western England and Wales stratus did not reform generally until after midnight and a few stations in these areas remained cloudless all night.

On the morning of the 4th low stratus began to break much earlier and over more extensive areas in England than on the 3rd. In many places stratus had dispersed by 0900 G.M.T. and there was prolonged sunshine but, as seems common in north-easterly types, there was some variation in local behaviour and a few stations, mainly in central England, had retained about half cover of low cloud until midday. In south-west England a sheet of stratocumulus (base 1,500-2,500 feet) formed under the influence of a depression slowly approaching Scilly from the south-west and restricted day-time maxima. At a few places in England isolated thunderstorms broke out after 1500 G.M.T. The reason for the earlier break up of the stratus cover is to be found on the tephigrams. Tephigrams for Hemsby and Crawley at 0200 G.M.T. on 3 and 4 June are reproduced in Figure 22. At Hemsby the low-level inversion extended from ground level to about 950 millibars and the temperature at the top of the inversion (950 millibars) was 13°C. (55°F.) on the 4th against 9.5°C. (49°F.) at about the same pressure level on the 3rd. On 4 June the air was wetter rather than drier in the levels below 850 millibars but the high dew-point at 900 millibars seems rather curious. Nevertheless

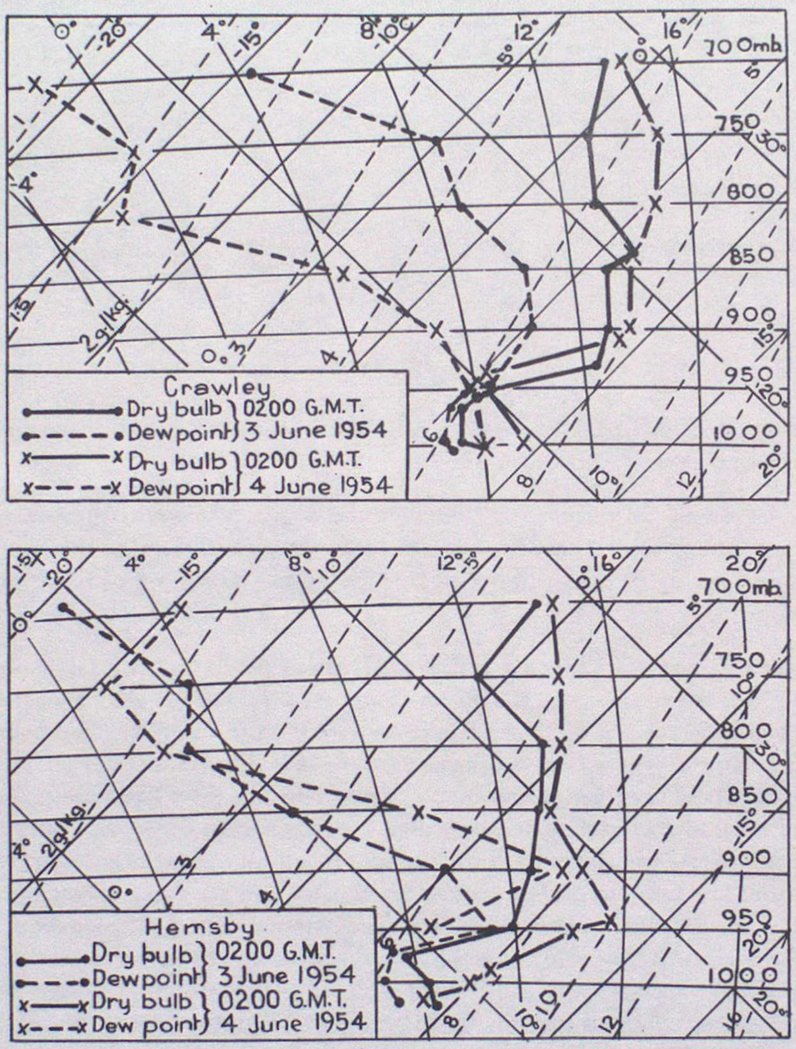

FIGURE 22 ***Tephigrams for Hemsby and Crawley, 0200 G.M.T., 3 and 4 June 1954***

the tephigrams indicate that temperatures in the mid fifties would not quite disperse the stratus on the 3rd but that on the 4th a temperature in the low fifties would suffice. At Crawley the changes in the upper air were rather different. On the 4th the inversion was no more pronounced than on the 3rd and indeed was at rather low temperatures. There was, however, noticeable drying out of air from 950 millibars to at least 800 millibars. Stratus cloud would therefore be eroded away from the top by dilution through turbulence from the very dry air above. A temperature of about 52°F. would in any case disperse the stratus by heating from below. The net result was that over England the stratus dispersed much earlier on the 4th and there was prolonged sunshine.

