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HANDBOOK

OF

WEATHER FORECASTING

METEOROLOGICAL OFFICE

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PREFACE

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

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

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13.1. INTRODUCTION

Wind is air in motion and 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 or inferred from the action and interaction of winds. Wind analysis and prognosis is therefore of great importance to the meteorologist.

Information on winds is also of importance to many users of meteorological forecasts. For many decades there has been a close association between meteorologists and mariners. Information on surface winds at sea is supplied for ocean-going and coastal shipping and for amateur yachting. The strength and direction of surface winds over confined seas 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. Wind information is also of great importance for the safe, efficient and economic conduct of air operations. Forecasts of surface winds are required both by aircrew and ground personnel (for example, air traffic control officers) for the take-off and landing of aircraft. For the transit flight at altitude and for the adequate separation of aircraft, particularly in congested air spaces, forecasts of upper winds are provided. With the current types of aircraft in service the heights for which forecasts of winds are required as routine range from near the surface to heights of about 55,000 feet. Forecasts of wind are of importance to farmers and may be crucial to the successful spraying of crops. Winds are also important for the non-specialized user of meteorological information. For example, a wind stream from sea to land may provide a welcome moderation to air temperatures near the coast on days with high temperatures inland but, when the sea is cold, may produce cold and damp days in winter or spring. Strong winds with a cool airstream feel colder than a consideration of the air temperature alone would imply. It is appropriate therefore that, at times, some indication of these effects should be included in forecasts for the general public.

Air motion occurs on many scales and it is useful for the forecaster engaged on day-to-day forecasting to be aware of the scales of wind systems with which he is concerned and of their inter-relations. Leaving aside the possibility of long-term climatic change the wind systems of the general circulation have the largest scale in time and distance. Some understanding of the general circulation is a very useful background for the forecaster even though it can seldom be directly applied in day-to-day forecasting. The next lower scale of wind systems which should be recognized by the forecaster may be conveniently referred to as the "long-wave patterns" which can usually be found at the 700-millibar level and above over substantial portions of the earth. The long waves are intimately linked with the major atmospheric systems readily recognized on synoptic charts covering extensive parts of the earth, for example, a single trough and ridge may cover North America and the North Atlantic Ocean. These wind patterns are often relatively slow-moving and determine the general character of the weather over an area for several days. The practical forecaster must recognize and understand these systems - a point already stressed in earlier chapters. Embedded in these major patterns are smaller-scale wind perturbations associated with the smaller disturbances, for example, waves on cold fronts, minor troughs, small secondaries, etc. These smaller disturbances are often the cause of the changes in our weather on a time scale of a few hours and over distances of very few hundreds of miles. The synoptic network over Europe normally reveals these disturbances but they may tend to slip through the more open networks at sea. Wind disturbances on about this scale can therefore be observed and dealt with by normal synoptic methods. At times in dense reporting networks over land somewhat smaller-scale surface wind disturbances can be detected and followed from chart to chart, for example, small heat lows in summer, onset of sea-breezes, etc.

There are still smaller-scale disturbances of the wind field in the nature of eddies, both in the free air and near the ground, air motions associated with individual clouds or showers and disturbed air flow near and to the lee of hills. These motions are too small to be regularly revealed by the synoptic network and in many cases the air disturbances are so transient and ephemeral that they can be treated by the forecaster only in a statistical manner.

In most of the larger-scale wind disturbances the horizontal component of motion is far greater than the vertical component but the vertical component may play a vital role in the atmospheric processes of weather. In the smaller-scale disturbances the ratio of the vertical component of motion to the horizontal component may be greater than in the larger scale. Current techniques do not, however, permit of the reliable observation or forecasting of vertical components of wind. In the rest of this chapter any reference to wind will mean the horizontal component unless the vertical component is explicitly mentioned.

