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WIND

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CHAPTER 16

WIND

16.1 INTRODUCTION

Wind is air in motion and, as such, it plays a vital part in atmospheric processes. Air motion provides a means of transporting heat and moisture from one region to another, and much of the weather can be related to the action and interaction of the winds. Wind analysis and prognosis is therefore of great importance to the meteorologist.

Information on winds is also important to many users of meteorological forecasts. Throughout the history of weather forecasting there has been a close association between meteorologists and mariners, and information on surface winds is supplied for a wide variety of activities from amateur yachting to the weather routeing of ships. The strength and direction of surface winds over certain sea areas are factors to be taken into account when considering whether there are likely to be abnormally high tides which may constitute a threat to sea defences and lead to extensive flooding. The introduction of offshore drilling for oil and gas has led to an increasing requirement for forecasts of wind speeds, with the associated wave and swell conditions, for the operators of oil rigs of various types. Particularly crucial is the requirement for light winds when the rig is being towed from one location to another, and the success of the operation is dependent upon the accuracy of the forecasts. Wind information is also essential for the safe and efficient conduct of air operations. Forecasts of surface winds are required for the take-off and landing of aircraft; for short-term forecasts it is often better to give the latest observation than to try to predict any changes that may occur, and this raises the problem of how to ensure that the wind measurements provide a reasonable estimate of the wind likely to be experienced by the aircraft. Not only are surface winds important, but the existence of strong wind shear in the lowest few hundred feet may, on occasions, constitute a hazard during take-off and approach. For transit flights at altitude and for the adequate separation of aircraft, particularly in congested air spaces, upper winds are provided; with the advent of supersonic aircraft winds are required at heights up to 20 kilometres or so.

Forecasts of strong winds are required for the construction of high buildings and the operation of tall cranes. Winds are also important to farmers, particularly when spraying crops, and to the non-specialist user of meteorological information. For example, a wind from sea to land may provide a welcome lowering of air temperatures near the coast on days with high temperatures inland but, when the sea is cold, may produce cold, damp days in winter and spring. Strong winds with a cool airstream feel colder than a consideration of the air temperature alone would imply. It is appropriate, therefore, that some indication of these effects should be included in forecasts for the general public.

Air motion occurs on a wide range of scales, and it is useful for the forecaster engaged on day-to-day forecasting to be aware of the scales of the wind systems with which he is concerned and their interrelationships. Leaving aside the possibility of long-term climatic change, the wind systems of the general circulation have the largest scale in time and space, and some understanding of the general circulation is a very useful background for the forecaster even though it can seldom be applied in day-to-day forecasting (see Chapter 1 - Introduction). The next lower scale of wind systems which should be recognized by the forecaster is that of the long-wave patterns which can usually be found at the 700-millibar level and above over substantial portions of the earth. The wavelength of these systems is of the order of several thousands of kilometres; these patterns are often relatively slow moving and, with their associated surface features, may determine the general character of the weather over an area for several days. The practical forecaster must recognize these systems and know how they are likely to behave (see Chapter 2 - Dynamical ideas in weather forecasting). Embedded in these major patterns are features of smaller scale, the so-called 'short waves', the smallest of which are associated with, for example, waves on cold fronts, minor troughs, etc. and often move through the large-scale flow without changing it a great deal. Other disturbances may grow sufficiently to modify or even change completely the long-wave pattern. Yet smaller are the mesoscale features, such as squall-lines and the small highs and lows associated with thunderstorms. Some local winds, e.g. sea-breezes and land-breezes, mountain and valley winds, are of a similar scale but occur generally when the larger-scale airflow is comparatively weak. The long waves are readily delineated by the synoptic networks over most of the northern hemisphere, but the short-wave features, although 'picked up' by the synoptic network over land, may well slip through the more open network at sea. The mesoscale features and local winds can at times be detected by careful analysis over an area where observations are closely spaced.

There are disturbances of a still smaller scale in the wind field, ranging in size from the air motions associated with individual clouds down almost to the molecular scale. These disturbances, or eddies, may be caused by mechanical friction with the ground, by heating of the ground surface giving rise to convection and by the break-down of strong vertical wind shear associated with stable stratification in the free atmosphere (clear-air turbulence, which, together with turbulence to the lee of high ground, will be discussed in Chapter 23 - Bumpiness in aircraft).

In most of the larger-scale wind disturbances the horizontal component of motion is far greater than the vertical component but the vertical component plays a vital role in the determination of the weather. In the smaller-scale disturbances the ratio of the vertical component to the horizontal component of motion may be greater than in the larger scale, and in the smallest scale of turbulence the two components are of the same order of magnitude. In general, however, any reference to 'wind' is usually taken to mean the horizontal component, and this will be so in the rest of this chapter unless the vertical component is explicitly mentioned.

For the purposes of this chapter, the atmosphere can be divided into three layers, viz:

(a) the surface layer;

(b) the planetary boundary layer ('spiral' or 'Ekman' layer);

(c) the free atmosphere.

The wind in the free atmosphere is determined largely by the pressure pattern and its changes, and the rotation of the earth, while friction can usually be neglected. Near the earth's surface, however, atmospheric momentum is lost to the surface by friction when the moving air encounters obstacles (the layer in which this happens has been termed the 'interfacial layer', see Sheppard[1]). In order to compensate for this loss, momentum is carried downwards from the free atmosphere by eddies; the depth and properties of the layer in which this downward eddy transport of momentum occurs, the planetary boundary layer, vary considerably with the vertical temperature structure of the atmosphere, since both the intensity of the turbulence and the vertical range of action of the eddies depend very much on the vertical stability of the lower layers of the atmosphere. On fairly clear days over land, heating of the surface results in an unstable lapse rate over a certain depth, often about $\frac{1}{2}$ to 2 kilometres, but on days of large cumulus there may be some difficulty about the definition of the boundary layer. In neutral conditions, turbulence is caused mainly by mechanical friction between the surface and the air, and the depth of the boundary layer varies with the roughness of the surface and with the wind speed, normally lying within the range of a few hundred metres to one kilometre or so. Stable stratification, on the other hand, considerably inhibits turbulence except near rough surfaces, and in light winds the boundary layer can be very shallow.

The lowest part of the planetary boundary layer has been called the 'surface layer'. Within this region, which extends to a few tens of metres above the ground, the motion is determined largely by the turbulent eddy structure, that is to say the movement of a given 'parcel' of air is mainly dependent upon the forces exerted on it by the turbulent motion of the surrounding medium; Coriolis and pressure-gradient forces play a relatively small part. This region has been extensively studied both theoretically and experimentally, but, although it plays an extremely important role in the vertical exchange of heat and water vapour in addition to momentum, the practical forecaster is not usually concerned greatly with the details of the flow in this limited region. Increasingly, however, he has to consider flow in the higher parts of the planetary boundary layer. Not only is this region very much less accessible for measurement than the surface layer, but considerable theoretical problems are encountered because the pressure-gradient and Coriolis forces assume an importance comparable with the eddy forces. The main requirements are for forecasts of strong winds during the construction of tall structures, usually over the very complex and rough 'surface' of an urban area. Forecasts of marked wind shear are also needed for take-off and landing of aircraft at locations where the topography favours the occurrence of such shears in certain conditions.

In the free atmosphere, the turbulent eddy stresses can usually be neglected in comparison with the pressure-gradient and Coriolis forces, and the wind can usually be estimated with reasonable accuracy from surface-pressure or isobaric-contour charts. Simplified mathematical relations are of much practical value and are discussed in the next section.

16.2 MATHEMATICAL EXPRESSIONS FOR WIND

The theoretical derivation of the general equations of air motion is included in many textbooks and a treatment from first principles is not repeated in this handbook.

The full equation of motion in vector form is:

$$\frac{d\mathbf{V}}{dt} = -\frac{1}{\rho}\nabla p + \mathbf{g} + \mathbf{F} - 2\mathbf{\Omega}\times\mathbf{V} \qquad \ldots\ldots \quad 16.1$$

where

$\mathbf{V}$ is the wind velocity.

$d\mathbf{V}/dt$ is the rate of change of wind velocity following the motion.

ρ is the air density.

∇p is the pressure gradient.

$\mathbf{g}$ is the gravitational acceleration.

$\mathbf{F}$ represents external forces, mainly friction.

$\mathbf{\Omega}$ is the angular velocity of rotation of the earth: if u, v and w are the components of motion along the x-, y- and z-axes, and ϕ is the latitude, then

$\mathbf{\Omega} = (w\Omega\cos\phi - v\Omega\sin\phi,\ u\Omega\sin\phi - u\Omega\cos\phi)$.

The equation is merely a statement of Newton's second law of motion, i.e. that the acceleration of unit mass of air is equal to the vector sum of the forces acting on it.

Although all the terms of equation 16.1 can be taken into account in numerical models of the atmosphere (see Chapter 3 - Background to computer models) the flow may quite often be represented sufficiently accurately for many purposes by much simpler expressions, in particular by the geostrophic and gradient-wind approximations which will be discussed below. There are regions, however, where the real winds are markedly different from the geostrophic or gradient winds, for example at the entrances and exits of jet streams, when vertical velocities are large, and in mesoscale systems which may last for only a few hours. Differences exist also where pressure patterns are changing rapidly, giving rise to an 'isallobaric' component of the wind, which will be briefly described in 16.2.3 (page 7).

16.2.1 The geostrophic wind

When the air is moving horizontally and experiencing no acceleration, and if the frictional forces are zero, the motion represents a balance between the pressure-gradient force and the Coriolis force, and is said to be geostrophic. Putting $d\mathbf{V}/dt$ and $\mathbf{F}$ in equation 16.1 equal to zero, and forming the cross product with $\mathbf{k}$ to obtain the horizontal components, we have

$$0 = -\frac{1}{\rho}\mathbf{k} \times \nabla p - f\mathbf{V}_g \qquad \ldots . \; 16.2$$

where $\mathbf{V}_g$ is the geostrophic wind and f is the Coriolis parameter, $2\Omega \sin\phi$.

Equation 16.2 shows that the geostrophic wind blows along the isobars, with low pressure to the left of the direction of motion in the northern hemisphere, and with a speed given by

$$V_g = \frac{1}{f\rho}\frac{\partial p}{\partial n} \qquad \ldots . \; 16.3$$

where $\partial p/\partial n$ is the pressure gradient along the normal to the isobars.

On isobaric contour charts, the vertical co-ordinate is p, and the geostrophic wind speed is given by

$$V_g = \frac{g}{f}\frac{\partial h}{\partial n} \qquad \ldots . \; 16.3a$$

where h is the height of the isobaric surface above mean sea level. The wind direction is along the contours of h.

This expression is simply derived by applying the hydrostatic equation to equation 16.3 (see Chapter 2 - Dynamical ideas in weather forecasting).

16.2.2 The gradient wind

If the air is following a curved path with a speed V, there is an acceleration towards the centre of curvature, of magnitude V^2/r, where r is the radius of curvature of the path.

The gradient wind, V_{gr}, is that steady wind for which the accelerations resulting from the pressure gradient, the Coriolis force and the curvature of the path exactly balance. In cyclonic flow, the centripetal acceleration acts in the same sense as the acceleration produced by the pressure gradient;

it follows that the Coriolis force must be smaller than for straight flow, in order to provide a net inward force for the centripetal acceleration. Therefore

$$f V_{gr} = \frac{1}{\rho}\frac{\partial p}{\partial n} - \frac{V_{gr}^2}{r}$$

$$= f V_g - \frac{V_{gr}^2}{r} \qquad \ldots . \; 16.4$$

For anticyclonic flow, the Coriolis force must be greater than the pressure gradient force, and V_{gr} is given by

$$f V_{gr} = f V_g + \frac{V_{gr}^2}{r} \qquad \ldots . \; 16.4a$$

The implications of these equations, quadratics in V_{gr}, are discussed fully in many textbooks on dynamical meteorology, e.g. Reference 41, and also in Chapter 2 of this handbook.

16.2.3 <u>The isallobaric wind</u>

The isallobaric wind component arises from changes of the pressure field with time. More strictly, the isallobaric component at a point is related to the instantaneous rate of change of pressure gradient at that point, or, by changing the order of differentiation, the gradient of the rate of pressure change:

$$V_i = \frac{1}{f^2\rho}\frac{\partial}{\partial t}\left(\frac{\partial p}{\partial n}\right) = \frac{1}{f^2\rho}\frac{\partial}{\partial n}\left(\frac{\partial p}{\partial t}\right) \qquad \ldots . \; 16.5$$

where $\partial/\partial n$ represents differentiation along the normal to the isallobars.

From equation 16.5 it can readily be seen that the magnitude of the isallobaric wind is obtained by treating the isallobars as isobars, and dividing the speed so measured by f. In practice, isallobars are usually drawn from reported pressure changes over three hours: the divisor should then be $\frac{1}{3}f$, where f is in units of h^{-1}.

The value of f varies from about 0.40 h^{-1} at latitude 50 degrees to about 0.45 h^{-1} at latitude 60 degrees; $\frac{1}{3}f$ can be taken as approximately 0.8 for the southern half of the British Isles and 0.75 for the northern half. The isallobaric wind component blows from (algebraically) high to low tendencies.

16.2.4 Thermal wind

It is sometimes desirable to estimate the thermal wind, that is to say the vector difference between the winds at two different levels. If h' is the thickness between these pressure levels it can be shown that the components u_T and v_T of the thermal wind are given by

$$u_T = -\frac{g}{f}\frac{\partial h'}{\partial y} \quad \text{and} \quad v_T = \frac{g}{f}\frac{\partial h'}{\partial x} \qquad \ldots . \ 16.6$$

These equations indicate that the thermal wind blows along the thickness pattern with smaller thickness to the left in the northern hemisphere. Furthermore, the thermal wind bears the same relation to the thickness pattern as the geostrophic wind to the contour lines. Thus if the same standard interval is used (for example, 60 metres) a geostrophic contour scale may be used to determine thermal winds from a thickness chart (see 16.3 below).

16.3 EVALUATION OF WINDS FROM METEOROLOGICAL WORKING CHARTS

16.3.1 Geostrophic wind

The basic tool for estimating the geostrophic wind from a working chart is the geostrophic wind scale. On many charts the appropriate wind scale is printed as an inset, and two examples of such scales are shown in Figures 1 and 2. The curves show the distance between contours or isobars at the specified intervals for various values of the geostrophic wind speed for a range of latitudes.

There are two main methods of use:

(a) A copy of the scale engraved on a transparent base is placed on the chart, the left-hand edge parallel to the flow along an isobar or contour. The wind speed, interpolated between the curves if necessary, is then read off at the point at which the next isopleth crosses the appropriate latitude line.

Alternatively, the wind speed may be read off at the nth isopleth, and the result multiplied by n to give an estimate of the mean speed over a greater distance.

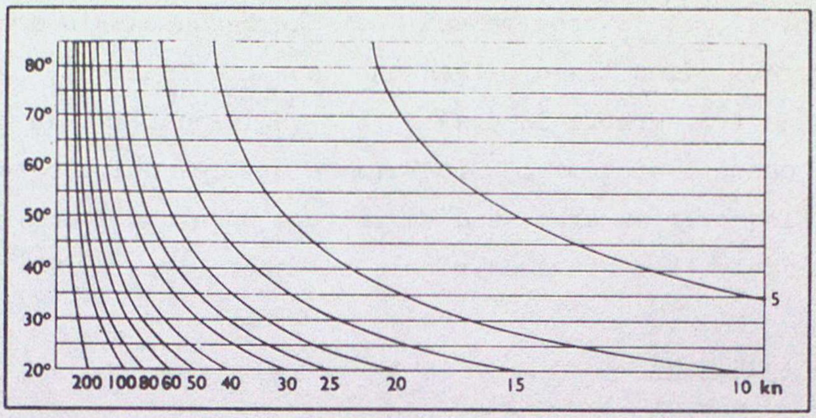


FIGURE 1. Geostrophic wind scale in knots for isobars at intervals of 4 mb, correct for 1013.250 millibars and 15°C (288.15 K)

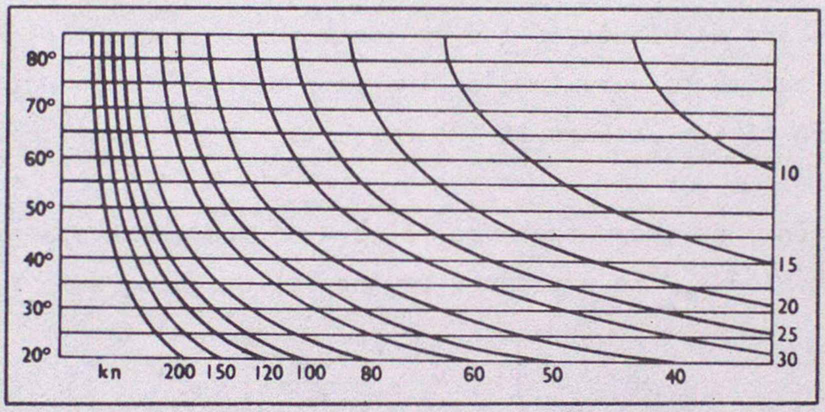


FIGURE 2. Geostrophic wind scale in knots for contours at intervals of 120 geopotential metres

(Polar stereographic projection, one standard parallel at 60° where natural scale is 1: 20 000 000)

A geostrophic wind scale for a given chart may be used on a chart of a different scale (but of the same projection); for a chart with a scale of half that of the original chart, for example, the wind speeds read off must be divided by two.

(b) A pair of dividers is set to the perpendicular distance between two contours and transferred to the appropriate part of the geostrophic scale. The wind speed is read off as in (a).

Alternatively, the dividers may be set to the distance, on the appropriate part of the scale, corresponding to a given wind speed. The dividers are transferred to the chart, along a line normal to the flow, and the number of isopleth intervals between the divider points is read off. The initial speed is then multiplied by the number of intervals to give the mean wind speed across the line.

In general, it is recommended that the measurements are made over a distance corresponding to more than one isopleth interval, partly to reduce errors arising from errors in drawing the individual isopleths, and partly because the concept of geostrophic balance applies only to uniform, straight and steady flow over an area, and variations on a small scale within such an area are unlikely to be in geostrophic balance.

16.3.2 Gradient wind

To calculate the gradient wind it is necessary to take account of the curvature of the flow. This complicates the problem for, in many cases, it is not easy to determine the radius of curvature of the path of the air and so determine the magnitude of the cyclostrophic component of the gradient wind. (The radius of curvature of the path of the air is not the same as the curvature of the isobars if the isobaric system is moving.)

Petterssen[2] has shown that the ratio M of the geostrophic wind (V_g) to the gradient wind (V_{gr}) is expressed by the following equation:

$$M\left[1 - M + \frac{C\cos\Psi}{fr_i}\right] = \frac{V_g}{fr_i} \qquad \ldots\ldots 16.7$$

where r_i is the radius of curvature of the isobar at the point where the geostrophic wind (V_g) is measured, C is the speed of movement of the pressure system, Ψ is the angle between the direction of movement of the pressure system and the geostrophic wind (V_g) (so that $C\cos\Psi$ is the component of movement of the pressure system in the direction of the geostrophic wind and in the same sense), and f is $2\Omega\sin\phi$ where ϕ is the latitude.

The calculation of gradient wind using a slightly modified form of equation 16.7 has been discussed by Gilbert,[3] who produced tables from which the cyclostrophic correction could readily be derived. More recently, the Meteorological Office has prepared a set of Gradient Wind Tables, from which Table 16.1 has been extracted, giving the cyclostrophic correction for various latitudes, radii of curvature, and geostrophic speeds for a stationary system; the values are given to the nearest 5 knots.

The measurement of the radius of curvature presents some difficulties, not least of which is allowing for the distortion resulting from the map projection. This has been discussed by Freeman[4] for two types of projection; the corrections that should be applied for map distortion can at

TABLE 16.1 Correction to be applied to geostrophic wind speed to obtain gradient wind speed when the system of isobars or contours is stationary

Latitude	Radius of curvature of isobar or contour									
	nautical miles									
70°	160	330	490	650	820	1220	1630	2040	2450	3260
60°	180	350	530	710	880	1330	1770	2210	2650	3540
50°	200	400	600	800	1000	1500	2000	2500	3000	4000
40°	240	480	720	960	1190	1790	2380	2980	3580	4760
30°	310	610	920	1230	1530	2300	3060	3830	4600	6130
Geostrophic speed	**Cyclonic curvature – correction to be subtracted**									
knots	***knots***									
20	5	0	0	0	0	0	0	0	0	0
40	10	5	5	5	5	0	0	0	0	0
60	20	15	10	10	5	5	5	5	5	0
80	30	20	15	15	10	10	5	5	5	5
100	40	30	25	20	15	15	10	10	5	5
120	55	40	30	25	25	15	15	10	10	10
140	65	50	40	35	30	25	20	20	15	10
160	80	60	50	45	35	30	25	20	15	15
180	95	70	60	50	45	35	30	25	20	15
200	105	85	70	60	55	40	35	30	25	20
Geostrophic speed	**Anticyclonic curvature – correction to be added**									
knots	***knots***									
20	20	5	0	0	0	0	0	0	0	0
40	–	35	10	5	5	5	0	0	0	0
60	–	–	55	20	15	10	5	5	5	0
80	–	–	–	70	30	15	10	10	5	5
100	–	–	–	–	90	25	15	15	10	5
120	–	–	–	–	–	45	25	20	15	10
140	–	–	–	–	–	80	40	30	20	15
160	–	–	–	–	–	–	60	40	30	20
180	–	–	–	–	–	–	95	55	40	25
200	–	–	–	–	–	–	180	75	55	35

Notes

(a) The theoretical maximum gradient wind for anticyclonic curvature is equal to twice the geostrophic wind.