The differences in the modifications to the fine structure of temperature and dew-point in stagnant anticyclonic conditions as shown on these two ascents on two consecutive mornings give an indication of the difficulty of accurately forecasting such changes and the profound effect on the weather experienced.

Table 24 shows some maximum and minimum temperatures and the duration of sunshine on 3 and 4 June 1954 at a selection of stations.

TABLE 24 *Temperatures and duration of sunshine 3-4 June 1954*

	Minimum 2-3 June	*Maximum 3 June*	*Duration of sunshine 3 June*	*Minimum 3-4 June*	*Maximum 4 June*	*Duration of sunshine 4 June*
	°F.	°F.	*hours*	°F.	°F.	*hours*
London Airport*	47	63	4.8	50	74	12.1
Lympne	48	59	6.4	47	70	10.8
Tangmere	47	64	5.8	45	70	10.9
Boscombe Down	45	61	6.4	48	69	5.6
Gorleston	49	53	0.3	50	60	13.1
West Raynham	47	52	0.5	45	70	10.2
Mildenhall	48	58	3.8	50	74	11.6
Elmdon	40	64	8.4	48	69	7.1
Shawbury	47	70	8.6	49	70	6.6
Plymouth	52	64	9.7	47	61	4.7
Pembroke	49	63	6.7	48	61	7.7
Valley	48	69	13.6	51	63	13.2
Spurn Head	46	53	0.3	49	62	11.3
Tynemouth	47	50	0.0	44	51	2.5
Leuchars	46	55	8.0	43	56	12.7
Renfrew	45	72	10.8	45	71	14.3
Wick	45	51	0.0	45	58	15.3

* Sunshine at Kew

Thermograms for London Airport and West Raynham for the period 3-4 June 1954 are reproduced in Figure 23. With only 0.5 hour of sunshine on the 3rd,

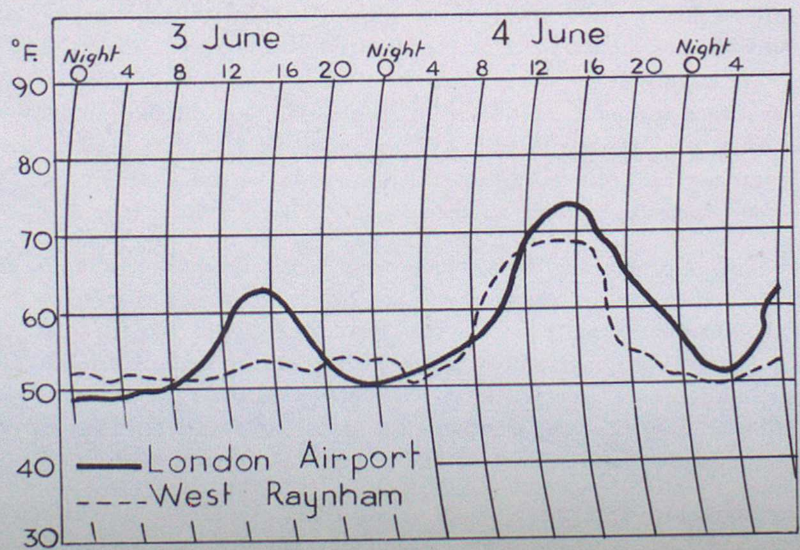

FIGURE 23 *Thermograms for London Airport and West Raynham, 3-4 June 1954*

West Raynham recorded little diurnal change of temperature. London Airport, situated further inland and downwind, had some breaks in the cloud (4.8 hours sunshine at Kew) and diurnal variations were restricted but noticeable. On the 4th, there was prolonged sunshine at both places and the enhanced maxima are clearly shown - particularly so at West Raynham.

14.12.6. Anticyclonic situations with and without a sheet of high stratocumulus cloud - 14 and 15 March 1952

This example has been chosen to show the effect of the presence and absence of a sheet of high stratocumulus on minimum and maximum temperatures.

Plate VIII shows the synoptic situation at 1200 G.M.T. on 14 March 1952 and the tracks of depressions and anticyclones for certain preceding and succeeding periods. During the period 13-15 March an anticyclone moved steadily south-south-eastwards down the North Sea turning later to a south-easterly track. At 1200 G.M.T. on the 13th it was centred some 100 miles north-north-east of the Shetlands and 48 hours later the centre was over western Austria. During this period its central pressure rose from 1025 millibars at 1200 G.M.T. on the 13th to a peak of 1031 millibars at 1200 G.M.T. on the 14th and declined to 1023 millibars at 1800 G.M.T. on the 15th. A fairly intense depression moved north-east from near the Azores on the morning of the 13th to the south-west approaches by the evening of the 15th. Low cloud and rain from this depression began to affect south-western districts on the 15th but apart from this the weather over the United Kingdom was dominated by the anticyclone.