A complete treatment of winds in the atmosphere would contain full consideration of winds from the level in contact with earth's surface to the upper limit of the earth's atmosphere. Wind flow in the very lowest levels is scarcely the concern of the practical forecaster. The subject is complicated and its theoretical aspects are well treated in Sutton's[1*] text on micrometeorology. The lowest level which will be treated in any detail in this chapter is the wind at normal anemometer level (10 metres) above the surface. At upper levels it seems appropriate to place an upper bound somewhat above the level at which manned vehicles are likely to operate regularly and as routine in the reasonably near future. At the time of writing, 50 millibars seems a suitable value. Accordingly the chapter will be concerned primarily with wind at levels from anemometer height (10 metres) to about 50 millibars (about 20,000 metres).

Wind is the only element of a forecast which can be satisfactorily measured from a forecast chart of isobars or contours without the necessity for a great deal of judgement. Some judgement is of course called for but in many cases the direct application of a suitable engraved scale enables a first numerical approximation to the wind to be made. Some allowance for topography or for any apparent inconsistency of gradient in the isopleths will normally improve on the forecast depending purely on measurement, but the straightforward measurement is often close to the final forecast. It does not follow that wind forecasting is of itself more quantitative since the preparation of the forecast chart involves much judgement and subjective assessment. However, the interpretation of a forecast chart in terms of wind is far more quantitative than its interpretation in terms of temperature, cloud, precipitation, visibility or almost any other element of a forecast. This arises from the fact that the principles of fluid mechanics have been successfully applied to atmospheric air motions. The equations which can be deduced to

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

describe the theoretical wind are, however, complicated by the inclusion of terms due to the rotation of the earth, the vertical components of motion and the acceleration of the wind. Simplified mathematical relations are of much practical value and these will now be discussed.

13.2. MATHEMATICAL EXPRESSIONS FOR WIND

The theoretical derivation of the general equations of air motion is included in many meteorological textbooks which are widely available at all outstations and a treatment from first principles is not repeated in this handbook. The following résumé is included for ease of reference by practising forecasters.

13.2.1. *Geostrophic wind*

The geostrophic wind is the steady, uniform, horizontal wind which would occur in a frictionless atmosphere when the accelerations due to the pressure gradient force and the earth's rotation are in balance and there is no external force other than gravity. The geostrophic wind blows along the isobars with low pressure to the left of the flow in the northern hemisphere (to the right of the flow in the southern hemisphere).

If V_g is the geostrophic wind speed, G the horizontal gradient of pressure ($= \partial p/\partial n$ where differentiation is along a line normal to the isobars), ρ the air density, Φ the latitude and Ω the angular velocity of the earth, then

$$V_g = \frac{G}{2\rho\Omega\sin\Phi} = \frac{G}{f\rho} \qquad \ldots\ldots (1)$$

where $f = 2\Omega\sin\Phi$ (the Coriolis force).

The corresponding formula for the geostrophic wind on a constant pressure (that is, contour) chart is

$$V_g = \frac{g}{f} \times \text{gradient of height} = \frac{g}{f} \cdot \frac{\partial H}{\partial n} \qquad \ldots\ldots (2)$$

where g is the acceleration due to gravity and H is the height on the contour chart.

From equations (1) and (2) it is seen that on isobaric charts the geostrophic wind is uniquely determined by the gradient of pressure, the Coriolis force and the air density; on contour charts, the gradient of height, the Coriolis force and the acceleration due to gravity determine the geostrophic wind.

13.2.2. *Gradient wind*

It is implicit in the definition of the geostrophic wind that the path followed by the geostrophic wind is straight. In most natural weather systems an air parcel follows a curved path. If a parcel of air is moving steadily with velocity V in a cyclonic sense around a path whose radius of curvature is r, then an acceleration of V^2/r exists directed towards the centre of curvature.

By definition, the gradient wind, V_{gr}, is the flow of air which is necessary to balance the pressure gradient in steady circular flow. It follows that in cyclonically curved flow,

$$V_{gr} = \frac{G}{2\rho\Omega\sin\Phi} - \frac{V_{gr}^2}{2r\Omega\sin\Phi}$$

$$= V_g - \frac{V_{gr}^2}{fr} \qquad \ldots\ldots (3)$$

Thus, in cyclonic flow, the gradient wind is less than the geostrophic value.