(b) If the system of contours or isobars is moving, the gradient-wind correction is related to the appropriate value in Table 16.1, as in Figure 3.

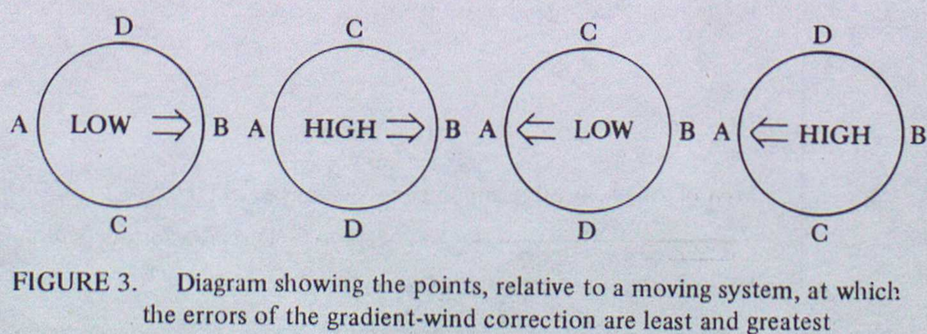


FIGURE 3. Diagram showing the points, relative to a moving system, at which the errors of the gradient-wind correction are least and greatest

At points A and B the table gives a correct determination.

At points C the table gives an overestimate of the correction, particularly if the speed of the system is similar to the geostrophic speed. (If these two speeds are equal, the correction is zero.)

At points D the table gives an underestimate of the correction.

times be very large. For further details the reader should consult Freeman's paper.

Zobel[5] and Boyden[6] have compared the errors of geostrophic and gradient winds. At 200 millibars, Zobel found that the gradient wind computed from countour curvature, even without adjustment for movement of the system, gave a better estimate of the actual wind than did the geostrophic wind. In contrast, Boyden's results suggested that the geostrophic wind gave a better estimate of the actual wind just above the friction layer than did the gradient wind computed from isobaric curvature, except when the radius of curvature was small in cyclonic flow. He concluded that when the radius is 500 n. mile (900 km), one-third of the cyclostrophic correction should be applied. For a radius of curvature of 250 n. mile (450 km) the correction should be increased to one-half of the cyclostrophic correction.

16.3.3 Isallobaric wind

A measure of the isallobaric wind may be obtained in the following way. Construct a set of isallobars (for three-hourly pressure change) and, using a geostrophic scale, measure a fictitious geostrophic wind. Isallobars are best drawn at intervals of one millibar pressure change in three hours and these must then be regarded as isobars at intervals of one millibar. Appropriate corrections are therefore necessary when geostrophic scales for isobaric spacings of 2 or 4 millibars are used. Then in latitudes near the British Isles the isallobaric wind is approximately 0.8 times the fictitious geostrophic wind so determined. It blows at right angles to the isallobars towards the (algebraically) lower pressure tendency as shown schematically in Figure 4 (see also 16.2.3, page 7).

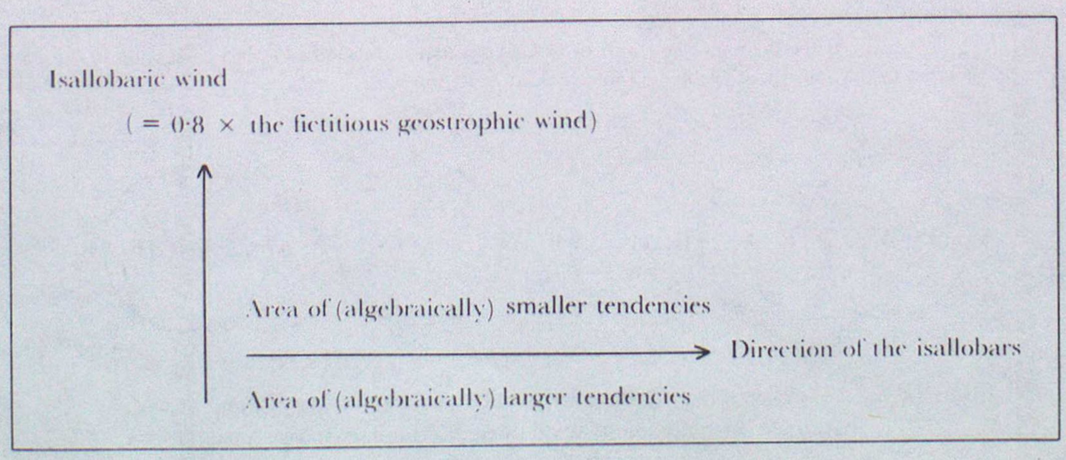


FIGURE 4. Isallobaric wind

16.4 ERRORS AND REPRESENTATIVENESS OF WIND OBSERVATIONS

16.4.1 Surface winds

Measurements of surface wind made with anemometers which are reasonably well exposed, correctly aligned and functioning satisfactorily may be regarded as sufficiently accurate for almost all needs of practical analysis and prognosis. For special requirements and investigations, however, the errors and characteristics in regard to lag, sensitivity, etc. of ordinary anemometers may be unacceptable. To meet these requirements special instruments and techniques may be required. For a more detailed consideration of the errors in surface wind measurements and the instrumental characteristics of anemometers, reference may be made to the 'Handbook of meteorological instruments',[7] and to an article by Maidens.[8]

For synoptic purposes, the 'observed wind' is a mean value over a period of 10 minutes, but for other purposes periods of different lengths may be more appropriate. Often the wind needs to be known at a location different from that at which it is observed, and this raises the question of how representative of the wind over a given area is the wind at a given spot. In particular, in the provision of winds for aircraft landing and take-off uncertainties arise because of the distance between the anemometer and the runway and the inevitable delay between the wind observation and the event for which the information is required. The nature of the fluctuations of the wind indicates that a mean wind, measured over a period of a few minutes, is more likely to be representative of the wind near the runway several minutes later than is a reading averaged over only a few seconds. The International Civil Aviation Organization[9] has recommended that a mean wind over a 2-minute interval should be used as an interim measure until more definite and reliable information is available. Sparks and Keddie,[10] using data for one period of strong westerly winds in a warm sector, found a 4-minute or 5-minute average to be better, but it would probably be unwise to extend these results to other synoptic situations. Hardy[11] has studied data from a wider variety of situations, including thunderstorms and squalls, taking into consideration

(a) the root-mean-square of all the differences, and
(b) differences greater than 10 knots only.

The latter study was carried out, and the criterion chosen, on the basis of a statement by Burnham[12] that changes of wind speed of as little as 10 knots, affecting an aircraft for one or two seconds, may in some circumstances be critical on approach. Hardy found that the lag between the observation and its use was far more important than the averaging period, an average over 2 minutes not generally being inferior to an average over a longer period for a lag of 4 minutes or less. For greater lags, up to 10 minutes, the data suggest that the averaging period need not be much longer, and that the value suggested by Sparks and Keddie, 4-5 minutes, is probably longer than necessary in most synoptic situations.

16.4.2 Upper winds

The errors of radar-wind measurements have been discussed in Chapter 11 - Upper-air charts. Winds reported by aircraft vary in accuracy, depending upon the method of measurement and the navigational facilities available (the latter in turn depending largely on location). Winds may be a mean value over part of the aircraft's track, or they may be 'spot' winds. The former are usually derived by comparing air speed and ground speed between navigational fixes, the latter by use of Doppler radar equipment. The aim should be to achieve a root-mean-square error of 10 knots or less in speed and 10 degrees or less in direction:[13] in favourable circumstances the error can be reduced to 5 knots or so.

Care must be exercised in the interpretation of wind reports from aircraft, particularly when mean winds over a distance are given. For example, if the aircraft is flying at an angle to a strong flow with marked horizontal shear, a mean wind may well be misleading. Errors in the reported height of the aircraft may well be transformed into wind errors in regions of strong vertical wind shear, and so on. Aircraft reports, by the nature of the conditions in which the observations are made, are more liable to coding and transmission errors than are routine reports from fixed stations, and all possible checks should be made if an aircraft report does not appear to fit in with the current analysis.

16.5 STATISTICS OF SURFACE AND UPPER-WIND RELATIONSHIPS AND THEIR USE IN FORECASTING THE SURFACE WIND

16.5.1 The relationship between surface and geostrophic winds

The basic tool used by the forecaster when attempting to predict the surface wind several hours ahead is the forecast chart of sea-level isobars, from which he is able to measure a geostrophic or gradient wind. He must then apply his knowledge and experience of the relationships between the geostrophic wind and the surface wind. These relationships depend upon the ways in which the momentum of the flow in the free atmosphere is brought down to the surface layer, which in turn depends upon the vertical temperature structure of the atmosphere at the time and upon the topography of the locality. The complexity of the relationships makes it impossible to give detailed guidance for individual stations or areas, but some of the studies which have been carried out illustrate some general principles and are discussed below.

16.5.1.1 Surface and geostrophic winds (Shaw). Some data regarding the ratio of the surface wind (at anemometer height) to the geostrophic wind for 16 wind directions at a number of locations were compiled by Shaw.[14] At some localities the ratios were determined by using measured surface winds for all speeds but at others the ratios were computed only from winds estimated as Beaufort force 4 (11 to 16 knots) or winds estimated at 12 metres per second (23 knots). Some values for the ratios are based on observations for a single year only. The data are therefore not homogeneous. Data for a number of sites are reproduced in Table 16.2. It is readily seen that there are substantial variations in the ratio from site to site and for different wind directions at each site.

16.5.1.2 Surface and geostrophic winds at Gorleston (Durst). A more detailed investigation for this site was carried out by Durst.[15] Gorleston is situated on the East Anglian coast, which in that neighbourhood runs in a broadly north/south direction. Coastal effects such as sea-breezes will exert some influence, but it is to be expected that offshore winds will show many of the properties of winds well inland, while onshore winds will show a reasonable resemblance to winds over the sea.

TABLE 16.2 The ratio of surface wind speed to geostrophic wind speed for a number of stations in the British Isles

Location	No. of years' observations	Forces compared	Height of vane or cups above ground	Nature of site	Height of ground	0° N	22½° NNE	45° NE	67½° ENE	90° E	112½° ESE	135° SE	157½° SSE	180° S	202½° SSW	225° SW	247½° WSW	270° W	292½° WNW	315° NW	337½° NNW	Mean of all
			m		m																	
Falmouth	1	All	12	Between cliff and harbour	51	.36	.36	.33	.30	.31	.32	.32	.31	.29	.28	.29	.33	.34	.35	.34	.36	.32
Pendennis Castle	1	All	20	Conical headland	78	.58	.66	.74	.83	.85	.83	.74	.69	.63	.58	.54	.52	.53	.56	.54	.54	.65
Pyrton Hill	2	All	30	Slope: hills to plain	151	.39	.38	.48	.48	.40	.37	.37	.42	.48	.40	.42	.43	.47	.49	.44	.40	.43
Southport	5	All	19	Flat sand shore	5	.69	.60	.53	.49	.48	.47	.47	.45	.41	.38	.38	.42	.51	.63	.65	.69	.52
St Mary's, Scilly	8	4	10	Hilly island	36	.74	.63	.59	.59	.67	.70	.59	.47	.51	.45	.49	.49	.54	.63	.65	.61	.58
Aberdeen	8	4	23	College roof	14	.35	.45	.61	.57	.51	.45	.43	.35	.36	.32	.32	.31	.39	.46	.54	.43	.43
Spurn Head	8	4	12	Spit of sand	8	.80	.85	.64	.72	.65	.72	.64	.54	.49	.49	.49	.53	.59	.76	.76	.73	.65
Yarmouth	8	4	12	Spit of sand	4	.72	.62	.66	.72	.62	.51	.54	.63	.38	.37	.36	.34	.39	.43	.46	.43	.51
Paisley	8	4	B	Inland station	-	.37	.33	.53	.53	.47	.33	.28	.35	.33	.35	.36	.36	.41	.36	.42	.47	.38
Carnforth	8	4	B	Inland station	-	.30	.37	.35	.40	.43	.39	.28	.25	.30	.35	.32	.31	.31	.36	.43	-	.33
Belvoir Castle	8	4	B	Inland station	-	.50	.59	.59	.44	.50	.58	.54	.50	.44	.36	.37	.37	.41	.40	.43	.36	.43
Geldeston	8	4	B	Inland station	-	.30	.30	.32	-	.42	.40	.36	.31	.29	.26	.29	.29	.29	.29	.30	.21	.30
Woburn	8	4	B	Inland station	-	.49	.49	.54	.57	.40	.45	.61	.42	.31	.43	.43	.44	.49	.50	.49	.46	.46
Holyhead	8	4	B	Flat island	5	.85	.68	.59	.57	.53	.37	.34	.49	.51	.61	.51	.48	.60	.69	.79	.87	.59
Holyhead	12	12 m/s	B	Flat island	5	.81	.75	.44	.43	.35	.17	.14	.34	.39	.49	.48	.40	.48	.60	.68	.80	.48
London, Brixton	4	12 m/s	B	Town garden	23	.28	.29	.36	.38	.31	.17	.17	.23	.24	.25	.24	.23	.23	.20	.23	-	.23
London, St James's Park	6	12 m/s	B	Town park	8	.25	.29	.29	.29	.24	.13	.11	.12	.13	.23	.24	.22	.18	.18	.24	.19	.21

B: Beaufort force estimated

Occasions when the pressure gradients lay between north-west and north-east and also between north-east and south-east were excluded from the investigation - presumably because they would lead to surface winds with directions which were close to that of the coastline. The remaining data were arranged in two groups: one with pressure gradients leading to onshore geostrophic winds (that is, between north-east and south-east) and the other with offshore geostrophic winds between south-west and north-west. The geostrophic values were compared with surface winds recorded at Gorleston. The data used were for the years 1936-40. Values for the months of May, June, July and August were used to determine the summer characteristics and those for January and December for the winter characteristics. The frequencies of the ratios of the surface wind to geostrophic wind were obtained and the results are shown in Figures 5 and 6.

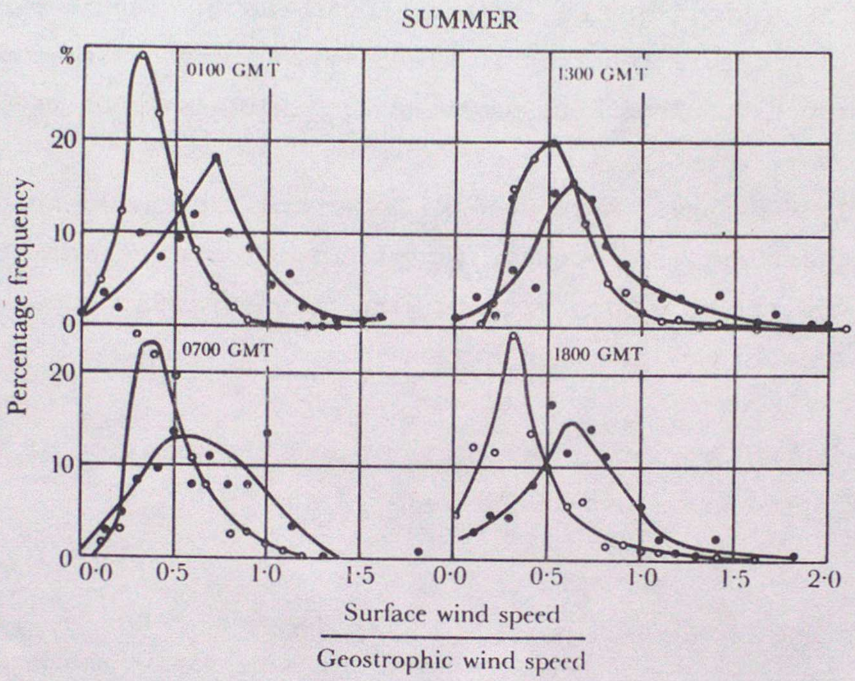
The diurnal variation of the ratio and the relation of the ratio to the wind speed are shown in Tables 16.3 and 16.4.

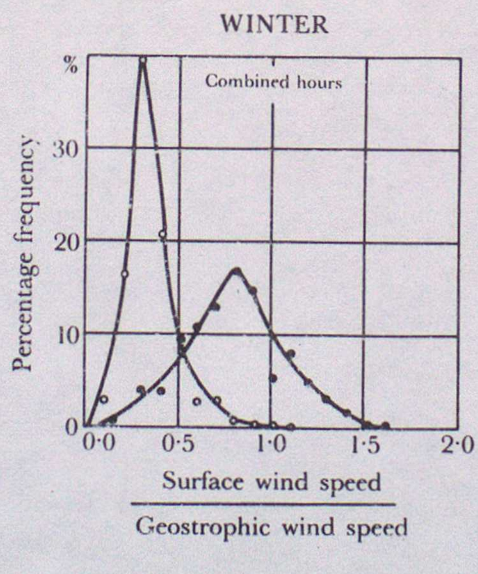
TABLE 16.3 Diurnal variation of the ratio of surface to geostrophic wind

	Time of observation (GMT)							
	0100		*0700*		*1300*		*0800*	
	on-shore	*off-shore*	*on-shore*	*off-shore*	*on-shore*	*off-shore*	*on-shore*	*off-shore*
Winter								
December, January	0.75	0.32	0.82	0.32	0.75	0.35	0.83	0.30
Summer								
May, June	0.65	0.39	0.62	0.52	0.63	0.58	0.64	0.46
July, August	0.70	0.35	0.75	0.41	0.92	0.49	0.63	0.30
All summer months	0.67	0.36	0.65	0.45	0.65	0.52	0.63	0.38

TABLE 16.4 Variation with wind speed of the ratio of surface to geostrophic wind

		Wind speed (knots)					
		9-17		*18-26*		*27-35*	
	Time	*on-shore*	*off-shore*	*on-shore*	*off-shore*	*on-shore*	*off-shore*
Winter	*GMT*						
December, January	(0100	0.87	0.40	0.88	0.44	0.77	0.26
	(1300	0.75	0.37	0.82	0.36	0.77	0.33
Summer							
May--August	(0100	0.65	0.42	0.68	0.32	(0.73)	0.32
	(1300	0.66	0.59	0.60	0.49	0.55	0.45





∘ Offshore • Onshore

FIGURE 5. Seasonal and diurnal variation of the ratio of surface to geostrophic wind at Gorleston

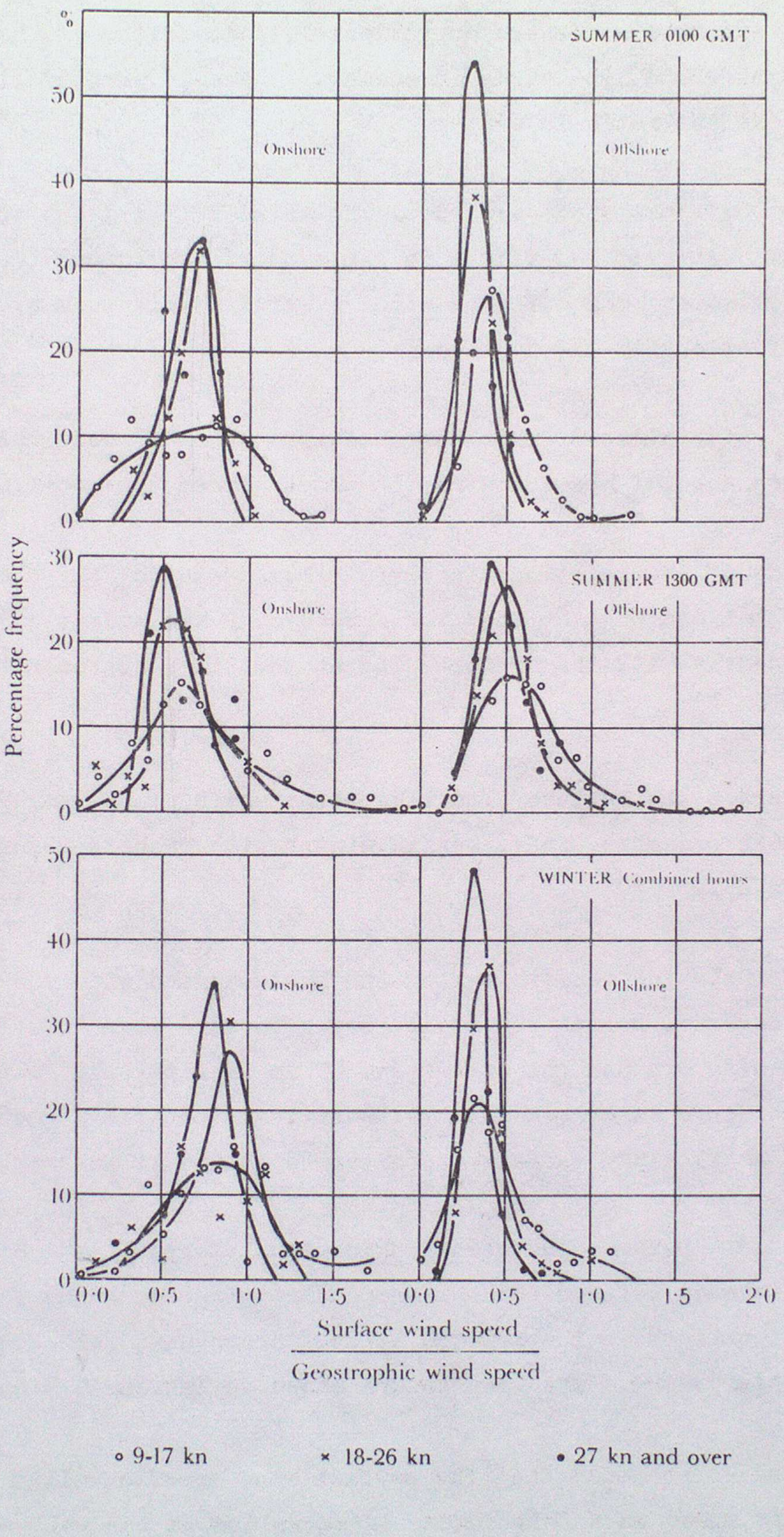


FIGURE 6. Variation with wind speed of the ratio of surface to geostrophic wind

The broad conclusions are that:

(a) Onshore winds are approximately four-fifths of the geostrophic wind in winter and two-thirds in summer. There is practically no diurnal variation in the ratio in winter.