During the 13th there was an extensive cover of low cloud with a main base between about 1,000 and 2,000 feet in the morning. By the evening the base had lifted to between 2,500 and 3,500 feet generally. Day maxima were mainly in the lower or mid forties.

This extensive high stratocumulus cloud sheet persisted throughout the night except in extreme eastern East Anglia where a complete clearance of cloud was first reported at 0400 G.M.T. on 14 March. This clearance was associated with drier air reaching England from the continent as the winds veered from a point north of east to east-south-east with the movement of the anticyclonic centre into Holland during the morning of the 14th. This clearance came too late in the night and was too localized to affect significantly night minima on the 13th-14th which were mainly in the upper thirties. The clearance of cloud spread steadily north-westward across England and by 0900 G.M.T. on the 14th had reached an approximate line from the Humber to Selsey Bill. By 1800 G.M.T. on the 14th the clearance of low cloud had extended over almost the whole of England and Wales apart from the extreme south-west.

Successive tephigrams from Hemsby and Larkhill corresponding to cloudy and cloudless conditions are shown in Figure 24. Apart from the drier air at levels below the stratocumulus inversion the remarkable dryness - particularly at Hemsby - of the air above 850 millibars is noteworthy. Application of Gold squares indicates day maxima in the mid forties on the 14th and it will be seen from Table 25 that this was a close estimate although there was some deviation due to topography, coastal effects and the time of clearance of cloud.

The night 14th-15th was virtually completely cloudless and surface winds were about the same as during the night 13th-14th although the pressure gradient was slightly greater. Radiational losses were much greater on the night 14th-15th and at many stations air temperatures fell below 32°F. At some stations