(b) Offshore winds are about one-third of the geostrophic wind in winter and just over two-fifths in the summer. In summer there is a marked diurnal variation; the offshore winds are relatively weak at night and relatively strong in the daytime.

(c) with a geostrophic speed of less than 18 knots the onshore wind has a wide range of speed and may at times exceed the geostrophic speed.

A study of the relationship for various states of sky showed little of value, but did suggest that the more cloudy the midday sky in summer the more nearly will the onshore wind at that time approach the geostrophic value.

On average the surface wind lies at an angle of 30° to 40° to the isobars with offshore winds, but in winter with onshore winds the angle is reduced to 24°.

The correlation coefficient of the surface wind with the geostrophic wind is, however, relatively small, being no more than about 0.4 or 0.5 with both offshore and onshore winds. This implies that, although some relationships do exist between the two winds, there is a good deal of random or unexplained variation from one occasion to another.

16.5.1.3 Surface and geostrophic winds (Marshall). Marshall[16] studied the relationship between the surface and geostrophic winds for the year 1946 for four locations in the British Isles: Stornoway, Bell Rock, Scilly Isles and Kew Observatory. The results are shown in Tables 16.5 and 16.6.

Marshall also found that the surface wind speed equalled or exceeded the geostrophic speed on a fair number of occasions at the well-exposed stations, as shown in Table 16.7. The occurrences were generally associated with marked low-level instability; marked anticyclonic curvature of the isobars was also apparent at times.

TABLE 16.5 Ratio of surface wind speed to geostrophic wind speed

Location	Geostrophic wind direction	No of cases	Time (GMT) 0000	0600	1200	1800	All hours
Stornoway	Northerly	218	0.38	0.41	0.54	0.50	0.46
	Easterly	184	0.53	0.54	0.66	0.64	0.60
	Southerly	392	0.53	0.49	0.54	0.53	0.52
	Westerly	462	0.47	0.48	0.58	0.53	0.52
Bell Rock	Northerly	198	0.62	0.77	0.66	0.64	0.67
	Easterly	188	0.90	0.85	0.85	0.80	0.85
	Southerly	327	0.58	0.58	0.60	0.66	0.61
	Westerly	504	0.69	0.64	0.67	0.72	0.68
Scilly Isles	Northerly	193	0.80	0.79	0.72	0.86	0.79
	Easterly	269	0.66	0.63	0.62	0.58	0.62
	Southerly	259	0.56	0.58	0.61	0.54	0.57
	Westerly	515	0.72	0.68	0.68	0.71	0.70
Kew Observatory	Northerly	174	0.27	0.26	0.48	0.38	0.35
	Easterly	255	0.38	0.37	0.56	0.50	0.45
	Southerly	200	0.27	0.28	0.42	0.33	0.32
	Westerly	563	0.28	0.29	0.41	0.36	0.34

TABLE 16.6 Percentage frequency of different deviations of the surface wind from the geostrophic wind

	Deviation from geostrophic wind: Backed 50-70°	Backed 20-40°	Within 10°	Veered 20-40°	Veered 50-70°
	%	%	%	%	%
Stornoway	18	58	19	1	0.4
Bell Rock	4	23	48	18	2
Scilly Isles	5	37	48	7	1
Kew Observatory	21	55	16	2	0.2

TABLE 16.7 Percentage frequency of occurrence of surface winds equalling the geostrophic wind

Geostrophic wind	*Stornoway*	*Bell Rock*	*Scilly Isles*
	%	%	%
Northerly	3	16	22
Easterly	11	39	8
Southerly	5	6	6
Westerly	5	13	12

16.5.1.4 Further remarks. Some very detailed information of surface winds observed at Cardington is contained in 'Geophysical Memoirs' No. 54 which is a comprehensive report of a detailed investigation of the structure of wind over level country. The reader who wishes to study this subject more deeply should consult the original papers since the results are, in general, too detailed and lengthy for inclusion in this handbook.

The statistics on surface winds over land which have been included in this subsection show quite clearly that there is no simple formula by which the surface wind can be accurately forecast from a forecast of the geostrophic wind. There are important variations with the time of day and also with the synoptic type. However, the characteristics of the site are probably of greatest importance in estimating probable surface winds from geostrophic estimates. Local investigations of past records of surface winds will often reveal some characteristics of surface winds which are of considerable value to forecasters, particularly to those fresh to that site or locality. To be useful such investigations need not be grandiose or ambitious and even quite limited studies, well within the compass of staff at outstations, are worth while. Outstations are strongly recommended to devise and undertake such work when resources are available.

16.5.2 Relationship between surface wind and the wind near the top of the friction layer

In order to use the forecast chart of surface isobars to predict the surface wind several hours ahead, the forecaster must have some idea of the relationships between the surface and the geostrophic or gradient wind for the forecast area. Often such knowledge is gained by experience over many years, usually being passed on from one forecaster to another by word of mouth or in local forecasting rules. The more quantitative and objective the information passed on, the better, but to form soundly based, objective forecasting rules requires data for a period of several years. This means that an investigation, once started, must be carried on for several years before useful results emerge, or that data for a number of past years must be studied - a daunting task when geostrophic winds must be measured at several hours for each day of the period! Also, geostrophic or gradient winds are often difficult to estimate, and uncertainties arise when the isobars have marked curvature or the pressure pattern is changing rapidly; the difficulties are magnified when geostrophic winds over the sea are required, since isobars often cannot be drawn with sufficient accuracy for the purpose. It is with these thoughts in mind that some investigators have turned to the use of actual winds at a height where it could be expected that they would be reasonably representative of the winds just above the boundary layer. The lowest level at which winds are reported from upper-air stations is either 900 millibars or 900 metres, nominally. In practice, the reported values are mean winds over a layer 700 or 1050 metres thick; reported winds at both 900 metres and 900 millibars are likely, therefore, to contain some of the effects of surface friction, while winds at the latter level will have the additional complication that the height of the 900-millibar surface varies over the approximate range of 400 to 1200 metres above mean sea level in the region of the British Isles. Nevertheless, some useful results have been obtained, particularly from the comprehensive investigation by Findlater et alii.[18] They studied the ratio of the surface wind speed, V_0, to the speed of the 900-millibar wind, V_{900}, and the angle, α, by which the surface wind was backed from the 900-millibar wind, for four locations: Ocean Weather Stations (OWS) 'I' and 'J', London/Heathrow and Leuchars.

Findlater et alii classified their results according to the overall lapse rate from the surface to 900 millibars and the 900 millibar wind speed. The lapse-rate classification is shown in Figure 7. The detailed reasoning behind this classification is given in the original paper; briefly, it was felt that errors in the temperature soundings would lead to errors in classification if the boundaries of the categories were along, for example, an isothermal or a saturated adiabatic.

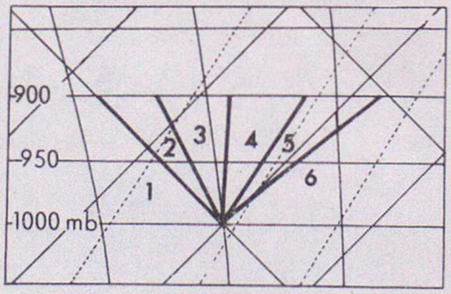


FIGURE 7. Schematic representation of lapse-rate classes

1. Superadiabatic
2. Conditionally unstable
3. Conditionally stable
4. Stable
5. Isothermal
6. Inversion

The main results of the investigation are summarized in Tables 16.8 and 16.9.

For many purposes, however, the forecaster has only the forecast surface isobars, from which he can derive a forecast geostrophic or gradient wind, to guide him in forecasting the surface wind. If he is to use the relationships discussed in this section, therefore, he must

TABLE 16.8 V_0/V_{900} and α for various wind-speed and lapse-rate classes for Ocean Weather Stations 'I' (59°00'N, 19°00'W) and 'J' (52°30'N, 20°00'W)

900-mb wind-speed class (knots)

Lapse class	*10–19* V_0/V_{900}	α°	*20–29* V_0/V_{900}	α°	*30–39* V_0/V_{900}	α°	*40–49* V_0/V_{900}	α°	*≥50* V_0/V_{900}	α°
1	0.97	0	0.91	0	0.85	0	0.81	0	0.78	0
2	0.90	5	0.84	5	0.79	5	0.76	5	0.74	5
3	0.85	10	0.77	12	0.71	15	0.67	12	0.64	10
4	0.80	17	0.72	20	0.66	20	0.61	22	0.59	20
5	0.75	15	0.68	18	0.63	20	0.58	20	0.54	25
6					see note below					

Note

Data for lapse class 6 are too few to allow of any classification according to wind speed. Overall values for all wind speeds are

$$V_0/V_{900} = 0.75$$
$$\text{and } \alpha = 25^\circ$$

TABLE 16.9 V_0/V_{900} and α for London/Heathrow Airport and Leuchars by day and by night

Lapse class	*Heathrow* *day* V_0/V_{900}	α°	*Heathrow* *night* V_0/V_{900}	α°	*Leuchars* *day* V_0/V_{900}	α°	*Leuchars* *night* V_0/V_{900}	α°
1	0.70	0)			0.60	5)		
		)				)	0.40	15
2	0.60	5)	0.35	25	0.45	15)		
		)						
3	0.45	20)			0.40	25	0.35	25
4	0.40	30	0.30	35	0.35	35	0.30	30
5)					0.30	40	0.25	40
)	0.40	35	0.30	40				
6)					0.20	50	0.15	20

Notes

1. The results were more variable over land than over the sea, but some tendencies do emerge, and the pattern of values at Leuchars is not very different from that at Heathrow, although the topography of the two stations varies markedly.

2. The values are given, to the nearest 0.05 for V_0/V_{900} and 5° for α, for all wind speeds together. There were variations with wind-speed class for some lapse classes, but no consistent pattern was apparent.

3. Upper-air data for Crawley, Sussex were used with the Heathrow surface data; the two stations are 39 kilometres apart. Leuchars (surface winds) is only 5 kilometres from Shanwell (900-millibar winds).

assume some kind of relationship between the 900-millibar wind and the surface geostrophic wind. Two relevant studies have been reported by Findlater et alii[18] and by Boyden.[6] Findlater et alii compared the 900-millibar wind at Crawley with the surface geostrophic wind; the results are given in Figure 8.

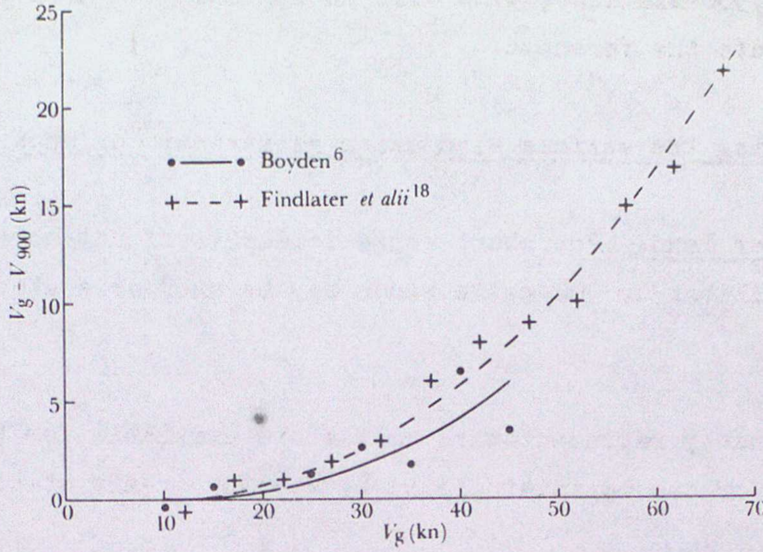


FIGURE 8. Difference between geostrophic (V_g) and 900-mb wind (V_{900}) for various values of geostrophic wind

Also given in this diagram are the results of a comparison between the 900-metre wind and the surface geostrophic wind given by Boyden. The difference between the two curves may be at least partly explained in two ways:

(a) the variation of the height of the 900-millibar surface from observation to observation, surface friction effects being particularly pronounced when the 900-millibar height is low;

(b) Boyden restricted his comparison to occasions when the 900-metre flow was straight and steady, while there was apparently no such restriction in the data of Findlater et alii.

Both reports provide a valuable account of the many factors determining the relationship between the surface geostrophic, the 900-millibar, the 900-metre and the actual surface winds.

For geostrophic winds of 40 knots or less, the differences are quite small, $\leqslant$ 5 knots, but the differences increase rapidly for greater speeds. However, errors in forecasting the surface isobars, and in estimating from them a geostrophic wind, will also increase, and it appears that the errors arising from the assumption that the geostrophic wind is equal to the 900-metre, and probably also the 900-millibar, wind will be small in comparison with other errors entering into the forecast.

16.5.3 Forecasting the surface wind using geostrophic or upper winds

16.5.3.1 Over land. For short-range forecasts of not more than a few hours, 900-millibar or 900-metre winds may be used as a starting point if:

(a) reasonably representative values are available for the area of interest or can be interpolated between upper-air stations, and

(b) allowance can be made, using surface prognoses as a guide, for changes occurring between the time of the latest observation and the time for which the forecast is required.

For longer-range forecasts, or if the above conditions are not met, the geostrophic wind must be used as the starting point. There may be occasions when both forecast geostrophic and reported winds will be useful.

In the following remarks it will be assumed that the surface geostrophic and 900-millibar or 900-metre winds are equivalent, but it must be borne in mind that some adjustment may be required, along the lines indicated in the last section, if one is substituted for the other in the relationships described in 16.5.1 and 16.5.2.

It is clear from the values quoted in 16.5.1 and 16.5.2 that merely to back the geostrophic wind invariably by, say, 30° and reduce its speed to a constant fraction will not always produce a good forecast of the surface wind. Nevertheless a general working rule might be to forecast surface winds as one-third of the geostrophic wind and backed by about 30°,

(a) by day, unless convection is occurring, when the ratio may be increased to about 0.5 to 0.6 and the backing decreased to $10\text{-}20^{\circ}$, or

(b) at night, except when skies are clear and there is likely to be an inversion of temperature, when the ratio may fall to about 0.25 and the backing may well increase to about 40°.

If information on lapse rate in the lowest hundred metres or so is available from a nearby upper-air station or from an instrumented mast, the values of Table 16.9 may be used.

In general, however, a good deal of caution must be exercised when forecasting the surface wind over land by means of the statistical data presented above. Topography plays an important part in determining the relationships, and the forecaster must also apply his local knowledge, together with that of other forecasters who may have set down their experience in the form of local forecasting rules.

16.5.3.2 Over sea. Over the sea the surface wind is more closely related to the geostrophic wind than over the land, and the values given in Table 16.8 will normally enable a reasonable forecast of the surface wind to be made (again assuming the equivalence of the 900-millibar and surface geostrophic winds). However, for short-range forecasts based largely on actual winds or actual surface-pressure patterns, errors may arise because the sparseness of, or uncertainties in, the data may lead to less accurate analysis than is possible over land.

16.5.4 Diurnal variation of surface wind

The relationship between the wind above the friction layer and the surface wind depends upon the way in which momentum is transported downwards from the top of the boundary layer to the ground, which in turn depends largely upon the wind speed in the free air and the lapse rate, and also to a greater or lesser degree on the topography. A full discussion of this topic must include an account of the wind profile in the friction layer and its variations; a more detailed treatment will be given in 16.8 but a few remarks on the diurnal variation of the surface wind in relation to the daily lapse-rate changes are appropriate at this stage.

As air moves over the earth's surface it loses energy and momentum to the surface and, if these were not replaced from higher levels, the air motion near the surface would soon cease. The vertical transport of momentum to maintain the surface wind is brought about by mechanical stirring ('forced' convection) or by free convection. Forced convection occurs when winds are fairly strong, and particularly when skies are cloudy, so that radiation plays only a small part in determining the lapse rate which is generally somewhere near the dry adiabatic value. It is friction at the surface which causes the formation of the eddies which achieve the downward momentum transport - the vertical extent and the intensity of the mechanical stirring depend upon the wind speed and the 'roughness' of the surface. Usually, in these conditions, the lapse rate varies little with time of day, and the surface wind consequently does not change a great deal. When winds are less strong over land, and especially when skies are at least partly clear, radiation plays an important role in determining the lapse rate and there are marked variations from day to night. During the day, the incoming short-wave radiation frequently establishes in the lowest layers a lapse rate of sufficient magnitude to create unstable conditions: 'free' convection (usually referred to as just 'convection' by the non-specialist in boundary-layer meteorology) starts, providing an efficient mixing mechanism in the lowest layers. In these conditions the variation of wind velocity with height is at a minimum. At night, however, long-wave radiation from the earth's surface and the lower atmospnere often results in a stable lapse rate or inversion. Then vertical motion and momentum transport to the near-surface layers is inhibited and the surface wind speed decreases, often falling to calm. At the same time, the wind speed at higher levels often increases because momentum is no longer being lost to the air at lower levels. The air motion at the surface may no longer be linked to that at a height of a few hundred metres or so; air motion may then still occur, but it will be more dependent upon topographically induced density variations (see 16.6 below).

Some of the effects of diurnal variation are apparent in the data presented in 16.5.1 and 16.5.2. In Table 16.9 it will be noted that classification into various surface to 900-millibar lapse rates has not removed all the diurnal variation. This is probably because the relationships probably depend upon the lapse rate on a finer scale; data from instrumented high masts (Jehn and Durie,[19] Thuillier and Lappe [20]) suggest that the lapse rate in the lowest 300 metres or so of the atmosphere is more important in determining the vertical wind profile than the temperature difference between the surface and 900 millibars.

Over the sea, the diurnal variation of temperature, and hence of lapse rate, is much smaller and the diurnal variation of surface wind speed may be neglected.

Near the boundary between extensive land and water surfaces, the differing reaction of the two surfaces to radiation leads to local circulations being set up - sea-breeze by day and land-breeze at night. These circulations will be discussed in 16.6.

16.6 LOCAL WINDS

It is probably true to say that the physical principles causing 'local' winds are fairly well known and simple to understand. However, completely satisfactory theoretical and mathematical treatments of the various local winds do not exist. Some of the treatments which are available are often fairly advanced and the theoretical approaches appear to have, at the time of writing, no direct application to practical forecasting. In the following treatment a few references to theoretical papers are given but the bulk of the text is confined to descriptions of local winds and to information of direct practical value.

16.6.1 Sea-breeze

16.6.1.1 Description and general remarks. The cause of the sea-breeze is to be found in the different behaviour of land and water surfaces and the overlying air when subject to incoming solar radiation. When radiation reaches the ground the temperature of the surface rises sharply. This sets up temperature gradients in the soil beneath the surface and in the lower layers of the air. A similar state of affairs occurs at a water surface. There is however, a fundamental difference in the effect. In the water, incoming radiation can penetrate beneath the surface and turbulent mixing of the surface layers by wind and waves causes a net downward transport of surface heat throughout relatively large masses of water so that the rise of temperature of the surface is moderated. Over land there is neither direct penetration of incoming radiation nor turbulent transport of heat downwards through the ground from the surface and the transmission of heat downwards is by conduction only. The land surface reaches higher temperatures than the water surface and so air in the lower levels over land reaches higher temperatures than at the same levels over water. Other things being equal, warmer air is less dense than colder air so that, at a given height above mean sea level within the stratum of air warmed by the heated land surface, the pressure is greater over the land than over the water.

At such an upper horizontal level there is a pressure gradient from land to sea and, in the absence of a general pressure gradient, an upper wind blows from land to sea. Air is thus exported from land to sea and the pressure at the sea surface commences to rise while at the land surface it commences to fall. This sets up a surface pressure gradient from sea to land and the sea-breeze flows down this pressure gradient. Since this gradient is normally at right angles to the shore the sea-breeze usually commences to blow also at right angles to the shore.

The surface sea-breeze is only one leg of a mesoscale circulation: ascending air over the land (often assisted by convection), outflowing air at higher levels, and gently subsiding air over the sea completing the picture. At first the surface sea-breeze flows directly from high pressure to low, that is from sea to land at right angles to the coastline. As the sea-breeze continues to blow, the rotation of the earth causes the wind to blow at larger angles to the thermally induced pressure gradient, so that late in the day it may be blowing almost parallel to the coast.

While the formation of a sea-breeze on a given day depends largely upon the setting up of a suitable temperature difference between the air over the land and that over the sea, the behaviour after formation depends largely upon other factors, such as the pre-existing wind, the temperature profile over the land, the topography of the area, and surface friction.

In the discussion above it has been tacitly assumed that the initial conditions were calm, but there is often a pre-existing wind (surface, geostrophic or gradient), and this is usually a major factor in determining the behaviour of the sea-breeze. If the geostrophic or gradient wind is initially onshore, the sea-breeze may penetrate well inland, but its arrival at any place may not be easy to detect, as changes of wind and temperature will not usually be very marked. If the pre-existing wind is light, the inland penetration may not be so great, but the effects will be more noticeable. If there is an offshore geostrophic wind, not too strong, penetration will be still slower, while the initial formation of the sea-breeze may take place out over the sea (see Koschmieder[21]). As the sea-breeze moves inland, however, it will produce marked changes of temperature, humidity and wind, the boundary between the land air and cooler sea air often having some of the characteristics of a macro-scale cold front. This advent of the sea-breeze is often marked by a 'wall' of haze or smoke and cumulus or cumulonimbus clouds in the rising air ahead of the front. The sea-breeze front is of particular interest and concern to glider pilots, and a number of aerial explorations have been made (Findlater,[22,23] Wallington[24]); in general, the front moves more slowly than the sea-breeze itself, with rising air near the front in the sea-breeze air itself, and a component of motion of the sea-breeze parallel to the front. Figure 9 shows some of the main features of the sea-breeze and associated front: the diagram is based partly on one by Munn,[25] with additions from Findlater.[23] The sea-breeze front will be more marked where the coastline is convex, leading to convergence in the sea-breeze air, and less marked where the coastline is concave and divergence occurs.

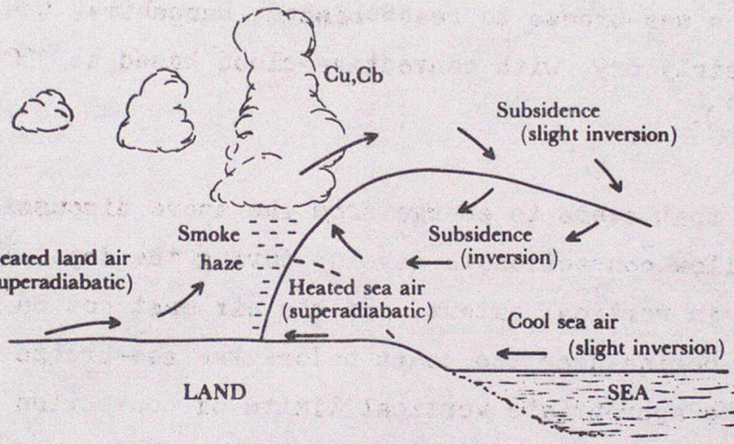


FIGURE 9. The structure of the sea-breeze

There are differing opinions on the effect of a geostrophic wind blowing parallel to the shore: Wexler[26] and Koschmieder[21] do not find them favourable for the inland penetration of a sea-breeze, but Wallington[24] found that, over southern England, a moderate or fresh conditionally unstable airstream blowing parallel to the coast was favourable for the development of a sea-breeze front. It seems quite likely that topography and the stability of the air are important additional factors.

Some doubt exists also about the precise role of convection inland in leading to the establishment of a sea-breeze and its subsequent movement. It is clear, however, that some degree of instability, leading to convection over the land is necessary, together with stability of the air over the sea.

Where the uncertainty lies is about the degree of instability and the vertical extent of convection over land necessary or favourable for sea-breezes to form. Brittain[27,28] has found that, for a sea-breeze to reach Manby, 9.6 kilometres inland from the Lincolnshire coast, convection to at least 1500 metres is necessary. On the other hand, if large amounts of deep convective cloud (extending to above 3000-4000 metres) formed, the sea-breeze front either did not move inland or did so erratically. In particular, it appears that the sea-breeze is unlikely to reach a given place if widespread convection is occurring between there and the coast. In contrast, in a study of sea-breezes at Worthy Down (Peters[29]) about half of the 58 occasions of sea-breeze were accompanied by cloudless or almost cloudless skies; this does not exclude dry convection beneath an inversion. For a sea-breeze to reach Lasham, Hampshire, the air should apparently be fairly dry, with convective cloud based at 1500 metres or above (Simpson[30]).

The pattern that seems to emerge from the above discussion is that land air must allow convection to develop during the day. The convection must be limited in vertical extent, and the air must not be so unstable that convection begins near the coast before the sea-breeze has become established. The appropriate vertical limits of convection are likely to vary from place to place in a way which may be linked to topography, aspect, etc. It is also essential that the air over the sea should not be unstable to the sea temperatures.

The part played by topography is likely to be complex. The shape of the coastline may, for example, be such that for slightly varying initial conditions the sea-breeze may set in from different directions. A range of hills near the coast is likely to obstruct the inland flow of air at low levels, but ground which slopes up from between south and east receives more heat per unit area from the sun than does a horizontal surface, and convection may be started earlier than over flat ground. Larger mountain ranges may lead to greater frequency of sea-breezes on the nearby coasts - the role of the Grampians has been discussed by Lamb;[31] the direction of the gradient wind will be important in the production of lee eddies which will exert some influence on the behaviour of the sea-breeze. The sea-breeze in valleys may be reinforced by local anabatic winds and may become quite strong.

Although it is impossible to give a comprehensive account of the effects of topography on sea-breezes, the above remarks may provide some indication of the sort of factors that local forecasters should look out for.

16.6.1.2 <u>Theoretical studies.</u> Theoretical studies of the sea-breeze have been carried out by Wexler,[26] Pearce,[32] Haurwitz[33] and Estoque,[34] among others. Pearce's model enables prediction graphs to be drawn and expressions were derived for the speed of the sea-breeze at a given time and also the inland penetration (see 16.6.1.6). Estoque presented diagrams showing the circulation for onshore, offshore and light geostrophic winds; simplified versions of his diagrams are given in Figures 10(a)-(c).

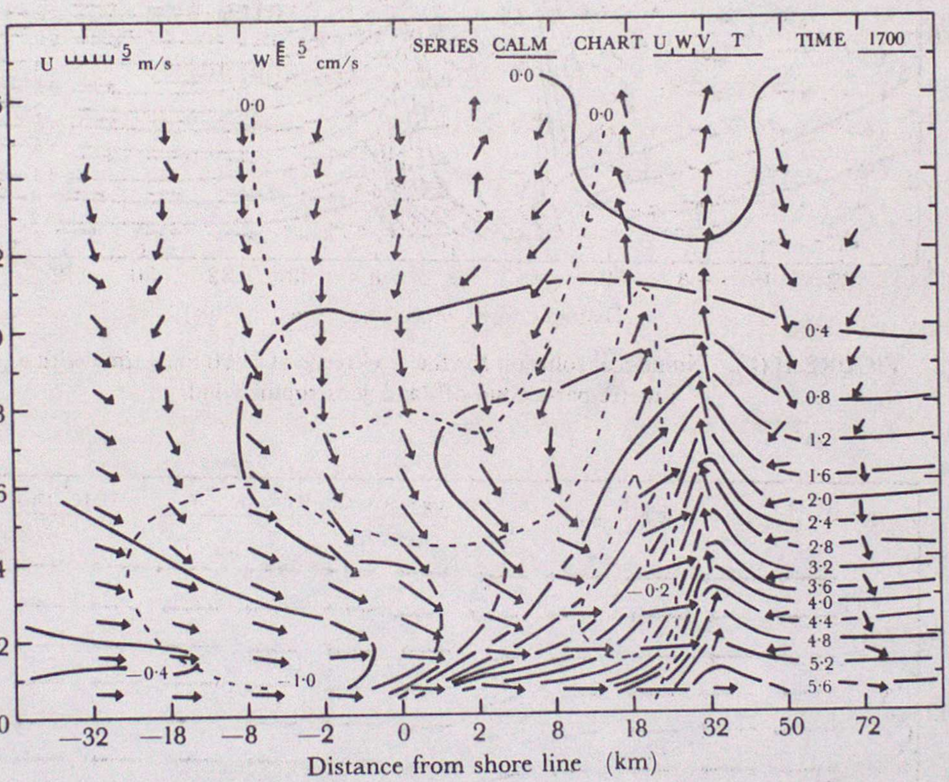


FIGURE 10(a). Numerical solution for the sea-breeze at 1700 local time with zero geostrophic wind

Vectors give landward and vertical circulation; ——— denotes temperature changes from 0800 local time; - - - - denotes velocity components in metres per second into the figure.

16.6.1.3 <u>Climatology of the sea-breeze.</u> Wickham[35] has produced a note on the climatology of sea-breezes and a useful map showing the prevailing direction of the sea-breeze at a number of locations, and the mean position of the sea-breeze fronts in various parts of Britain at

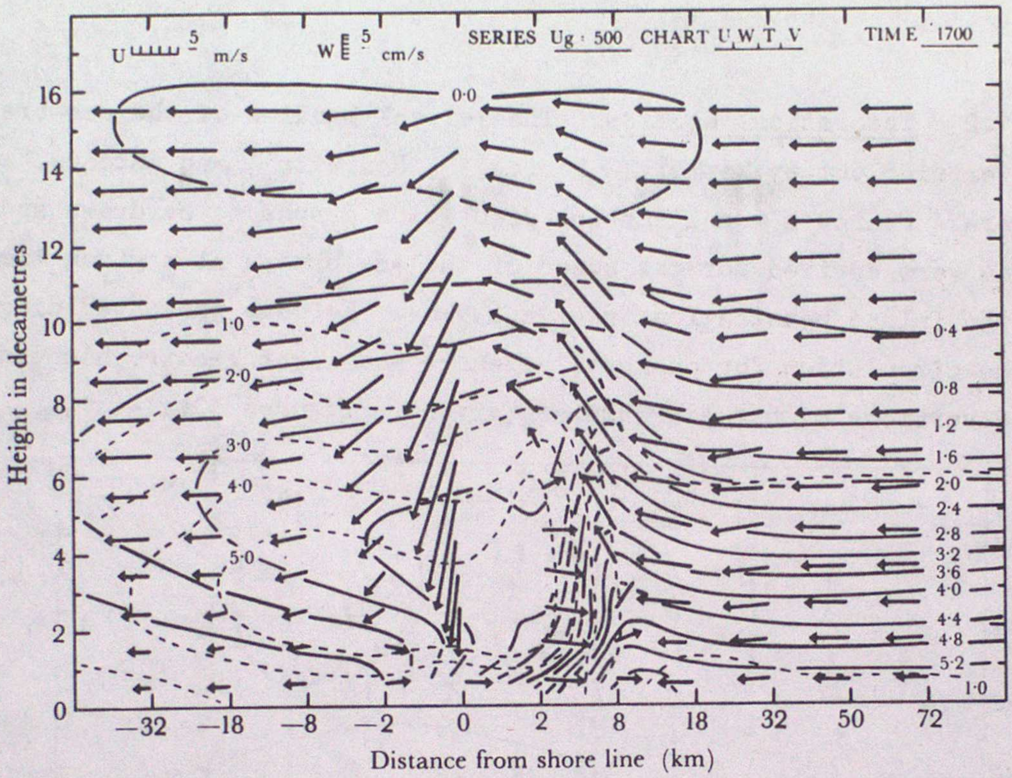


FIGURE 10(b). Numerical solution for the sea-breeze at 1700 local time with a 5-metre-per-second off-land geostrophic wind

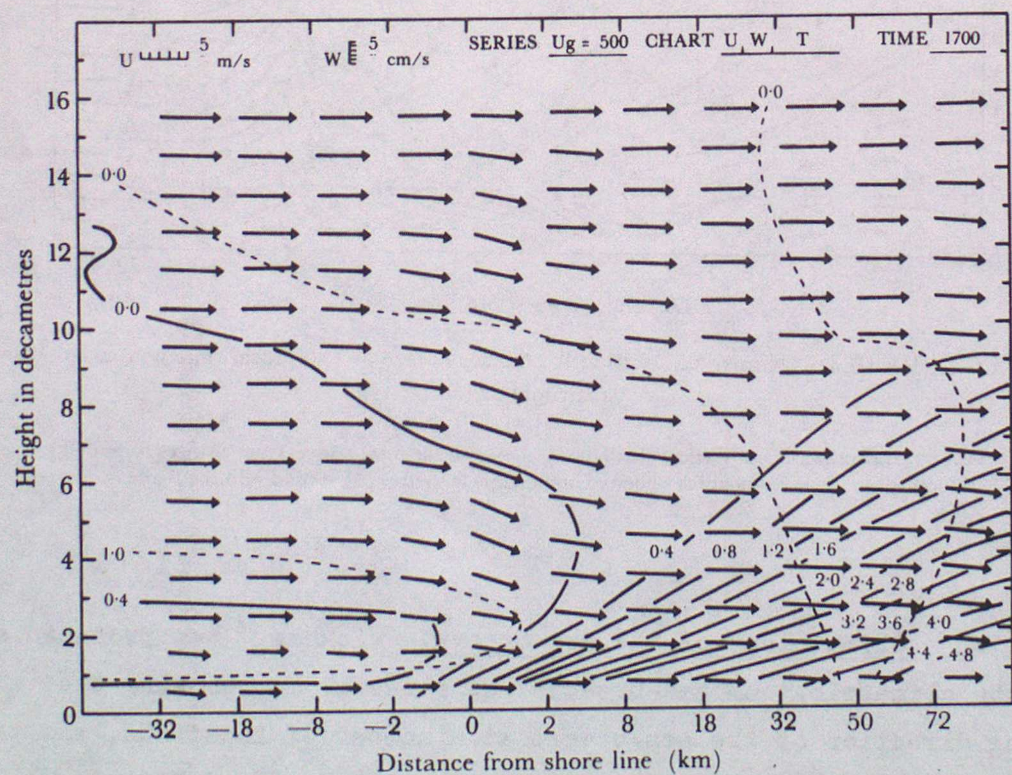


FIGURE 10(c). Numerical solution for the sea-breeze at 1700 local time with a 5-metre-per-second off-water geostrophic wind various times of day (Figure 11). The results presented do not imply that sea-breezes and sea-breeze fronts do not exist elsewhere, but that although known to exist they are not as well documented or are less consistent in their behaviour.

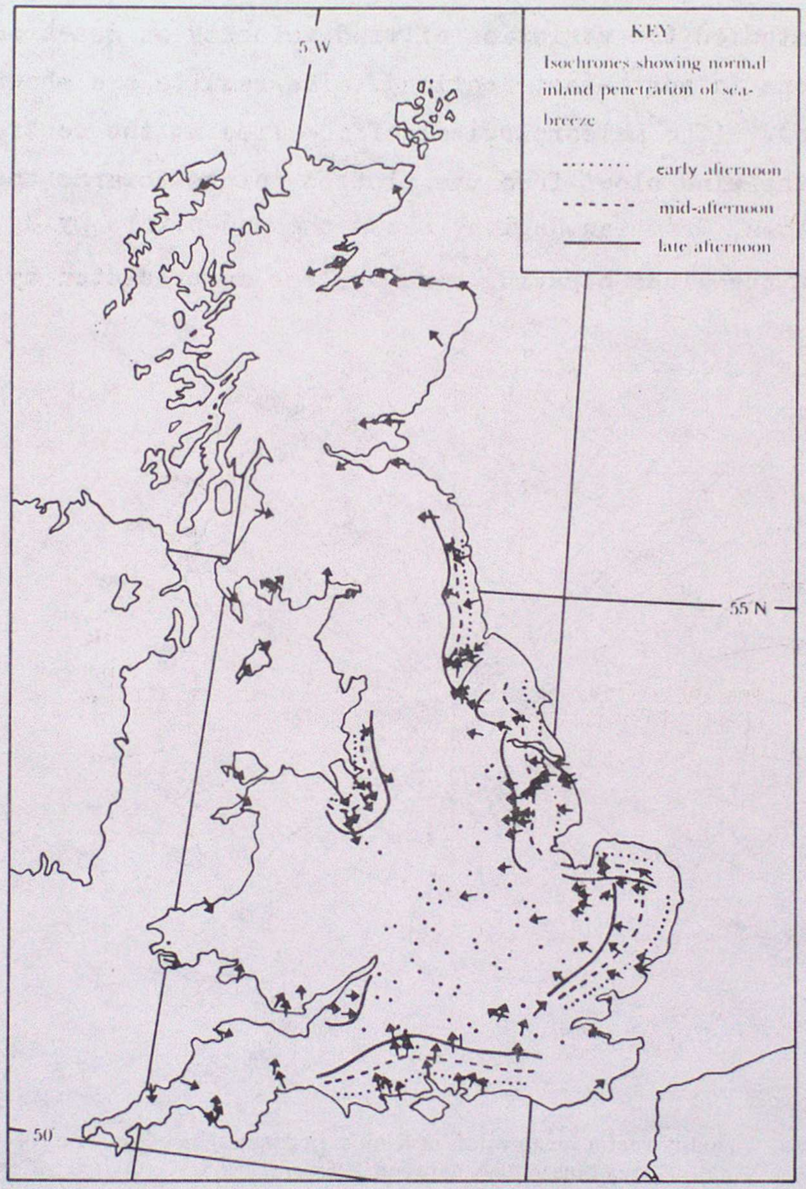


FIGURE 11. The normal direction and penetration of sea-breezes during summer months

Each arrow or dot represents an individual airfield. Arrows show the late afternoon direction of the sea-breeze (a few places having two preferred directions). Dots show airfields where sea-breezes have not been specially recorded in the local weather notes. The isochrones show the normal rate of progress inland of the sea-breeze on a summer afternoon.

16.6.1.4 <u>Diurnal variation of the sea-breeze.</u> Haurwitz[33] has given a theoretical treatment of the diurnal variation of the sea-breeze. Allowing for surface friction and for the fact that the flow is not balanced, he found that the ends of wind vectors plotted on a hodograph lay on an ellipse, and that the end-points moved round the ellipse in a clockwise direction. The eccentricity of the ellipse varied with the frictional force, being low when the frictional force was small.

Gill[36] has studied the variation of wind velocity on sea-breeze days for three stations in north-east Scotland. The results are shown in Figures 12(a)-(c). (The meteorological office lies at the centre of the hodograph, and the wind blows from the plotted points towards the origin.) The diagram for Kinloss, Grampian Region, shows the end-points lying almost on an ellipse, with the winds behaving more or less as predicted by the theory.

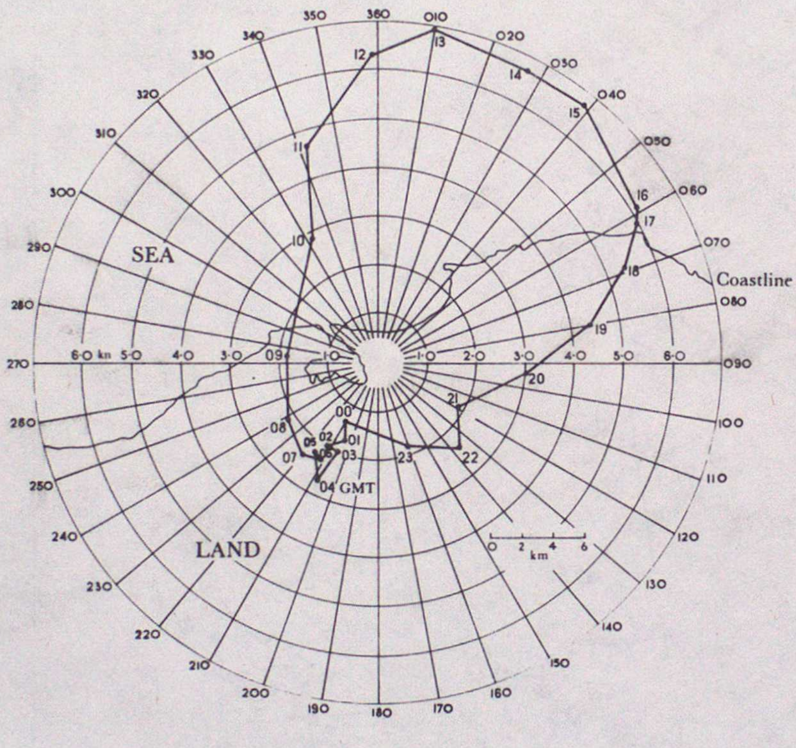


FIGURE 12(a). Hourly vector mean winds at Kinloss in August, based on data for 40 sea-breeze days between 1959 and 1965

The diagrams for other months from April to September are similar, the main difference being in the times at which the sea-breeze starts and ends - taken to be the times at which the wind vectors lie parallel to the coast. As expected at a coastal station with relatively small surface friction effects the eccentricity of the ellipse is low.