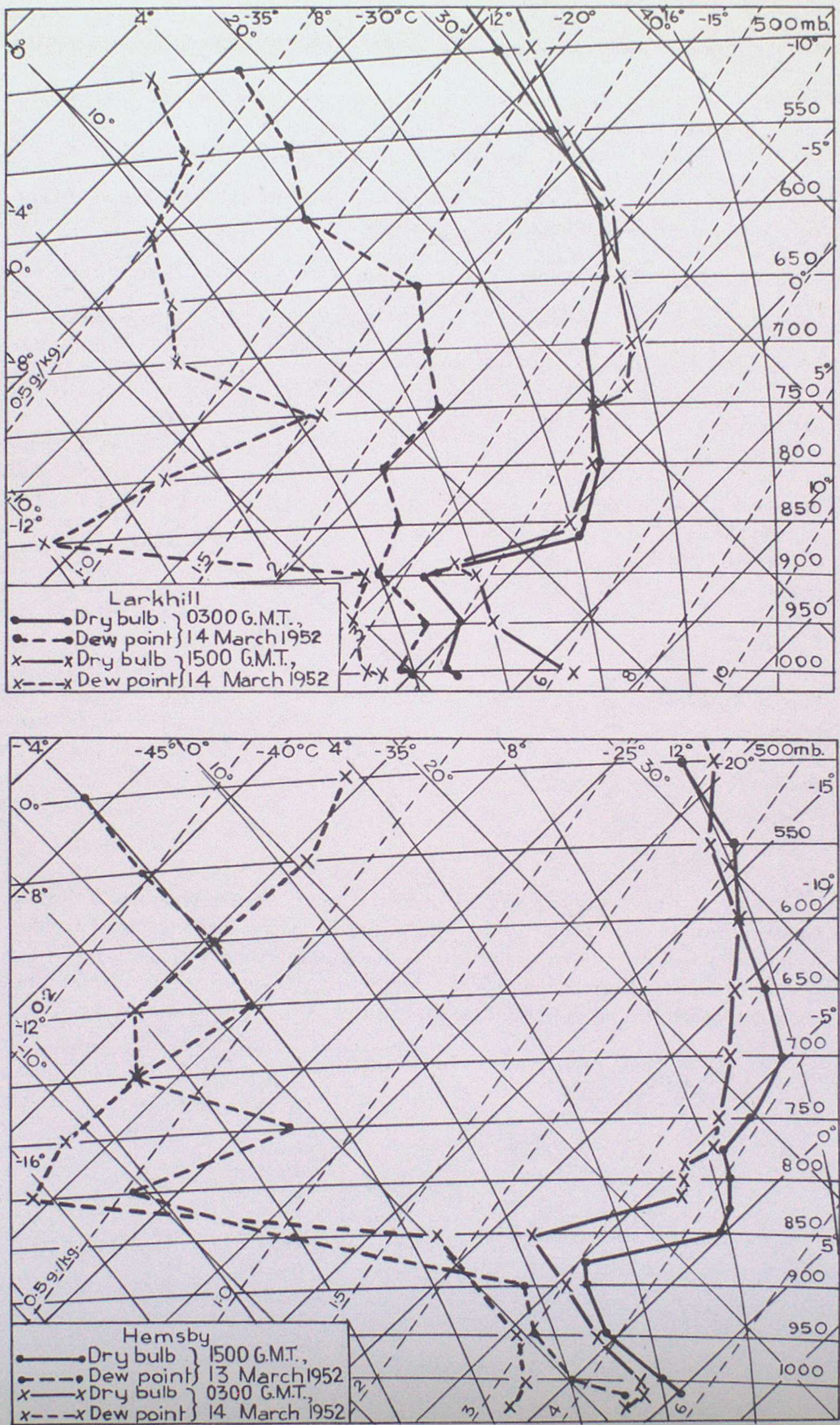

FIGURE 24 *Successive tephigrams from Hemsby and Larkhill corresponding to cloudy and cloudless conditions*

night minima on 14th-15th fell more than 10°F. below the minimum on 13th-14th. Ground frost was widespread.

On the 15th skies remained almost cloudless all day in England and Wales except the extreme south-west and, in spite of the lower night minima, day maxima on the 15th generally exceeded the maxima on the 14th.

Table 25 shows the maximum and minimum temperatures and duration of sunshine at a selection of stations during 14-15 March 1952.

TABLE 25 *Temperatures and duration of sunshine 14-15 March 1952*

	Maximum			*Minimum*		*Duration of sunshine*		
	13th	*14th*	*15th*	*13th-14th*	*14th-15th*	*13th*	*14th*	*15th*
	°F.	°F.	°F.	°F.	°F.	*hr.*	*hr.*	*hr.*
London Airport*	42	48	50	37	32	0.0	9.6	8.4
Gorleston	44	40	42	37	38	2.8	10.0	10.3
Elmdon	39	46	49	37	29	0.0	8.7	9.4
Plymouth	47	45	49	41	41	1.8	2.4	1.6
Pembroke Dock	44	45	48	41	39	0.0	3.9	4.0
Valley	49	48	54	40	30	6.1	5.6	9.9
Squires Gate	45	47	51	38	25	2.5	4.5	9.2
Leuchars	46	42	44	37	27	3.4	0.0	7.9
Renfrew	43	42	46	39	29	0.0	0.0	7.5
Wick	42	45	45	38	33	0.4	7.5	6.7
Watnall	42	45	49	36	27	0.0	8.6	9.5
Mildenhall	43	45	50	32	30	0.1	10.3	10.0
Boscombe Down	42	46	45	36	29	0.6	4.6	6.3
Dishforth	43	46	48	37	25	0.0	7.9	9.0
Tynemouth	45	43	44	37	30	6.1	5.0	6.9

* Sunshine at Kew

Figure 25 shows thermograms for London Airport, Squires Gate and Valley for 48 hours from 13 to 15 March 1952. The uniformity of the thermogram for London Airport on the 13th and part of the 14th is due to the presence of the layer of cloud. At Squires Gate and Valley the day-time performance of temperature was not greatly different on the 14th from that on the 13th, possibly due to the

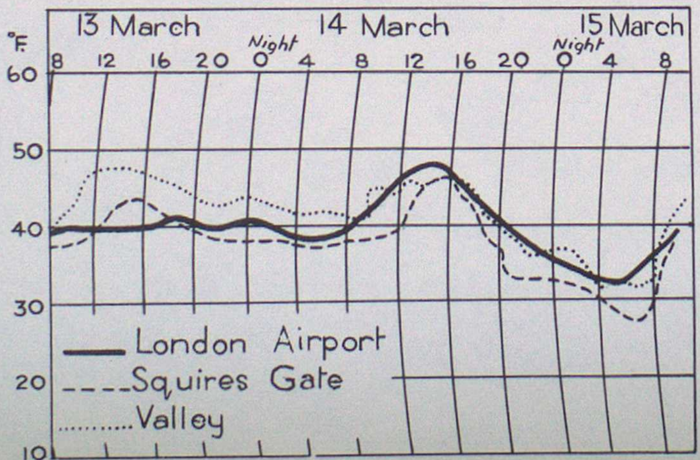

FIGURE 25 *Thermograms for London Airport, Squires Gate and Valley for 48 hours from 13-15 March 1952*

sheltering effect of the Pennines and Welsh Mountains. At all three locations the minima on 14th-15th were much lower than on 13th-14th, due to the absence of cloud on the night 14th-15th.

14.12.7. Examples of reversal of usual diurnal change of temperature

If temperatures are to fall throughout the day or to rise throughout the night it is clear that there must be fairly vigorous advection of air masses whose temperatures differ widely and such that the temperature difference is concentrated over a fairly narrow belt.