In contrast, the eccentricity of the figure for Dyce (Aberdeen) is very high, showing marked effects of friction at an inland station 8 kilometres from the coast. The hodograph for Wick, Highland Region, is not very

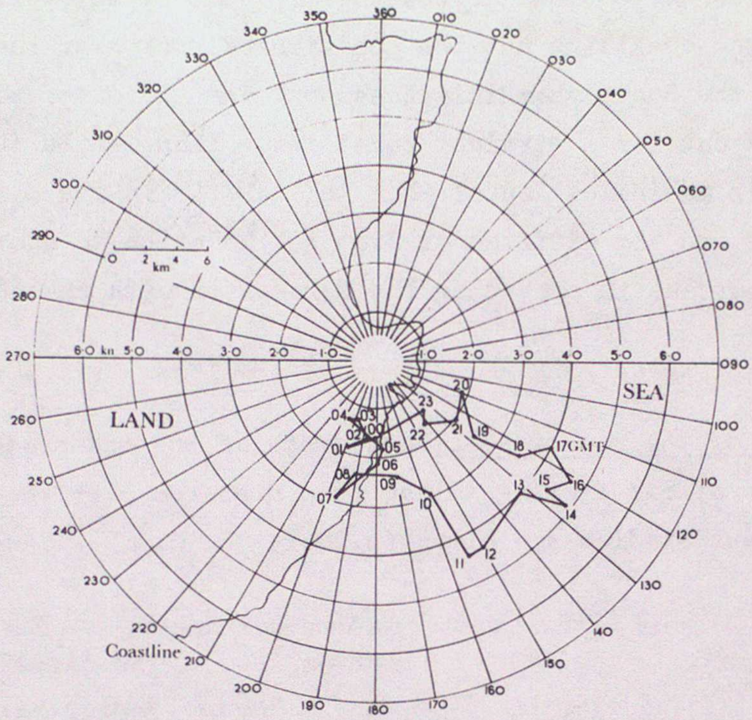


FIGURE 12(b). Hourly vector mean winds at Wick in August, based on data for 42 sea-breeze days between 1958 and 1964

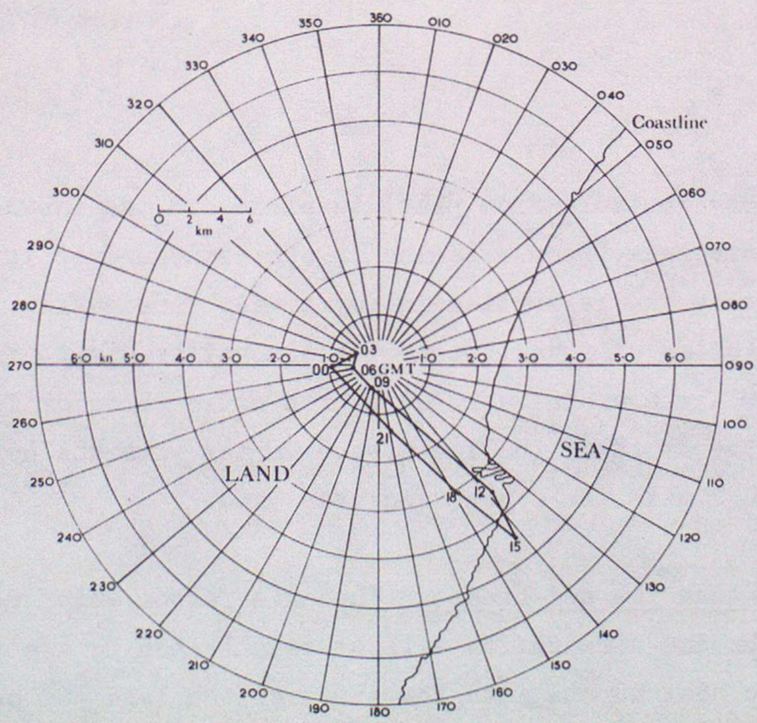


FIGURE 12(c). Hourly vector mean winds at Dyce in August, based on data for 42 sea-breeze days between 1958 and 1964 different from an ellipse, but the end-points of the vectors move around the ellipse in a counter-clockwise direction, opposite to that predicted by the theory. The reason for this is not certain, but it may well be that the irregularity of the coastline and the position of Wick near the north-eastern extremity of the Scottish mainland account for the discrepancy - the theory was worked out for a straight coastline. Thus, while the above results can be used as an indication of what to look for in the data, they go to show that there are big differences from one location to another, and local studies are important in providing the forecaster with the necessary background knowledge.

16.6.1.5 Depth of the sea-breeze. The depth of the sea-breeze is taken to be the depth of air flowing inland from a source over the sea. Values reported in four studies are given in Table 16.10.[37,38]

TABLE 16.10 Depth of sea-breeze, standard deviation and extreme values at four stations

Station	*Distance from sea*	*Average depth*	*Standard deviation*	*Extreme values*
	km	*metres*	*metres*	*metres*
Felixstowe (Suffolk)	1.6	500–600	–	–
Kinloss (Grampian Region)	2.4	300	180	60–850
Tangmere (East Sussex)	8.0	750	–	150–1200
Worthy Down (Winchester, Hants.)	40.2	500	–	–

The results are very variable from place to place and, as shown by the Kinloss and Tangmere data, from occasion to occasion, and it is difficult to see how they can be applied in day-to-day forecasting. There is some evidence at Kinloss of a variation with time after onset, from 200 metres or so after one hour to over 300 metres after three or four hours. It seems likely that the depth of the sea-breeze varies with the height of the inversion which limits the convection over land.

16.6.1.6 Forecasting the sea-breeze. The main factor which determines whether or not a sea-breeze circulation will be established is the difference between the heat inputs into the air over the land and over the sea; in most forecasting techniques this is expressed in terms of a closely related, but not synonymous, quantity, the difference between the temperature of the air at screen level over the land and the sea-surface temperature.

Also probably important at times in determining the formation of the sea-breeze, but of prime importance in influencing its penetration inland, is the wind field which would exist in the absence of the sea-breeze: this is usually expressed as the geostrophic or gradient wind appropriate to the forecast sea-level isobars or as the actual surface wind early in the morning before the start of the sea-breeze. Also important on many, if not all occasions is the stability of the air mass and the vertical extent and horizontal distribution of convection. All these factors have been discussed in the preceding sections; the forecasting problem is to find quantitative relationships between the predictors and the behaviour of the sea-breeze, in particular the critical value of each predictor determining whether a sea-breeze forms or not. These relationships will be different for different places for various reasons - distance from the sea, topography, aspect, and the method of measurement or estimation of sea temperatures, etc.

The most common way to derive the necessary relationships is to plot a scatter diagram with the offshore component of the geostrophic, gradient or pre-existing actual wind along one axis and land-sea temperature difference or some measure of inland heating along the other; a line is then drawn which best separates the occurrences from the non-occurrences. Slight variations on this theme have been used: for example, a diagram by Watts[39] for Thorney Island (between Chichester and Portsmouth) uses the actual wind speed at 900 metres, five demarcation lines being drawn for various wind sectors. This diagram is reproduced in Figure 13; the 'sea' temperature was measured at a depth of 60 centimetres in Chichester Harbour.

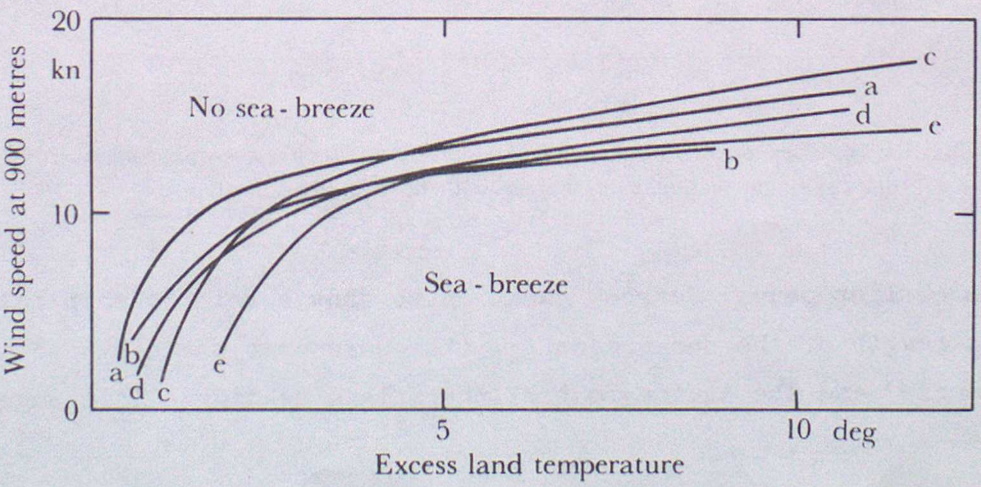


FIGURE 13. Occurrence of sea-breeze at Thorney Island

Direction of 900-metre wind:

(a) 280°–319° (d) 020°–059°
(b) 320°–349° (e) 060°–089°
(c) 350°–019°

This diagram is reproduced as an example, and should not be used for other locations. A similar diagram has been obtained by Pearce[40] on theoretical grounds, the ordinate being the offshore component of the general wind and the abscissa the amount of inland heating expressed in terms of the large squares on the tephigram (also given is the equivalent depth, d, in millibars, of the layer which is heated from an isothermal to a dry adiabatic state; see Chapter 17, section 17.6.1.1). The diagram is for what Pearce calls 'average stability', with no highly stable layers below 2 kilometres but no vigorous convection.

In the original diagram the lines separating the various types of occasion are parabolas, i.e. for a given stability the resultant effect on the circulation depends upon the ratio A/u^2. If the diagram is redrawn with d as abscissa, the lines become straight, as in Figure 14, a result of the approximate proportionality of A and d^2.

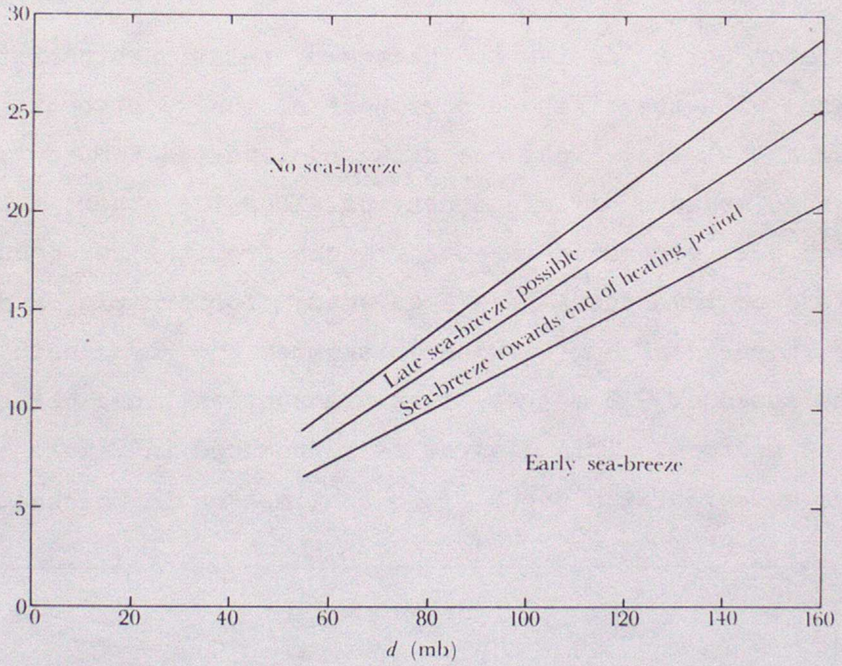


FIGURE 14. Dependence of formation of sea-breeze on (a) off-shore wind component, u_p, and (b) inland heating, expressed as depth, d, in millibars, heated from isothermal to dry adiabatic state

In his earlier paper, Pearce[32] went on to show a relationship between the mean strength of the sea-breeze U_m, (the mean over the distance inland from the shore) and the square root of the inland heating. This suggests that

$$U_m \propto d$$

where d is the depth of the layer heated from an isothermal to a dry adiabatic state by the time for which U_m is to be estimated. Values of d

can be obtained from Table 17.15 of Chapter 17 - Temperature. The constant of proportionality, although worked out by Pearce for use with A, is likely to vary from place to place and should be determined empirically for each location.

Similarly, Pearce suggests that the inland penetration by time T after sunrise is proportional to $TA^{\frac{1}{2}}$; by using d^2 instead of A, this becomes

$$y \propto Td.$$

Again the constant of proportionality must be determined from local observations. The maximum inland penetration is probably given reasonably well by the values of T and d at the time of maximum heating.

It should be emphasized that the above formulae were derived by Pearce on theoretical grounds for a simple case. Although no practical results have been published to support the theory, the ideas seem worthy of a trial.

Brittain[27,28] studied the inland penetration of the sea-breeze over Lincolnshire and put forward a model based on several simplifying assumptions, principally the following:

(a) The sea-breeze does not start until convection reaches 1500 metres.

(b) The pressure gradient set up at low levels by inland heating corresponds to a geostrophic wind, V_g, of 15 knots blowing parallel to the coast, the corresponding surface wind, V_0, being 2/3 of the geostrophic, i.e. 10 knots.

Otherwise, friction is neglected.

Following Gordon,[41] Brittain assumed that an initially stationary particle coming under the influence of the pressure gradient set up by inland heating would begin to move inland with increasing speed, but that as soon as it started moving it would come under the influence of the Coriolis force and would acquire an increasing component of motion parallel to the shore. The components of the sea-breeze would be given by

$$u_s = V_0 (1 - \cos ft)$$

$$u_p = V_0 \sin ft \qquad \ldots . \; 16.8$$

where u_s is the component parallel to the shore, u_p that perpendicular to the shore, f is the Coriolis parameter and t is the time after the setting-up of the pressure gradient.

If there is already a component of wind, v_p, perpendicular to the shore before the sea-breeze formation (v_p positive onshore) then the sea-breeze will not start to move inland until

$$u_p + v_p \geqslant 0 \qquad \text{. . . . 16.9}$$

which occurs at a time, t^1, given by

$$t^1 = \frac{1}{f} \sin^{-1}\left(-\frac{v_p}{V_0}\right) \qquad \text{. . . . 16.10}$$

After t^1 the sea-breeze moves inland at a speed, $u_p + v_p$, which increases to a maximum at $t = \pi/2f$ and falls to zero at $t = (\pi/f) - t^1$ when inland penetration ceases. The distance penetrated inland is given by

$$y = \int_{t^1}^{t} (u_p + v_p)\, dt \qquad \text{. . . . 16.11}$$

The results are given in Figure 15 (on the facing page), each curve giving the inland penetration after time t for a given value of v_p. Of course, the land-sea temperature difference must be suitable for the formation of a sea-breeze; the relevant conditions are given in Figure 16.

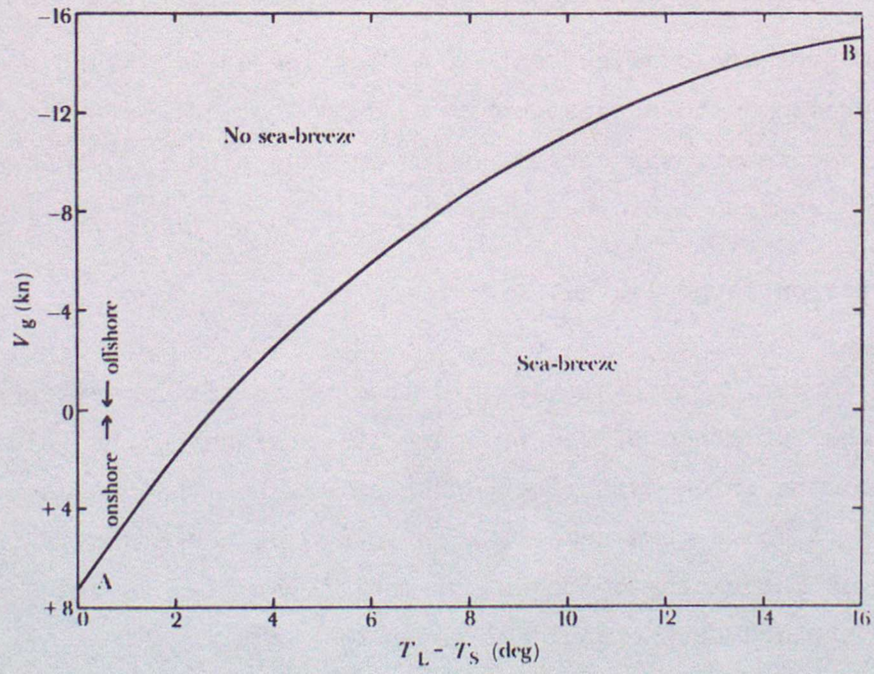


FIGURE 16. The relationship between $T_L - T_S$ and V_g for a sea-breeze to reach Manby

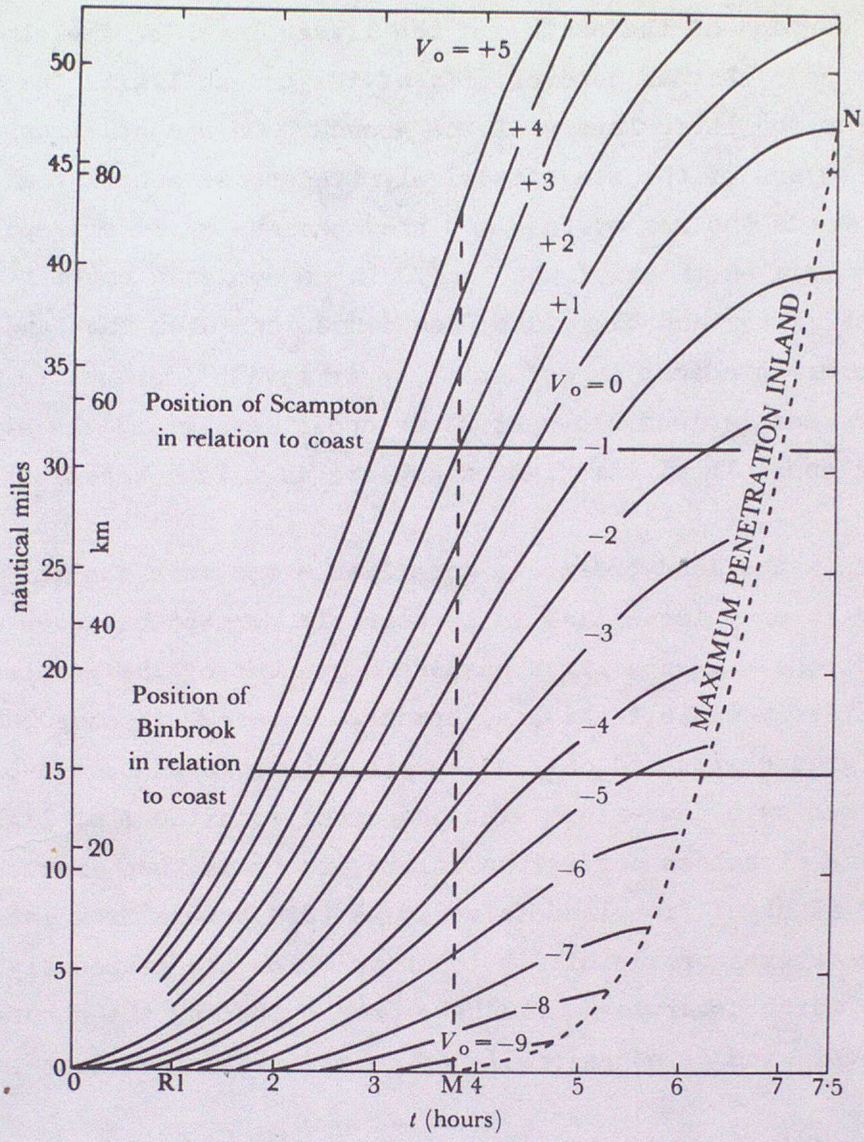


FIGURE 15. Inland penetration of the sea-breeze $y = V_0/f\,(\cos \mathrm{ft}^1 - \cos \mathrm{ft}) + v_p\,(t - t^1)$

Although Brittain's results may not be directly applicable elsewhere, a summary of the method has been included to show that relatively simple methods can often bring useful results; in a trial covering a period of six years (but using actual values of the predictors, V_g, T_L, T_S **and** v_p **,** where T_L and T_S are the land and sea temperatures), Brittain forecast the arrival or non-arrival of the sea-breeze at Binbrook (in the Lincolnshire Wolds) and Scampton (on the Lincoln Edge) correctly on 87 per cent of occasions.

16.6.2 **Land-breezes**

The land-breeze is the night-time counterpart of the sea-breeze and it blows when the land and the overlying air become colder than the sea and its

overlying air. On a clear night with light winds outgoing radiation soon cools the surface of the earth and the lowest layer of the atmosphere. The relatively poor thermal conductivity of the ground limits the upward flow of heat from the lower layers of the ground. On the other hand mixing in the upper layers of the sea effectively transports substantial quantities of heat towards the sea surface and prevents sea surface temperatures from falling as much as those of the land. In consequence the air over the sea tends to be warmer and therefore less dense than over the land. In this way a pressure gradient is set up at upper levels from sea to land, that is, a wind from sea to land blows at these upper levels and the other horizontal leg of the small local circulation appears as a land-breeze.

Generally the land-breeze is more feeble and more shallow than the sea-breeze and in many localities it is scarcely perceptible even under favourable conditions. Even a light onshore component of the gradient wind will often completely inhibit the land-breeze. However, in some coastal areas with high ground situated only a few kilometres inland, land-breezes may be reinforced by a down-slope wind and, under conditions of light pressure gradient and cloudless conditions inland, there is then frequently a noticeable wind at night from land to sea. As this breeze from land to sea may cause temperatures or visibility to occur which are noticeably different from those which otherwise would have been expected, the accurate forecasting of this local wind is occasionally of considerable importance.

Lawrence[42] examined the magnitude of land-breezes at anemometer height for a number of stations in England for the months of March, April and May. Figure 17, due to Lawrence, indicates the magnitude of the land-breeze. Directions are not plotted since these were normally from land to sea and at right angles to the isopleths but, where the latter wavered for small sea inlets, Lawrence found that the general direction was outward from the centre of the land mass. Where there is a major concavity in the coastline the land-breezes in the vicinity tend to be deflected towards the axis of the concavity.

Regarding the magnitude of land-breezes to seawards of the coastline, Lawrence suggests that 'the isopleths of the magnitude of the land-breeze over the sea are roughly a "mirror" of those over land'.

Land-breezes in the British Isles are normally very shallow and probably do not exceed 60 metres in depth.

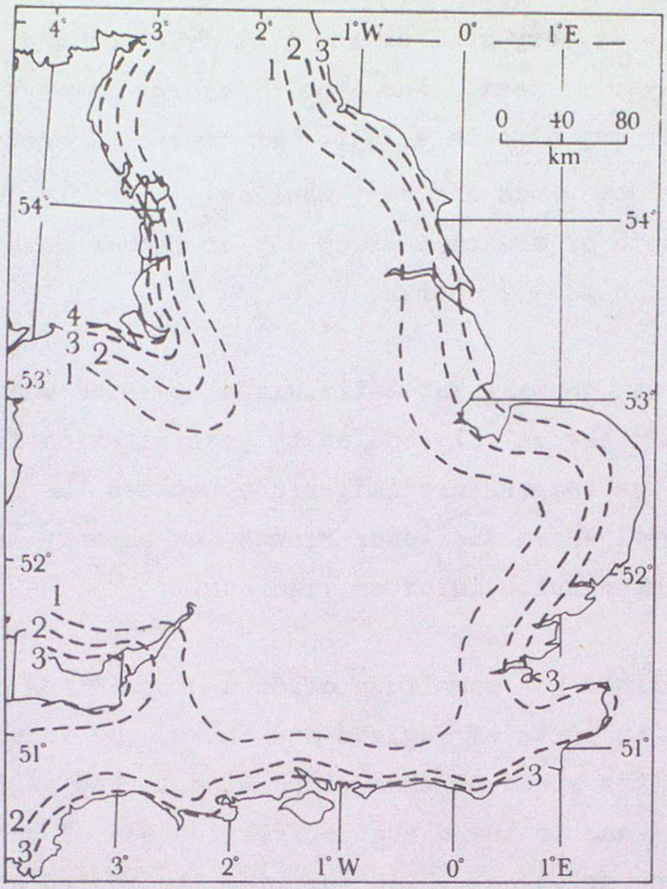


FIGURE 17. Lines of equal magnitude, in knots, of land-breeze (March, April, May)

There is a tendency for land-breezes near flat coasts to increase during the night. On general physical grounds it would be expected that the strength and frequency of the land-breeze would be greatest when the contrast between land and sea temperatures was at a maximum, namely, in anticyclonic conditions in early autumn.

16.6.3 Katabatic and anabatic winds

A katabatic wind occurs when air in contact with sloping ground is cooled and flows down the slope. Radiation nights, with clear skies and very slack pressure gradients, favour the formation of katabatic winds. Once katabatic flow has started, further cooling does not necessarily lead to an increase in its strength. Heywood[43] found that the katabatic flow did not increase throughout the night and that at times it reached a maximum quite early in the night. It should be noted, however, that very intense surface inversions inhibit down-slope winds and there is often a dead calm. Both Cornford[44] and Heywood[43] have commented that the katabatic winds they observed seemed to cease by the time a large inversion had built up. It may be that with such strong surface inversions the layer of air induced to move is very shallow and that friction and ground roughness then bring that layer to rest. The loss of energy involved in stirring a layer of stable air may also be a factor at times. In general, it seems likely that down-slope winds are very shallow - possibly not exceeding 30 metres. The depth of mountain winds may be rather greater and, on occasions, may extend to 150 metres.

Another mechanism causing katabatic winds operates when the upper slopes of a mountain are quickly cooled by precipitation falling upon them, by day or night. The temperature difference between the cooled air and air at the same level above the lower slopes can be very large and the katabatic flow which results is often vigorous.

Lawrence[42] examined the magnitude of down-slope winds on relatively broad, flat slopes in parts of England and Wales. He found that it was possible to relate the speed of this slope wind to the angle of the slope of the ground and to the distance from the sea of the site (sometimes called the 'donor' area) which supplied the cooled air to the slope. The relation between the smoothed values is shown by the curves in Figure 18. These curves satisfy the following formula:

$$\text{Speed of down-slope wind} = \sqrt{(2 \times \text{distance of donor site from the sea} / 1.4 + 100 \tan\theta)}$$

where θ is the angle of slope of the ground.

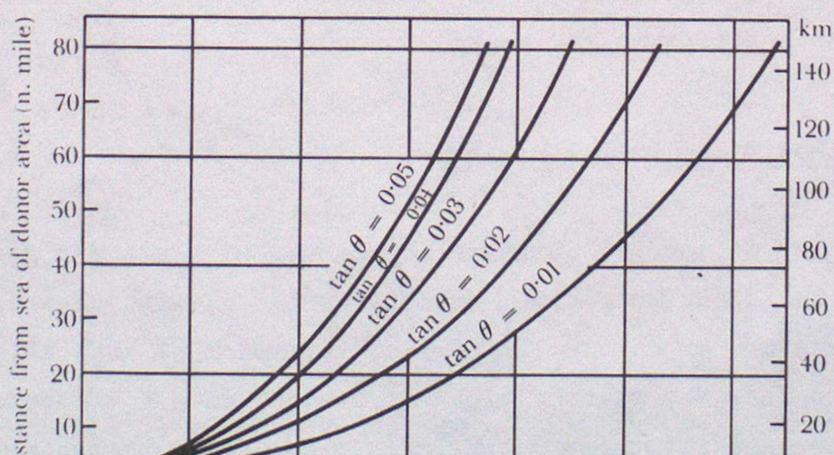


FIGURE 18. Relation of speed of down-slope wind (on a broad slope) to distance from sea and angle of slope of land

It should be noted that these general results were derived for broad, smooth slopes and cannot therefore be expected to apply to sites near hilltops, on valley floors or on banks of narrow valleys. Down-slope flow is strongly influenced by topography and, if the formula is used to calculate a down-slope wind for any particular site, the value so obtained should be used with caution and discretion.

In areas where a line of hills runs more or less parallel to the coast a katabatic wind may reinforce a land-breeze (see 16.6.2) so that a stronger flow from land to sea develops, the 'nocturnal wind'. Moffitt[45] has made a detailed study of the nocturnal wind at Thorney Island, near Portsmouth. To the north the South Downs rise to a height of nearly 200 metres or so within 13 kilometres of the station, lying almost parallel to the coast. Moffitt was able to produce a diagram relating the occurrence or non-occurrence of the nocturnal wind at Thorney Island to the difference between the day maximum and night minimum temperature and the strength of the gradient wind. The diagram is reproduced as Figure 19. For forecasting purposes, the night minimum temperature would have to be predicted (see section 17.6.2 of Chapter 17 - Temperature).

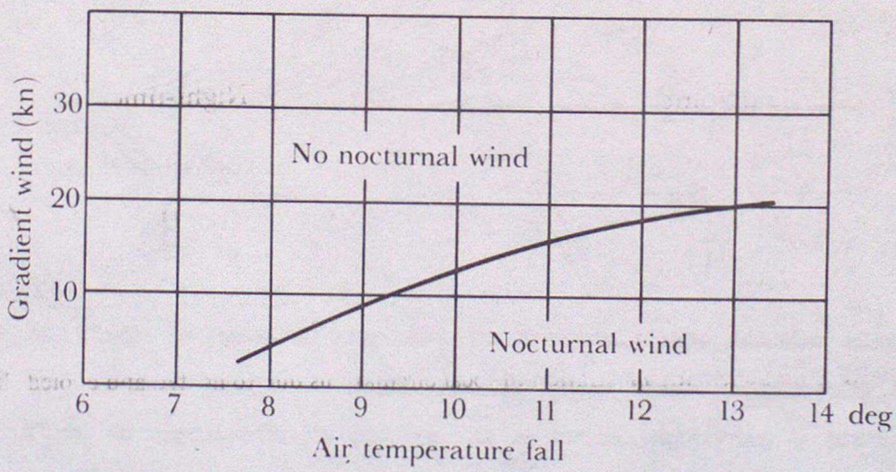


FIGURE 19. Graph of gradient wind and temperature fall at Thorney Island

A very persistent katabatic or 'nocturnal' type flow may occur when the ground is snow-covered, the nearly constant temperature difference between the snow surface and the sea causing a seaward thermal flow for most of the day. McGinnigle[46] has given an account of this type of occurrence at Acklington.

An anabatic wind occurs when a slope is heated by the sun; air in contact with the ground is warmed and becomes less dense than the air at the same level above the lower slopes, leading to upward motion along the slope. In the British Isles these anabatic winds are less frequently observed and are weaker than katabatic or nocturnal winds.

16.6.4 Mountain and valley winds

Although much of the British Isles consists of low-lying ground and gentle hills there are some areas of mountainous terrain. It appeared desirable therefore that a brief description of mountain and valley winds should be included in this handbook.

Consider first idealized long straight ridges rising uniformly from a level plain. Insolation on the slopes heats the air in contact with them amd this air is at a higher temperature than air over the plain at the same level. Air over the slopes rises in day-time and surface air flows from the plain towards the slopes and local circulations are thermally driven. These local circulations in a broad flat valley are illustrated schematically in Figure 20.

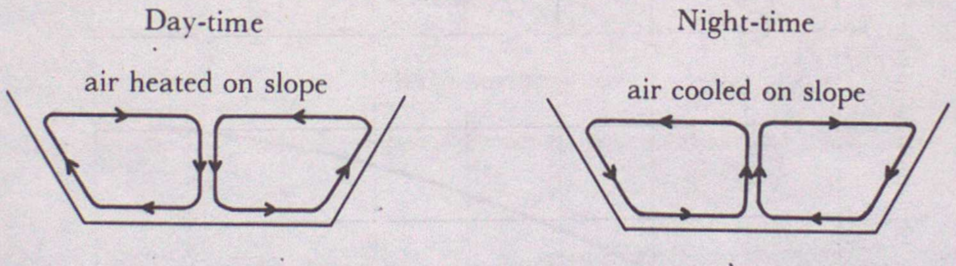


FIGURE 20. Schematic illustration of local circulations due to heated and cooled slopes

Southward-facing slopes receive more solar radiation than northward-facing slopes and it is on slopes with a southern aspect that this up-slope type of wind is usually most developed. The intensity of the up-slope wind varies greatly with differences in the local terrain while even a temporary shading of part of a slope from direct sunshine will cause an almost immediate variation in the local up-slope wind.

At night air in contact with the slope is cooled more rapidly than air over the plains at the same level and a local circulation which is the reverse of the day-time one is set up. This is also shown schematically in Figure 20. These circulations lead to ascending air over ridges by day and over valleys by night and descending currents over valleys by day and ridges by night. Clouds over ridges with clearances over valleys by day and conversely at night sometimes produce visual evidence of the upward and downward currents in these local circulations. The surface winds are, of course, well established by direct observations.

The up-slope wind in mountains normally commences shortly after sunrise, reaches its maximum intensity about the middle of the day and then diminishes. Shortly after sunset the down-slope wind sets in.

Consider now a wide and deep valley in mountainous country. Observations in the Alps have shown that in the day-time from about 0900 or 1000 until sunset a 'valley' wind blows up the valley towards the head and the mountains. By night an opposite wind - the mountain wind - blows down the valley and continues until after sunrise. These phenomena are also explained by the different behaviour to heating and cooling of the air in the valley and near the mountains and of that over neighbouring lower-lying country - the plains. The diurnal variations of temperature in the valley air are greater than those at corresponding levels above the plains. As a consequence a local pressure gradient from plain to valley is set up by day and from valley to plain by night. These pressure gradients cause the 'valley' wind to flow up the valley by day and the 'mountain' wind to flow down the valley by night.

Thus in mountainous or hilly country there are two local winds of thermal origin to be considered, namely, the up-or-down-slope wind and the up- or down-valley wind. The development of these systems is unlikely to be in 'phase' and the pattern of these winds throughout a period of 24 hours is one of some complexity. Naturally these winds are rather more strongly developed in more hilly country than occurs in the British Isles. Many studies have been made of these winds in the Alps. Defant[47] has produced a series of schematic diagrams illustrating both these types of winds. These are reproduced in Figure 21.

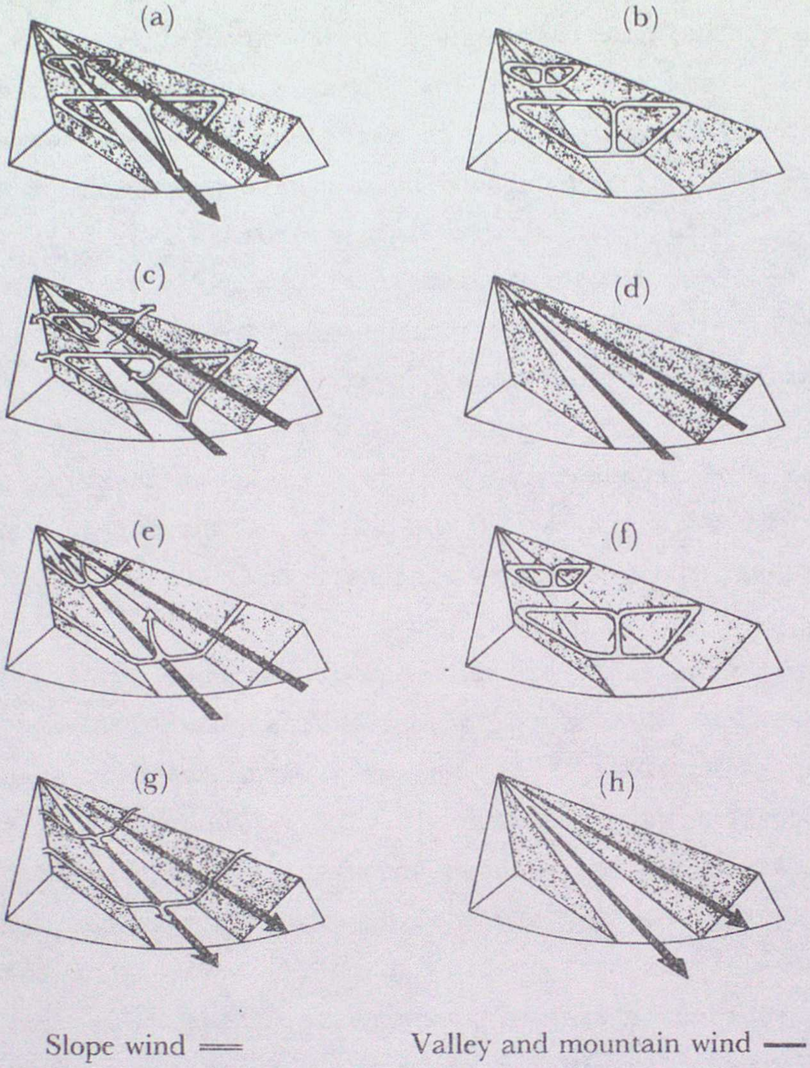


FIGURE 21. Schematic illustration of the normal diurnal variations of the air currents in a valley (after Defant[47])

(a) Sunrise – up-slope winds commence, mountain wind continues to blow down valley; pressure gradient is down the valley; temperatures: valley cold, plains relatively warm.

(b) Forenoon (about 0900) – strong up-slope winds, transition from mountain to valley wind; nil pressure gradient in direction of valley; valley temperature same as plains; strong warming in the valley, little change of temperature above the plains.

(c) Noon and early afternoon – diminishing up-slope winds, valley wind well developed; pressure gradient is up the valley; temperatures: valley warm, plains relatively cold.

(d) Late afternoon – up-slope winds have ceased, valley wind continues; pressure gradient is up the valley; temperatures: valley still warm and plains relatively cold; slow cooling in the valley.

(e) Evening – down-slope winds commence, valley wind diminishes; pressure gradient is still up the valley; temperatures: valley slightly warmer than above the plains; strong cooling in the valley, only slight cooling in the plains.

(f) Early night – down-slope winds well developed, transition from valley to mountain wind; nil pressure gradient in direction of valley; valley temperature same as plains; strong cooling in the valley continues.

(g) Middle of the night – down-slope winds continue, mountain wind well developed; pressure gradient is down the valley; temperatures: valley cold, plains relatively warm; cooling in the valley, slight cooling above the plains.

(h) Late night to morning – down-slope winds have ceased, mountain wind throughout the valley; pressure gradient is down the valley; temperatures: cold in the valley, warmer above the plains; changes slight in the valley and above the plains.

16.7 GUSTS AND SQUALLS

16.7.1 Gusts

Wind seldom occurs as a smooth-flowing airstream but as a stream continually fluctuating in speed and direction. The fluctuations in the stream are due to eddies which vary both from place to place and from time to time at the same place. Inspection of pressure-tube anemograms or a careful examination of the records of some cup-type anemometers yields some quantitative estimates of the fluctuations. It is readily apparent from these records that it is not possible to forecast the effects of any individual eddy on the mean flow. They are nearly all far too transient - a fact readily borne out by common experience of the surface winds when out of doors. The approach to either an examination or the forecasting of gustiness must be a statistical one.

If a single eddy, embedded in the general wind, flows past an anemometer there will be an increase in speed when the eddy flow reinforces the general flow and a decrease when the eddy flow detracts from the general flow. If eddies are regarded as entities with rotating winds, then it is clear that the passage of eddies will be associated with changes in wind direction as well as speed. If the eddies which move across an anemometer were of a similar type and size a form of periodicity in the gusts and lulls would be apparent on the anemogram. The very detailed measurements and the open time-scale of the recording instruments used in the Cardington investigations reported by Giblett et alii [17] showed, however, that gusts were very heterogeneous. The time interval between gust and ensuing lull was very varied. Giblett et alii found that in spite of this heterogeneity in gusts and lulls, the pattern of any one gust and lull showed the following general characteristics. The increase in wind at the onset of the gust was very rapid - sometimes the maximum wind in the gust occurred one second or so after its initial onset. The subsequent decay from maximum gust wind to lull occurred much more slowly (but not uniformly) over periods of perhaps 10 to 30 seconds. These characteristics of rapid increase and relatively slow decrease were noticeable on all scales of gusts. During the gust some change in wind direction also occurred but these changes were not uniform about the mean wind direction. On the whole the wind direction in the gust was likely to be veered from the preceding wind direction but this was not always the case. On any one day gust followed gust with variations in the intensity of the gust, the duration to the lull and the interval to the next gust. In fact it was clear that the gusts were often due to eddies of markedly different size or character. On occasions when deep convection was occurring in the troposphere and cumulonimbus clouds were present 'high gusts were spaced out at comparatively wide intervals but the whole trace was irregular both in direction and velocity'. With less vigorous convection occurring in the lower layers of the troposphere the major fluctuations in wind speed and direction were still considerable but were more rapid. The lapse rate when this type of fluctuation occurred was approximately adiabatic but when the vertical temperature gradient near the ground was less than the adiabatic 'the trace on the anemograms was broad both in velocity and direction but the fluctuations appeared to be very rapid more so than with either of the two previous types'. When there was a decided inversion in the lowest layer of the atmosphere (the lowest 60 metres or so) fluctuations almost entirely disappeared from the trace.

The longer-period gust and lull described above may be ascribed to convectional currents set up in an unstable atmosphere. The shorter-period disturbances which occur when the lapse rate of temperature near the ground is stable but is not a strong inversion are eddies caused by frictional effects.

As stated earlier it is not possible to forecast individual gusts but it is possible to indicate the probable value of the maximum gust which may occur. Typical values of the ratio of the maximum gust speed to the mean hourly wind speed are shown in Table 16.11[17] (on the facing page).

Gold[48] determined the distribution of ratios of maximum gust to the maximum hourly wind for some 42 stations in the British Isles. The frequencies are shown in Table 16.12.

TABLE 16.12 Frequency of various values of the ratio of maximum gust to maximum hourly wind for 42 stations in the British Isles

Ratio	*1.3*	*1.4*	*1.5*	*1.6*	*1.7*	*1.8*	*1.9*	*2.0*	*2.1*	*2.6*
Number of occasions	1	9	10	6	5	3	4	1	2	1

TABLE 16.11 Frequency of the ratios of maximum gust to mean wind at Cardington

Time GMT	*Wind speed at 45m*	*1.2*	*1.3*	*1.4*	*1.5*	*1.6*	*1.7*	*1.8*	*1.9*	*2.0*	*2.1*	*2.2*	*2.3*
	knots					*Height 15m – June 1929*							
0100	(9–16	2	4	8	13	11	6	1	1	–	–	–	–
to	(17–25	–	–	2	1	–	–	1	–	–	–	–	–
0300													
1300	(9–16	–	2	1	7	16	7	12	6	2	2	–	–
to	(17–25	–	–	1	9	6	2	–	–	–	–	–	–
1500	(26–34	–	–	1	1	1	–	–	–	–	–	–	–
						Height 15m – December 1929							
	(< 9	1	–	2	–	2	–	–	1	1	–	–	–
0100	(9–16	–	3	4	5	2	4	1	–	1	–	–	–
to	(17–25	–	–	7	14	2	1	1	–	–	–	–	–
0300	(26–34	–	–	2	6	9	1	–	–	–	–	–	–
	(35–42	–	–	–	1	–	–	–	–	–	–	–	–
	(43–51	–	–	1	2	–	–	–	–	–	–	–	–
	(9–16	1	1	8	10	11	1	3	1	–	–	–	1
	(17–25	–	–	2	7	8	2	–	1	–	–	–	–
1300	(26–34	–	–	2	6	2	1	1	–	–	–	–	2
to	(35–42	–	–	–	1	1	3	–	–	–	–	–	–
1500	(43–51	–	–	–	1	1	–	–	–	–	–	–	–

For coastal stations the ratios were mainly between 1.4 and 1.5, for inland country stations between 1.6 and 1.8, but for inland town exposures the ratios were mainly between 1.9 and 2.1.

Shellard,[49] in a detailed study of the variation of maximum gust with mean hourly wind speed, obtained very similar results, as shown in Table 16.13.