14.12.7.1. Fall of temperature during the day 31 January 1956. Plate IX shows the synoptic situation at 0001 G.M.T. on 31 January and the frontal positions at 0001 on 1 February. The depression in northern France had formed some 36 hours previously near the Azores as a wave on a cold front and had moved steadily east-north-east. A deep depression was situated in the Greenland - Iceland area maintaining a fairly vigorous cyclonic flow of air across the North Atlantic from the North American seaboard towards north-western France and the United Kingdom. Air to the north of the fronts WH and CH was maritime polar with a long sea track. An intense blocking high was situated over Scandinavia and occlusion OG was the main front separating very cold continental air from the maritime polar. Over the preceding four days three occlusions had moved north-east across the United Kingdom and had become stationary on the western side of the blocking high. During the day of 30 January occlusion OG had lain almost stationary along the east coast of England and Scotland but during the evening the front commenced to move west or west-south-west as the depression in northern France turned and moved south-east towards Italy and the Adriatic. By 1800 on the 31st the front OG had cleared the whole of the United Kingdom. Rain to the west of the occlusion turned to sleet or snow near the front and to the east of the front there was increasingly strong advection of increasingly colder air throughout the day. Table 26 shows the march of temperatures on 31 January 1956 at the four main synoptic hours at a selection of

TABLE 26 *Temperatures on 31 January 1956*

	Time (G.M.T.)			
	0001	*0600*	*1200*	*1800*
		°F.		
London Airport	44	41	31	27
Gorleston	36	32	28	26
Elmdon	40	32	28	25
Plymouth	42	42	41	33
Pembroke	41	42	38	31
Valley	41	40	35	30
Tynemouth	38	33	31	30
Leuchars	37	34	32	29
Renfrew	39	35	33	31
Wick	36	34	32	30

stations. It will be seen that during the day there was a pronounced drop in temperature which exceeded 10°F. in some places.

The thermogram for Elmdon from midday on 30 January to midday on 1 February 1956 is shown in Figure 26. There is a steady fall in temperature during the day on 31 January and the marked difference in the temperature régime of the air masses on either side of occlusion OG is clearly shown.

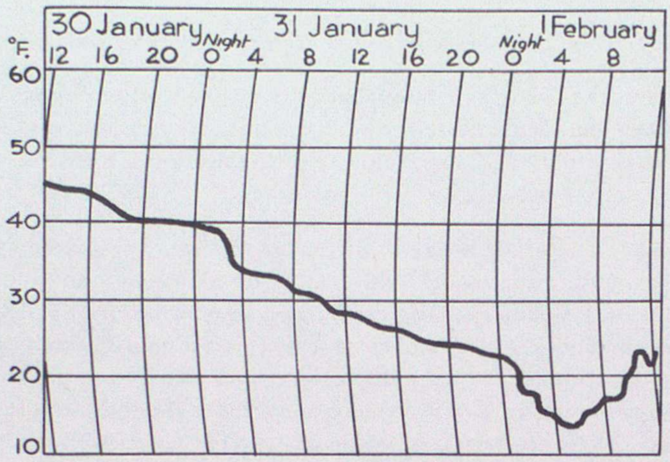

FIGURE 26 *Thermogram for Elmdon from midday on 30 January to midday on 1 February 1956*

14.12.7.2. Rise of temperature during the night - 4-5 February 1956. The cold spell which set in during 31 January 1956 persisted until the 4th. During this cold spell mild air had remained just off our western seaboard. On 4 February this mild air commenced to move eastward and by 0600 G.M.T. on the 5th covered the whole of the country. Plate X shows the synoptic situation at 0600 G.M.T. on the 4th and, in addition, frontal positions at 0600 on the 5th.

Table 27 shows the rise of temperatures from 0600 G.M.T. on the 4th to 0600 on the 5th.

TABLE 27 *Temperatures on 4-5 February 1956*

	4 February				*5 February*	
	0001	*0600*	*1200*	*1800*	*0001*	*0600*
				°F.		
London Airport	21	18	32	35	35	37
Gorleston	24	23	29	34	35	37
Elmdon	18	22	32	33	38	40
Plymouth	33	32	39	40	39	41
Pembroke	39	40	41	44	47	46
Valley	39	41	44	45	45	47
Tynemouth	27	29	33	37	38	47
Leuchars	30	32	34	34	34	34
Renfrew	37	37	36	36	46	48
Wick	37	36	38	35	35	39

It will be seen that almost all the stations showed an increase of temperature during the period of normal night cooling. The rise of temperature reached 10°F. or more at Tynemouth and Renfrew but the penetration of warm air to ground level at Leuchars was considerably delayed.

The rise of temperature during the night 4-5 February due to the continued advection of milder air is shown on the thermogram for Elmdon, reproduced in Figure 27.