TABLE 16.13 Ratio of maximum gust to mean hourly speed for open sea and various terrains

	Range of ratios found	*Estimated average ratio*
Open sea	1.3	1.3
On isolated hill tops well above general level	1.4–1.5	1.4
Flat open country, fens, etc.	1.4–1.8	1.6
Rolling country with few windbreaks such as trees or houses, e.g. farmland	1.5–2.0	1.7
Rolling country with numerous windbreaks, forest areas, towns and outskirts of large cities	1.7–2.1	1.9
Centres of large cities	1.9–2.3	2.1

Notes

(a) Effects in valleys have not been considered here, and these values refer to the ratio in strong winds in the free air.

(b) Conditions in a town or city in the layer between rooftop and street level are usually determined by the local configuration of streets and buildings.

The maximum gust speed may also be related to the geostrophic wind speed. For strong geostrophic winds ($\geqslant$ 35 knots) the geostrophic speed gives a close estimate of the maximum gust speed. When geostrophic winds are light, however, the maximum gust may be greater than the geostrophic speed, especially on spring and summer days, mainly as a result of convection. Some further remarks on this topic are given in Chapter 23 - Bumpiness in aircraft.

16.7.2 Squalls

Gusts are essentially very short-lived and transient in character but occasionally a strong wind rises suddenly, lasts for a time which can be measured usually in minutes, and then dies away, often comparatively suddenly. This is known as a squall. It is usually associated with a temporary change in wind direction but, when a squall is associated with a front, the change in direction may be of a semi-permanent nature, that is, a feature of the flow of the post-frontal air mass.

Squalls are relatively infrequent phenomena in the British Isles. However, the sudden increase in wind speed and the shift in direction render them dangerous and, when they are violent, much damage may be caused. In this country they are often associated with deep convection cloud and heavy precipitation and are potentially dangerous to aviation.

When squalls are associated with features (for example, fronts or troughs) which can be recognized on synoptic charts, the normal processes of analysis and prognosis will go a long way towards providing a forecast of the probable movement of a 'squall line' and values of wind speed observed at upwind stations will give some guide to probable wind speed. However, some squalls are associated with a smaller individual storm; the existence of the storm and the precise location of such storms and their future movement are very difficult to observe and predict on normal synoptic charts. Radar weather echoes may give a very useful guide for the short term (see Chapter 19 - Clouds and precipitation).

The air reaching the surface in the squall has probably descended from upper levels. Air at upper levels is cooled by evaporation from precipitation and subsequently warmed by compression during descent to the surface where, in spite of the warming, the air in the squall usually arrives as a relatively cold current. There are very few rules which enable a forecast of the speed of the wind in the squall to be made but it seems likely that the temperatures of the air immediately preceding and during the squall may be important parameters. Further if a parcel of air does descend to the surface the wind speed and direction at levels of 3000 to 5000 metres may also be important since the descending air will initially have the momentum of the wind at the upper level.

In a well-marked frontal or line squall the wind direction before and after the squall can usually be forecast adequately by normal methods but the directions during the squall are very difficult to forecast. Wind usually veers in a squall but not invariably so and at times directions may be temporarily almost reversed. In isolated squalls the variations of wind direction are difficult to forecast except perhaps in localities where topography may channel the winds. The upper wind around the levels from which the chilled air descends may also provide a clue. Before and after such squalls the winds usually conform to the general isobaric pattern although a very large storm centre may maintain cross-isobaric surface winds for a few hours - particularly when it is slow moving and in a slack pressure distribution.

Fawbush and Miller[50] have made a study of peak wind speeds experienced when non-frontal thunderstorms passed over a number of stations in the United States. They found a high correlation (0.86) between the peak wind speed and the difference between the surface temperatures just prior to and immediately following the first heavy rain shower. This relationship is shown graphically in Figure 22.

Fawbush and Miller further stated that practical experience indicated that the temperature of strong down-draughts reaching the surface in thunderstorms was very close to the surface temperature shown by the wet adiabat through the intersection of the $0^{o}C$ isotherm and the wet-bulb curve of a representative tephigram. This is illustrated in Figure 23. (Note: The wet-bulb temperature is readily obtained on a tephigram from dry-bulb and dew-point temperatures by the method described in Chapter 10.) Thus a forecast of dry-bulb temperature and an estimate of the 'downrush' temperature, together with Figure 22, yield an estimate of the peak wind

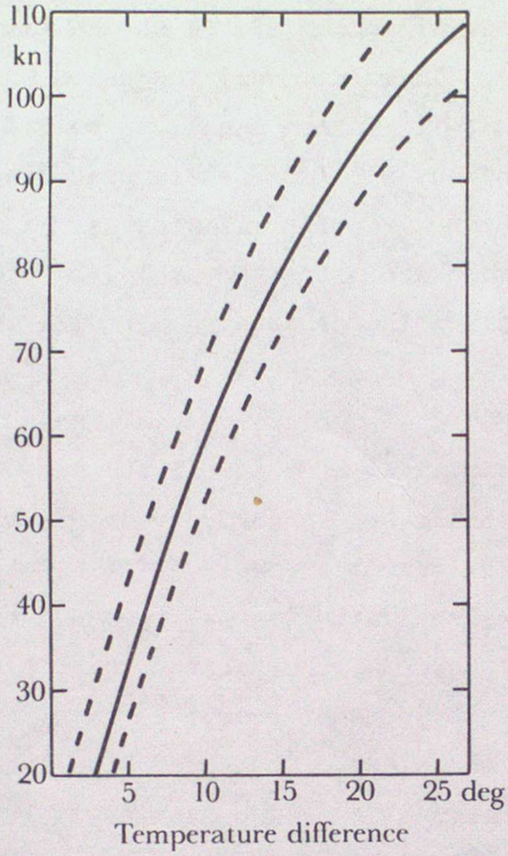


FIGURE 22. Peak wind speed and temperature differences in thunderstorms in the United States of America

—— denotes regression curve

– – denotes standard error of estimate

speed (with a standard error of estimate of 8 knots) likely to occur in non-frontal thunderstorms in the United States.

McNair and Barthram,[51] investigating a squall line over England, found that the technique gave a good estimate of the maximum gust occurring during the squall.

16.8 FORECASTING THE WIND IN THE FRICTION LAYER

The structure of the boundary layer was briefly discussed in 16.1 (page 3). The depth of the boundary layer is highly variable but, for the purposes of this section, will be taken as the lowest kilometre. Interest will mostly be concentrated in the lowest 100-200 metres or so, partly because it is within the lowest layer that the largest variations normally occur, and partly because these regions are the least inaccessible to measurement.

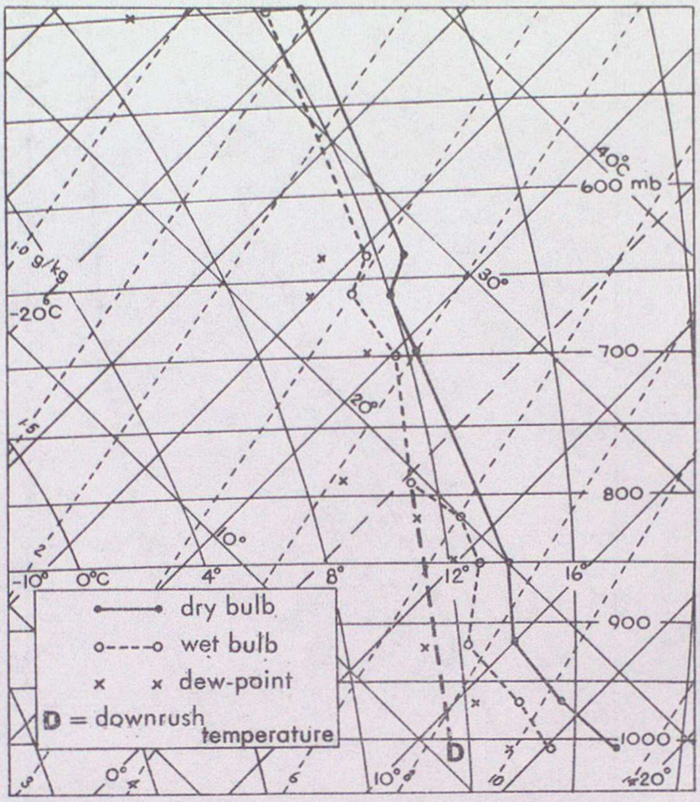


FIGURE 23. Example of tephigram to illustrate the method recommended by Fawbush and Miller[50] for computing the 'downrush' temperature in non-frontal thunderstorms in the United States of America

Even so, there is not a great deal of information to help the forecaster, for in micro-meteorological studies every effort is made to obtain the ideal conditions for measurement - statistically steady, homogeneous, flow with a long fetch over uniform smooth terrain, etc. The forecaster often has to deal with non-uniform conditions and rough terrain, but nevertheless the following sections should contain some useful background information, and some indication of how to forecast the variations of wind within at least the lowest part of the friction layer.

16.8.1 Vertical wind profiles in lapse and neutral conditions

Figure 24 is reproduced from a paper by Pasquill:[52] it shows the results of an analysis by Lettau of part of a classical series of pilot-balloon ascents at Leipzig. The profile is the mean of ascents made over a period of 7 hours. The main features are the rapid decrease of $\partial V/\partial z$ with height, and the nearly linear variation with height of the departure of the wind direction from that of the geostrophic wind.

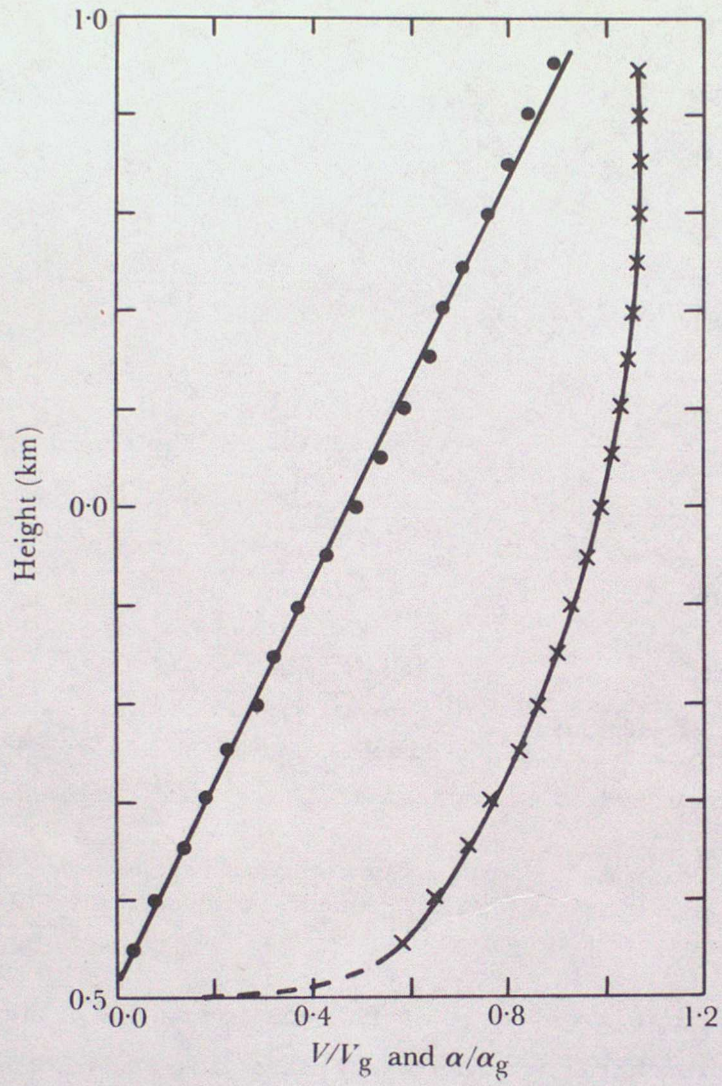


FIGURE 24. General nature of wind profile in near-neutral conditions (from re-analysis by Lettau (1950) of Leipzig pilot-balloon data)

×—× V/V_g; •—• α/α_g
V_g = geostrophic wind = 17·5 m/s
α = angle between surface wind and V
α_g = angle between surface wind and V_g = 26°

In neutral conditions, with flow over fairly smooth and uniform terrain, the mean wind speed, $\overline{u}$, and height, z , can be represented by the formula

$$k\overline{u}/u_* = \ln(z/z_0) \quad . \quad . \quad . \quad . \quad (16.12)$$

where k is a constant (von Kármán's constant, ≈ 0.35 to 0.4), u_* is the friction velocity and z_0 is the roughness length. (For the definition of these quantities and for the physical meaning and reasoning behind the

derivation of the equation, the reader should consult Pasquill,[52] Smith and Carson,[53] (or any text on micrometeorology). If z_0 is known for a given site (and wind direction, if the terrain varies with direction from the site), u_* can be found from a wind observation at a single height, and the equation then used to determine the wind at other heights. Unfortunately, the equation applies only in neutral conditions and for heights up to a few tens of metres, and there exists no straightforward formulation of the wind profile throughout the whole of the boundary layer. However, by applying similarity-theory arguments to the problem (see Sutton[54], pages 99 ff), Smith and Carson[53] have devised a scheme whereby day-time wind-speed profiles can be deduced for given months and times of day, depending upon cloud amount and the surface geostrophic wind and/or the actual surface wind (at 10 metres). The following account is based on their paper.

The parameters needed to specify the wind profile are:

(a) z_0, the surface roughness. The effective z_0 generally varies from one part of the profile to another. For the near-surface profile the very local nature of the surface (e.g. short grass) is important, but away from the ground z_0 represents an average roughness over a greater upwind area which is significantly affecting the wind profile at the height concerned. For a well-exposed anemometer at a rural site the effective z_0 will increase with height as there are likely to be larger roughness elements (trees, hedges, buildings, etc.) at a distance upwind than in the immediate vicinity of the site. For such a rural area it is assumed that z_0 = 10 cm for the 10-metre wind, increasing to 30 centimetres above 100 metres. For an urban profile, applicable only at heights well above the heights of the roughness elements themsleves, z_0 = 1 metre from about 20 metres, decreasing to 30 centimetres for heights above 100 metres where the urban environment will no longer significantly affect the profile unless the city extends for more than a few kilometres upwind or there are many tall buildings.

(b) Some measure of the degree of instability in the surface layer.

(c) The total incoming short-wave radiation, I.

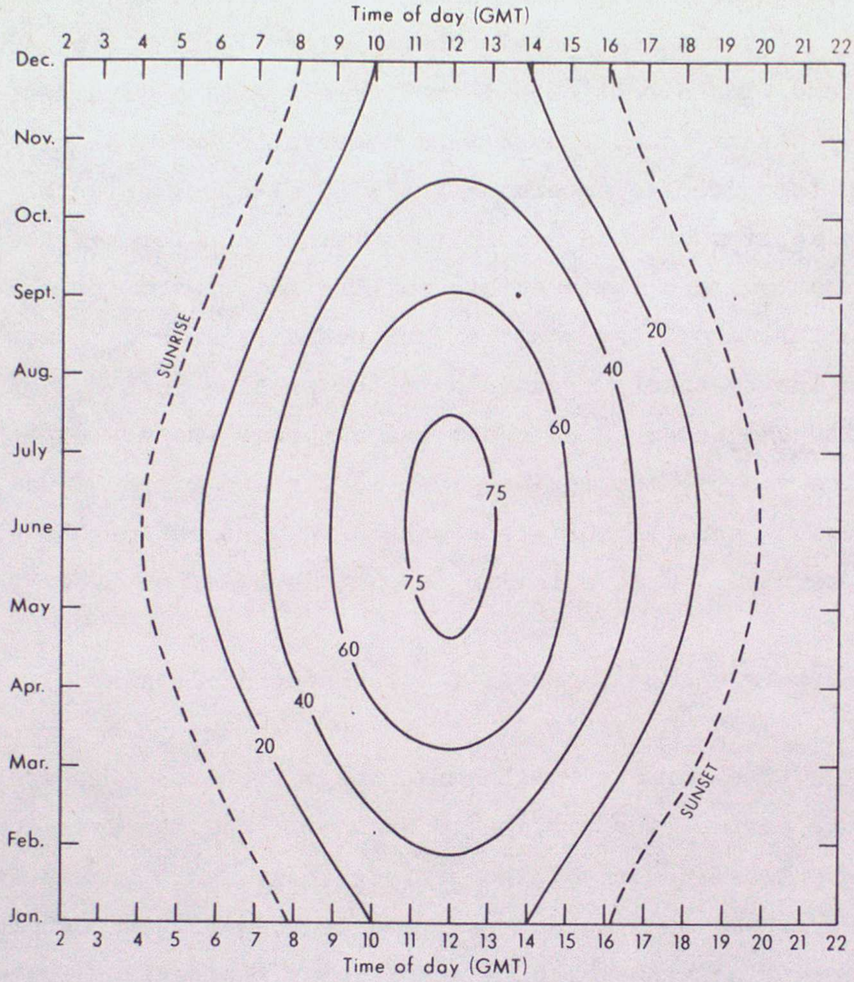


FIGURE 25. Variation of incoming solar radiation, *I*, with time of year and time of day for Cardington, in milliwatts per square centimetre for 0–1 okta of cloud

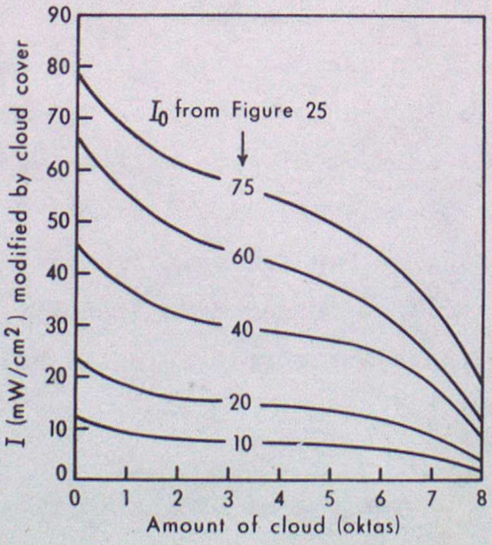


FIGURE 26. Incoming solar radiation modified by cloud cover

(d) The depth, h, of the atmospheric boundary layer.

(e) The value of the 10-metre wind speed, u_{10} (strictly for $z_0 =$ 10 centimetres only, i.e. in a rural area), or the surface geostrophic or gradient wind, V_g.

Smith and Carson show that the wind speed at height z can be represented simply as

$$\frac{u_z}{u_{10}} = f(z, z_0, V_g/u_{10}, h) \qquad \text{. . . . 16.13}$$

where V_g/u_{10} has been found experimentally to be an adequate representation of the stability parameter mentioned in (a) above. The function f in equation 16.13 can be expressed in graphical form, and thus u_z can be found quite quickly when the other parameters are known.

The first stage is to determine the incoming solar (short-wave) radiation, I. Figure 25 shows the variation of I with month and time of day for 0-1 okta of cloud; Figure 26 shows the allowance to be made when more cloud is present. The ratio V_g/u_{10} is then read off from Figure 27(a) if u_{10} is known, or from Figure 27(b) if V_g is known.

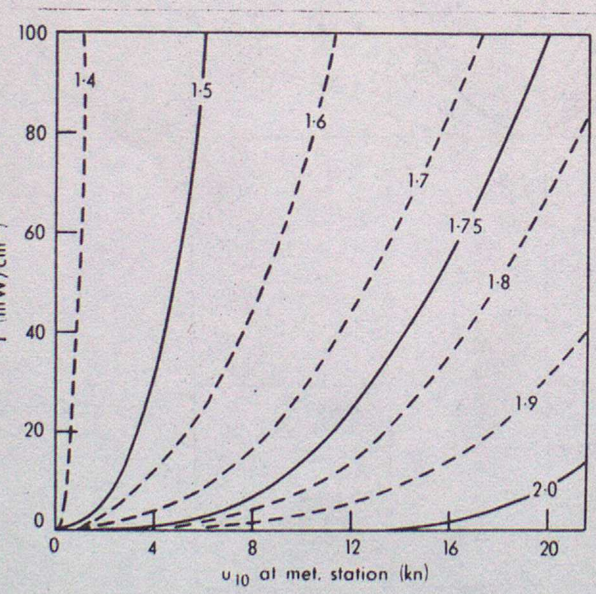


FIGURE 27(a). Ratio of geostrophic wind, V_g, to the wind at 10 metres, u_{10}, in relation to u_{10} and the incoming solar radiation

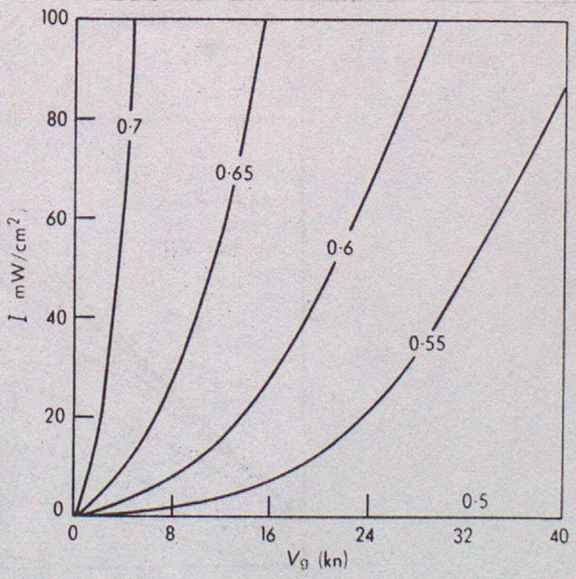


FIGURE 27(b). Ratio of the wind at 10 metres, u_{10}, to the geostrophic wind, V_g, in relation to V_g and the incoming solar radiation

If both u_{10} and V_g are known, this step may be omitted, or used as a check. In situations where topographical effects are likely to be significant, or in urban areas, any measured values of u_{10} may not be representative, and V_g above must be used, as indeed it also must when forecasting for several hours ahead. If the curvature of the air trajectories is significant, the gradient wind may have to be used instead of the geostrophic wind.