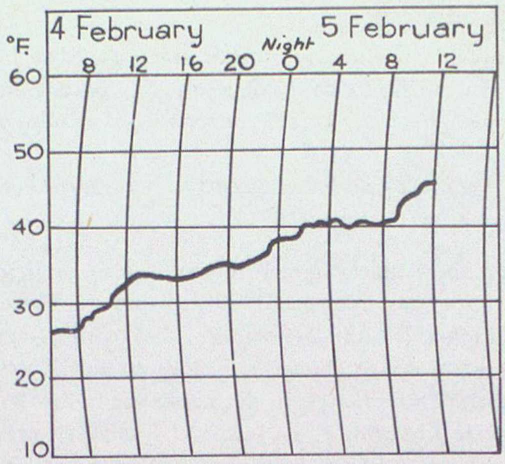

FIGURE 27 *Thermogram for Elmdon 4-5 February 1956*

14.12.8. Cold front passage followed by mobile anticyclone leading to extensive ground frost 10 May 1949

Plate XI illustrates the synoptic situation at 0600 G.M.T. on 9 May 1949. A frontal system had moved south-west and cleared almost the whole of the country. The major anticyclonic centre, HN, was situated to west of Ireland. To north and east of the frontal system a smaller anticyclonic centre, HI, had moved from a position between Iceland and Faroes at 0600 G.M.T. on the 8th to a point near Cape Wrath. During the 24 hours from 0600 G.M.T. on the 9th this small cell continued its south-south-easterly movement and by 0600 on the 10th was located in central England. To the north of this cell the warm front, WP, was bringing milder and cloudy air to Scotland. The cold air which followed the frontal system into England and Wales was quite shallow. The deepest layer of cold air did not exceed 7,000 feet at Downham Market.

The crux of the problem in forecasting night minimum temperatures on this occasion rested primarily on accurate timing and location of the anticyclonic cell, HI, during the night 9th-10th and the spread of warmer cloudy conditions on its northern flank. This determined those areas for which a full radiation night with light winds was likely to occur.

Plate XII shows the synoptic situation at 0600 G.M.T. on 10 May. It will be seen that the anticyclonic cell was located in central England and there were clear skies almost everywhere in England with light winds. Air frost was reported at several places and ground frost was extensive. Table 28 shows temperatures

TABLE 28 *Temperatures on 9-10 May 1949*

	Screen minimum		*Screen maximum*	*Grass minimum*	
	8th-9th	*9th-10th*	*9th*	*8th-9th*	*9th-10th*
			°F.		
London Airport	44	32	52	44	27
Gorleston	41	41	51	37	37
West Raynham	40	39	47	36	31
Plymouth	47	41	60	38	34
Aberporth	45	40	56	39	30
Valley	42	39	58	34	28
Squires Gate	41	32	54	39	25
Leuchars	...	38	53	35	32
Honiley	42	32	52	41	24
Little Rissington	42	37	52	42	25

recorded during 9-10 May 1949 at a selection of stations. The moderating effect of the sea can be seen in the figures for some stations on or near the coast but it is noteworthy that on the 9th-10th Exeter recorded a grass minimum of 22°F.

14.12.9. Stagnant anticyclonic conditions and the persistence and clearance of fog and/or low stratus - 2 and 3 March 1953

Plates XIII and XIV show that stagnant anticyclonic conditions covered most of England and Wales. Fog was dense and widespread and there was night frost in many places on the nights 1-2 and 2-3 March. The morning tephigrams for Larkhill on both 2 and 3 March are shown in Figure 28 and it will be seen that there was very little noticeable change in the upper air below 850 millibars. The differences in the isobaric situations in Plates XIII and XIV are small and it is very doubtful whether they could have been forecast with certainty in any detail. The persistence of fog in the west country on the 3rd was associated with the small anticyclonic cell which can be identified in Plate XIV. The clearance and persistence of fog was quite patchy and maximum temperatures correspondingly so. Tables 29 and 30 indicate the variability of temperatures on the 2nd and 3rd at a selection of stations. On the 4th some differences in maxima were even more marked than those tabulated for the 2nd and 3rd: for example, Dishforth 63°, London Airport 59°, Squires Gate 39°, Valley 40°, Aberporth 34°, Plymouth 44° and Chivenor 39°.

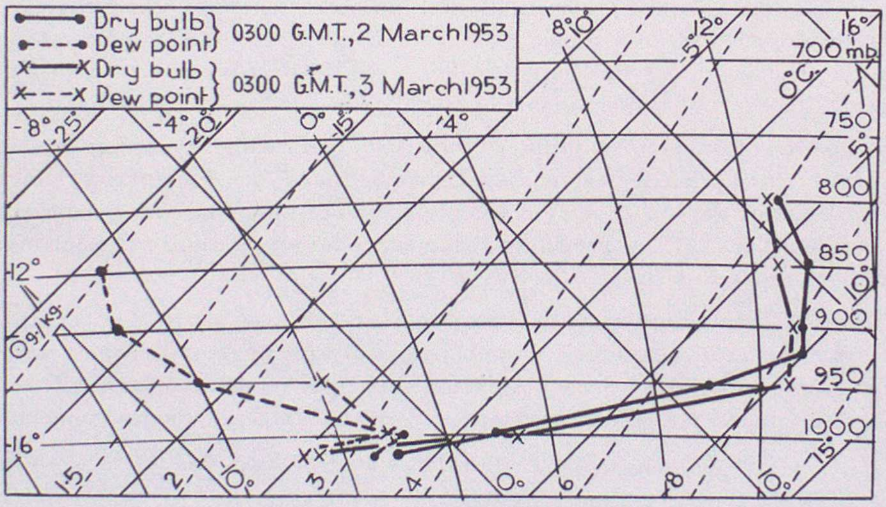

FIGURE 28 *Tephigrams for Larkhill, 0300 G.M.T., 2 and 3 March 1953*

Some very large diurnal ranges of temperature also were recorded at some places on 3 and 4 March. On 3 March the range was 33°F. (23° to 56°F.) at Boscombe Down and 30°F. (24° to 54°F.) at Dishforth. Boscombe Down recorded a similar range 34°F. (22° to 56°F.) on the 4th but Dishforth experienced the remarkable range of 43°F. (20° to 63°F.). The main factors causing these very large ranges were anticyclonic conditions with light winds and mainly clear skies, a very strong inversion from the ground up to 600-800 metres (2,000-2,500 feet approximately) with very low dew-points and a dry ground. The thermograph trace for Dishforth for 2, 3 and 4 March 1953 is reproduced in Figure 29 and

TABLE 29 *Temperatures and remarks on the weather on 2 March 1953*

Station	*Screen min. 1st-2nd*	*Screen max. 2nd*	*Duration of sunshine 2nd*	*Remarks*
	°F.	°F.	*hr.*	
London Airport	33	43	(Nil at Kew)	Fog all day
Tangmere	28	51	9.7	No fog
Boscombe Down	25	44	2.5	Fog cleared 1200-1500
West Raynham	28	35	0.0	Fog cleared to mist temporarily around 1500
Mildenhall	30	39	1.1	Fog lifted to 8 oktas low stratus by 1200 which cleared for a time in late afternoon
Elmdon	27	45	4.1	Fog cleared by 1200
Watnall	28	42	3.6	Fog cleared for a time in afternoon
Ross-on-Wye	29	38	2.4	Fog cleared by midday
Shawbury	29	46	4.3	Fog cleared by midday
Plymouth	33	56	8.3	Fog cleared by midday
Chivenor	28	60	9.3	Fog cleared very early
Pembroke Dock	31	51	6.0	No fog
Squires Gate	29	38	0.0	Fog lifted to low stratus
Finningley	29	46	3.5	Fog cleared by midday
Dishforth	28	42	2.6	Fog all day but thin in late afternoon
Tynemouth	29	52	7.1	Fog cleared by midday

TABLE 30 *Temperatures and remarks on the weather on 3 March 1953*

Station	*Screen min. 2nd-3rd*	*Screen max. 3rd*	*Duration of sunshine 3rd*	*Remarks*
	°F.	°F.	*hr.*	
London Airport	26	51	(5.7 at Kew)	Fog cleared after midday
Tangmere	21	46	3.8	Fog all day
Boscombe Down	23	56	8.3	Fog cleared by midday
West Raynham	25	37	0.0	Fog all day
Mildenhall	26	33	0.3	Fog all day
Elmdon	26	45	2.6	Fog cleared for a time in afternoon
Watnall	24	49	4.1	Fog all day but thin after midday
Ross-on-Wye	27	35	0.0	Fog all day
Shawbury	29	38	0.0	Fog all day except for a time in late afternoon
Plymouth	28	49	8.3	Fog formed after midday
Chivenor	29	50	9.3	Fog formed after 1500
Pembroke Dock	26	40	0.0	Fog all day
Squires Gate	35	39	0.0	Misty but overcast all day
Finningley	20	52	5.3	Fog cleared by midday
Dishforth	24	54	6.5	Fog cleared by midday
Tynemouth	33	50	3.4	Fog cleared by midday

vividly illustrates the ranges and the variation in ranges at that station over the three-day period.