The estimation of the depth, h, of the boundary layer comes next, and this is done by means of Figure 28, based on the developing boundary-layer model of Carson.[55] The method of use is straightforward and is illustrated for the example quoted by the dashed lines and arrows. If there is a subsidence inversion based at a height lower than the derived h, the lower value is taken as the depth of the boundary layer. Over an urban area, u_{10} will have to be derived from V_g and the value of V_g/u_{10} obtained from Figure 27(b).

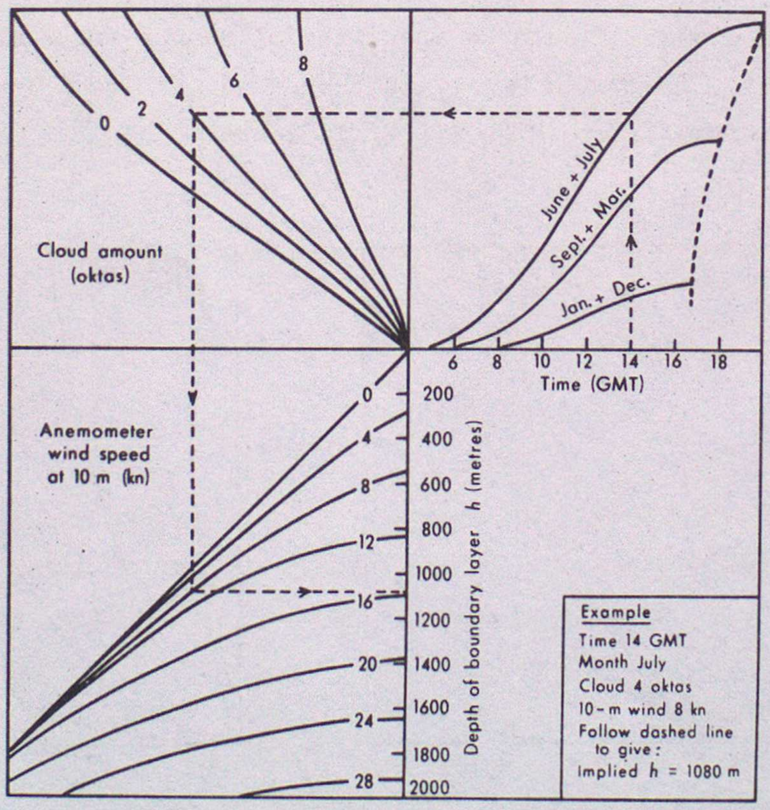


FIGURE 28. Nomogram for determining the depth of the day-time boundary layer in relation to time of day, month, cloud amount and the wind speed at 10 metres

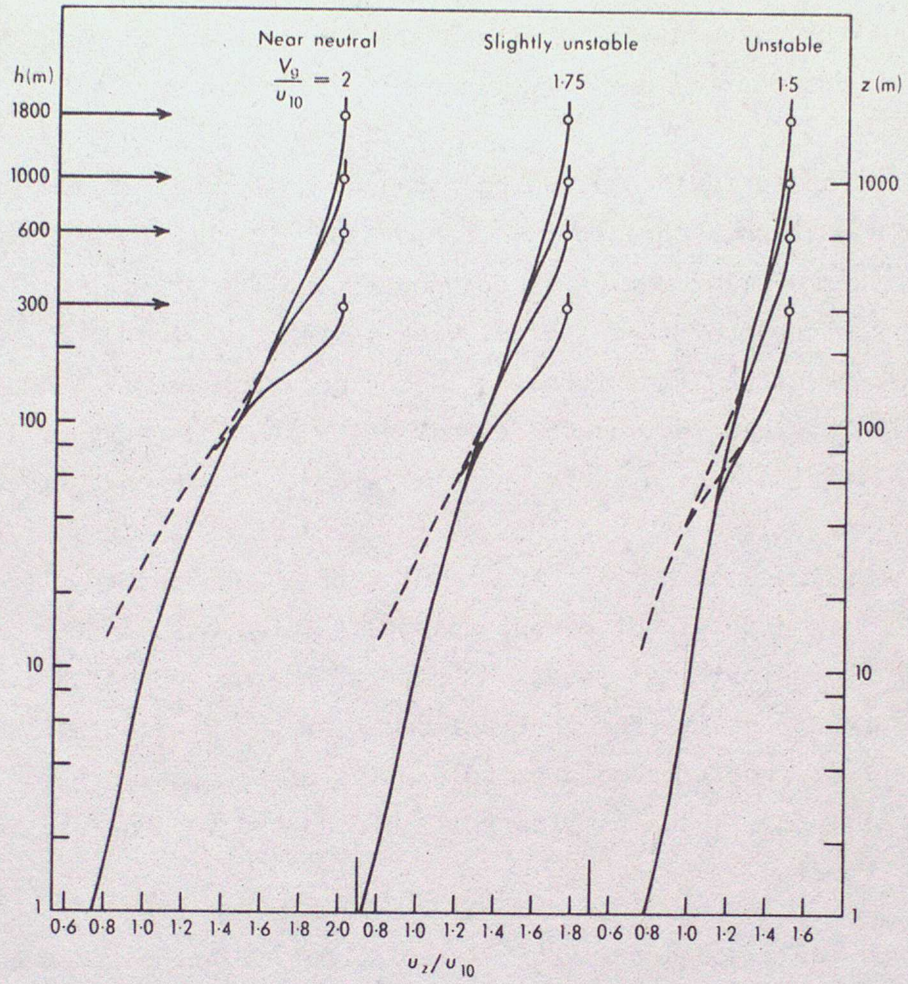


FIGURE 29. Wind profiles for near neutral, slightly unstable and unstable conditions for various depths of boundary layer

Below 100 metres --- denotes urban profiles and —— denotes the profiles at meteorological stations.

The last stage is the use of Figure 29 to derive the wind profile, or the wind at a specified height. In the diagram, specimen profiles for three values of V_g/u_{10} are given, appropriate to near-neutral stability, slight instability and unstable conditions respectively. Above 100 metres, four profiles are given, for values of h indicated by the arrows and by open circles at the upper end of each profile. Some interpolation may be necessary, but often it will be sufficient to take the nearest curve. Below 100 metres or so, two profiles are given, one for a rural site, the other for an urban site. The profile for an urban site should not be continued below about twice the height of the buildings; below this height winds are very variable and are strongly influenced by individual buildings or groups of buildings. Usually (linear) interpolation between the profiles for the given values of V_g/u_{10} will be necessary, and for urban areas a fictitious value u_{10} must be calculated, but these are the only two arithmetical manipulations not carried out directly on the diagrams.

The method as so far described has assumed uniformity in the horizontal; if there is a horizontal gradient of temperature in the lowest layers a significant thermal wind may exist between the bottom and top of the boundary layer. The component of this thermal wind perpendicular to the general wind direction alters mainly the direction of the geostrophic wind and has comparatively little effect on the speed, while the component of thermal wind parallel to the flow affects only the speed. Smith and Carson have shown how to make allowance for the thermal wind in a fairly simple way, but it has not been included here as it is not certain whether the forecaster on the bench will need to use it often, and those interested in the details should refer to the original paper.[53] Briefly, the horizontal temperature gradient is used to derive the vertical shear of the geostrophic wind. The vertical shear is then extrapolated to the top of the boundary layer to give a new value of V_g/u_{10}, which may lie well outside the values used in the profiles of Figure 29.

The method has a sound theoretical basis, but it has not yet been tested extensively in practice. It would be a useful exercise to test the technique at an outstation if suitable winds can be obtained for verification purposes.

16.8.2 Vertical wind profiles in stable conditions

When the atmosphere is stable, vertical eddy motion is damped and the transport of momentum from the free atmosphere to near ground level is reduced, particularly when winds are not strong. The result is that the wind speed near the ground decreases while that in the free atmosphere may increase. Events may be complicated, however, by the decay of the turbulent boundary layer at night and by its re-formation, making it difficult to say where the 'free atmosphere' begins. In stable conditions, wind profiles are often extremely difficult to predict, and they may be very variable in both space and time, depending on the temperature structure of the air and on topography, etc. An example of the change in wind profile with change of atmospheric stability is given in Figure 30, showing mean profiles in spring at Cardington over a 10-year period for gradient winds of 170-250°/15-30 knots for the lapse-rate classes of Figure 7 (page 23). The variation among individual profiles would probably be considerably greater.

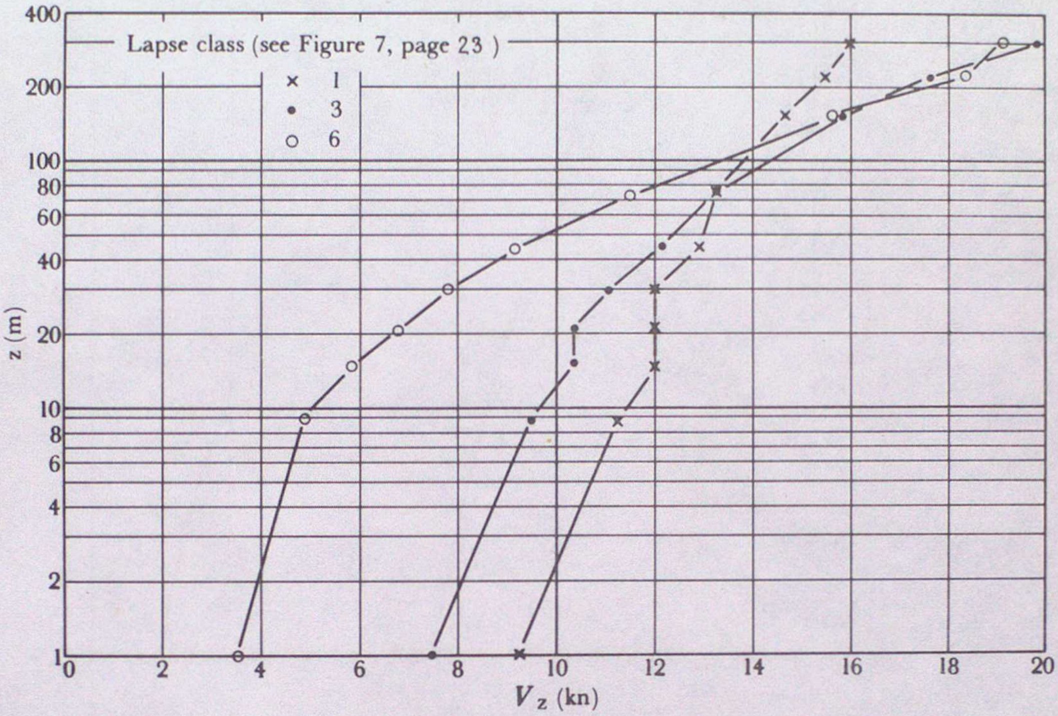


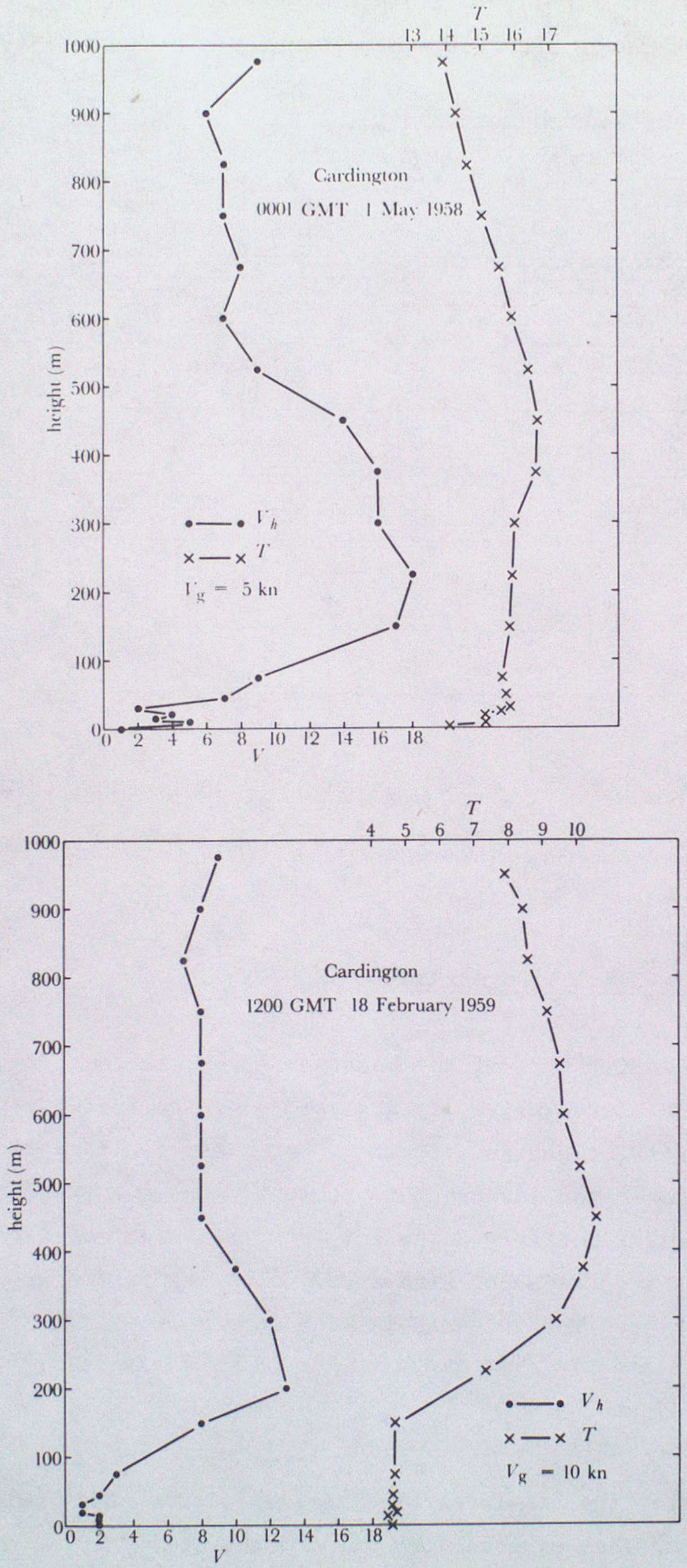
FIGURE 30. Mean wind profiles at Cardington for various lapse classes

Spring: geostrophic wind speed (Vg) = 15–30 kn, direction = 170°–250°.

16.8.3 The low-level wind maximum

Usually the upper part of the boundary layer, above a few hundred metres or so, is characterized by a marked decrease in the rate of increase of wind speed with height (except when, for example, there is a strong thermal wind). When an inversion is present, the wind speed may reach a maximum at a height not very different from the height of the top of the inversion, with a decrease of wind speed occurring in the upper part of the boundary layer. The tethered-balloon soundings (BALTHUM) at Cardington often show this feature, and two examples of typical profiles are shown in Figures 31(a) and (b).

At Cardington the low-level wind maximum occurs about twice as frequently at night (10 per cent of occasions at 0001 and 0600 GMT) as during the day. (In this comparison the criterion has been that the maximum speed should be



FIGURES 31(a) and (b). Wind and temperature profiles at Cardington on two occasions of a low-level wind maximum

10 knots or more, and that the speed should fall to half of the maximum value or less in the 300 metres above the height of the maximum speed.) Although the low-level maximum is associated with gradient winds from all directions, the maximum frequency occurs with winds from between west-south-west and north. The horizontal extent of the phenomenon and how it varies from place to place are not known, but when it occurs at Cardington it is often detected by pilot-balloon soundings at Odiham (Hampshire), 100 kilometres away.

Low-level wind-speed maxima occur most frequently at night when the circumstances attending the formation of a nocturnal inversion in the lowest 100 to 300 metres tend to promote or intensify their development. They also occur during the day in stable conditions, with about half the frequency of the night-time occurrences. The height of the speed maximum is usually, but not always, a little below the top of the inversion. However, a significant wind-speed maximum does not occur often enough in apparently suitable conditions for it to be forecast with confidence. When a knowledge of the wind profile in the boundary layer is of importance the only safe guide is a measurement of the wind speeds at the required heights, by tethered-balloon soundings, by observations from high masts or by pilot-balloon soundings; winds observed by radar are average values over a layer several hundred metres deep, obscuring any low-level maximum which may be present.

Low-level wind maxima occur in other parts of the world, and in particular the southerly low-level jet which occurs to the east of the Rocky Mountains in the U.S.A. has been intensively studied (see, for example, Hoecker [56]). It seems, however, to bear little resemblance in magnitude, scale or associated synoptic situation with its British counterpart, although both may be at least partly a result of an inertial osciallation occurring when the 'braking' effect of surface drag is removed as the lowest layers stabilize (see Petterssen,[57] Blackadar [58]).

Low-level jets have also been observed[59] ahead of active cold fronts, and are discussed in Chapter 6 - Depressions and related features. They are essentially geostrophic phenomena, or nearly so, and differ in this respect from the low-level wind maxima described in this section.

16.8.4 The power 'law'

The simplest way of expressing the results of studies of the variation of wind speed with height is the so-called 'power law', in which

$$V_h = kh^p \quad . \quad . \quad . \quad . \qquad (16.14)$$

where V_h is the wind speed at height h, k is a constant and p depends upon the roughness of the terrain and upon the lapse rate. Individual profiles may deviate a good deal from those derived from equation 16.14, which perhaps finds its most valuable application in climatological studies rather than in forecasting, but the power law often does give a worthwhile indication of how the wind speed varies with height. Figure 32, based on work by Shellard,[60] shows the variation of V_h/V_{10} with h for various values of p for neutral conditions (adiabatic lapse rate) and fairly strong winds; the appropriate values of p for various types of terrain are shown in Table 16.14.

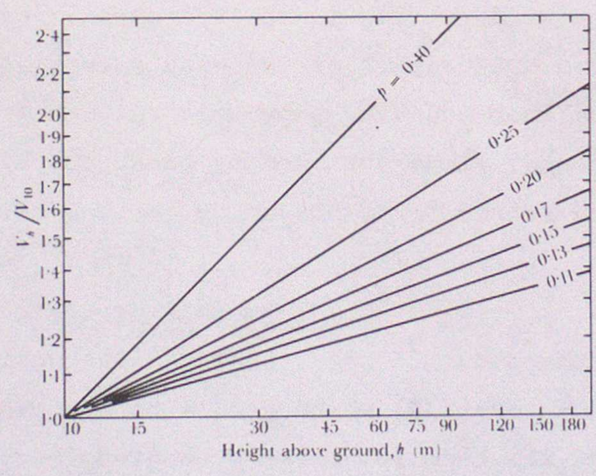


FIGURE 32. Ratio of wind speed at height $h(V_h)$ to that at 10 metres (V_{10}) for adiabatic conditions

TABLE 16.14 Values of p for various surfaces

Type of surface	*Approximate value for p*
Very smooth – marshlands, flat coasts	0.11 to 0.13
Open country with only isolated trees and low hedges	0.15 to 0.17
Well-wooded farmland and suburban areas	About 0.20
Cities	May exceed 0.25 though few reliable measurements are available. Some workers have suggested values of 0.4 or more but these results are open to criticism.

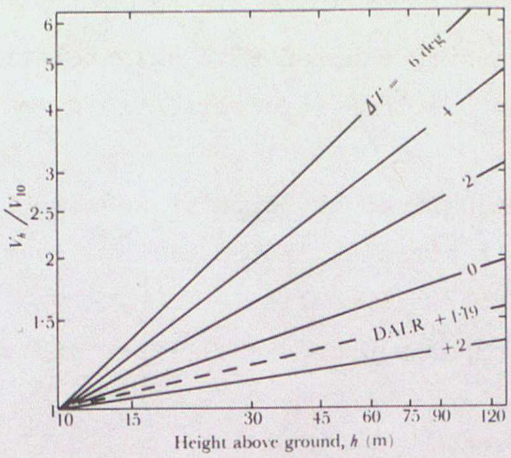


FIGURE 33. Ratio at Cardington for wind speed at height $h(V_h)$ to that at 10 metres (V_{10}) for different values of ΔT

ΔT = temperature at 1·2 m minus temperature at 120 m (heights above ground)

Frost[61] has considered the variation of wind speed with height in varying lapse conditions at Cardington; his results are shown in Figure 33. Individual profiles are likely to differ more from the mean than in neutral conditions with strong wind, and the diagram should not be regarded as giving more than a rough indication of the possible variation with height over flat, open countryside.

16.9 FORECASTING UPPER WINDS

Above the friction layer, forecast winds are derived mainly from computer-produced prognoses. For many purposes, such as aircraft route planning, the winds may be taken straight from the computer model, but for some other purposes it is at times advantageous to modify the computer products subjectively in regions where they are known to be prone to error. This modification is carried out before the prognoses are passed to the out-stations, as the process demands a thorough knowledge of the current computer model's characteristics in addition to a deep understanding of the behaviour of atmospheric systems; erroneous adjustments are likely to be worse than none at all. At 200 millibars and above, computer prognoses are less satisfactory than at lower levels, and more emphasis is placed on the use of the latest reports coupled with extrapolation of observed gradients and movements. Further information is given in Chapters

3 - Background to computer models
8 - Jet streams, tropopause and lower stratosphere
11 - Upper-air charts
13 - Computer prognoses: types and uses.

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