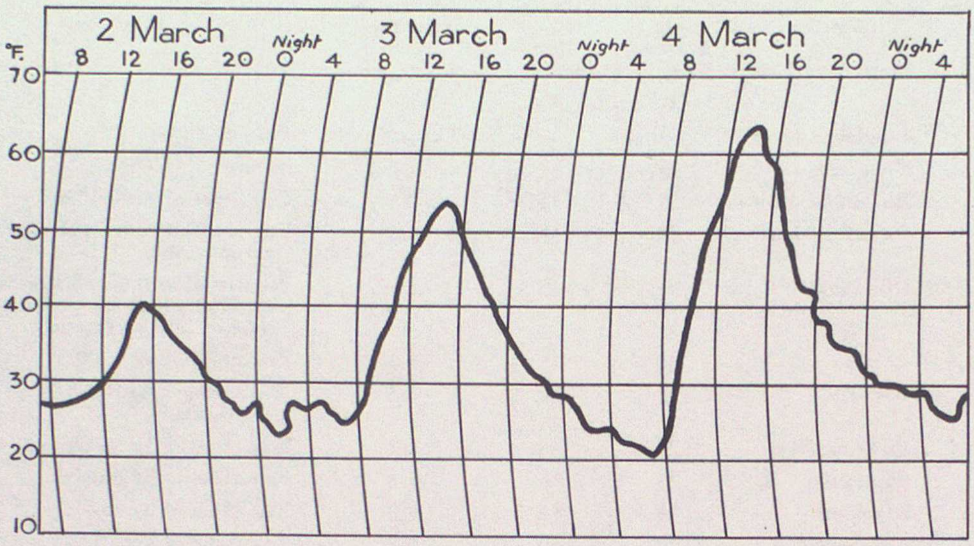

FIGURE 29 *Thermogram for Dishforth, 2, 3 and 4 March 1953*

APPENDIX I

KEY TO PLOTTING ON SYNOPTIC CHARTS

(i) The method of plotting is that in use on 1 January 1958.

(ii) Screen and dew-point temperatures are plotted in degrees Fahrenheit.

(iii) Horizontal visibilities are plotted in code figures according to International Code 84 of which an extract is given below.

(iv) Total cloud amounts are indicated conventionally by hatching in the station circle as detailed on p. 84. On some charts individual cloud amounts are plotted in oktas and cloud heights (above station level) are plotted in code figures according to International Code 40 except for decade 90-99 for which only the last of the code figures is plotted. An extract from Code 40 is given on p. 84. Cloud types are plotted in conventional symbols.

(v) Surface wind directions are plotted in the form of an arrow flying with the wind. Surface wind speeds are represented by feathers, each full feather represents 10 knots and a half feather 5 knots. An arrow with no feather indicates a wind of 1-2 knots. A calm is indicated by a circle concentric with the station circle.

(vi) Present and past weather are plotted conventionally.

Extract from Code 84: horizontal visibility

Code figure	*Distance*	
	km.	*yd. (approx.)*
00	<0.1	<110
01	0.1	110
02	0.2	220
..	..	..
..	..	..
..	..	..
09	0.9	990
10	1.0	1,100
11	1.1	1,200
12	1.2	1,300
..	..	..
..	..	..
..	..	..
19	1.9	2,000
20	2.0	2,200
21	2.1	2,300
22	2.2	2,400
..	..	..
..	..	..
..	..	..
29	2.9	3,100
30	3.0	3,300
31	3.1	3,400
32	3.2	3,500
..	..	..
..	..	..
39	3.9	4,200

Code figure	*Distance*	
	km.	*miles (approx.)*
40	4.0	2½
41	4.1	2½
42	4.2	2⅝
43	4.3	2¾
..	..	..
46	4.6	2⅞
..	..	..
48	4.8	3
..	..	..
50	5	3⅛
51-55 are not used		
56	6	3¾
57	7	-
58	8	5
..	..	..
60	10	6¼
..	..	..
66	16	10
..	..	..
70	20	12½
..	..	..
79	29	-
80	30	18¾
81	35	-
82	40	25
..	..	..
88	70	43¾
89	>70	>43¾

Extract from Code 40: height of cloud

Code figure	*Height ft.*
00	<100
01	100
02	200
03	300
..	..
10	1,000
11	1,100
..	..
20	2,000
21	2,100
..	..
30	3,000
31	3,100
..	..
40	4,000
41	4,100
..	..
49	4,900
50	5,000
(51-55 not used)	
56	6,000
57	7,000
58	8,000
59	9,000

Code figure	*Height ft.*
60	10,000
61	11,000
..	..
..	..
70	20,000
71	21,000
..	..
..	..
80	30,000
81	35,000
82	40,000
..	..
..	..
88	70,000
89	>70,000
90	150
91	150 - 300
92	300 - 600
93	600 - 1,000
94	1,000 - 2,000
95	2,000 - 3,000
96	3,000 - 5,000
97	5,000 - 6,500
98	6,500 - 8,000
99	≥8,000 or no low clouds

Plotting symbols for total cloud

Nil trace	1/8	2/8	3/8	4/8	5/8	6/8	7/8	8/8	Sky obscured

APPENDIX II

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