For anticyclonic flow the relation is

$$V_{gr} = V_g + \frac{V_{gr}^2}{fr} \qquad \ldots . \ (4)$$

In anticyclonic flow the gradient wind exceeds the geostrophic value.

The term V_{gr}^2 / fr is often called the cyclostrophic term.

13.2.3. *Isallobaric wind*

If the pressure field is not constant, as assumed in deriving the geostrophic wind, but pressure changes are going on, then the wind differs from the geostrophic. Brunt and Douglas[2] have shown that an approximation to the true wind can be obtained by adding to the geostrophic wind an additional wind with components u_i and v_i which arise from the effect of the pressure changes. This wind is known as the isallobaric wind and is given by

$$u_i = -\frac{1}{f^2\rho}\frac{\partial}{\partial x}\left(\frac{\partial p}{\partial t}\right) \quad \text{and} \quad v_i = -\frac{1}{f^2\rho}\frac{\partial}{\partial y}\left(\frac{\partial p}{\partial t}\right) \qquad \ldots . \ (5)$$

It can be seen from these equations that the isallobaric wind is proportional to the isallobaric gradient and is directed towards the (algebraically) smaller pressure tendencies.

13.2.4. *Thermal wind*

It is sometimes desirable to estimate the thermal wind between two contour charts at 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

$$u_T = -\frac{g}{f}\frac{\partial h'}{\partial y} \quad \text{and} \quad v_T = \frac{g}{f}\frac{\partial h'}{\partial x} \qquad \ldots . \ (6)$$

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.

13.3. SOME PRACTICAL METHODS FOR EVALUATING WINDS FROM METEOROLOGICAL WORKING CHARTS

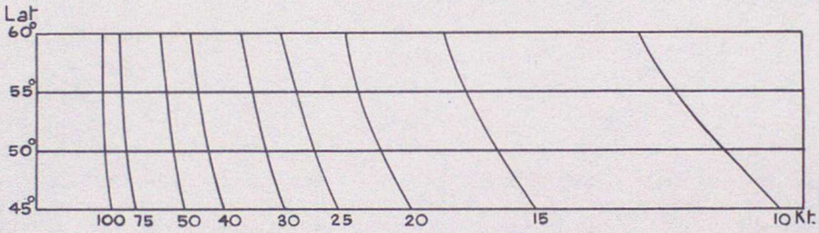
Some theoretical and observational results regarding the deviation of the actual from the geostrophic or gradient winds are considered later in Sections 13.5., 13.6., 13.7., 13.9. and 13.10. This section deals solely with some of the practical methods available for computing winds from working charts.

13.3.1. *Geostrophic wind*

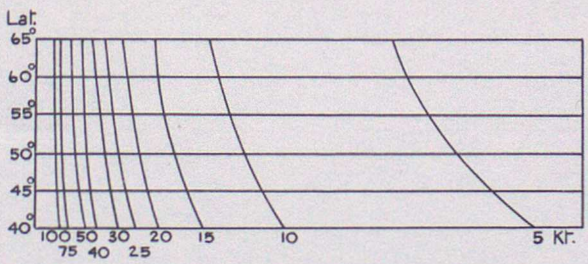
Many Meteorological Office printed working charts contain an inset on which a grid is printed. Such grids often contain curves indicating the geostrophic wind for suitable isobar or contour intervals at various latitudes appropriate to the particular chart. Some forecasters prefer to use dividers to step off the intervals along the normals between neighbouring isopleths and then to obtain the corresponding geostrophic wind from the printed grid. Other forecasters prefer to have geostrophic scales engraved on a transparent base and to use these for calculating geostrophic winds. The scales are used by placing the engraved scale(s) at right angles to the isopleths with the base line of the scale in contact with one of the isopleths. The distance along the normal between the base and a neighbouring isopleth, when read off the appropriate engraved scale, provides a meausre of the geostrophic wind.

When charts on several scales are in use several transparent scales are required but some forecasters use only one or two scales and apply conversion factors either directly from the relative scales of the chart in use and the chart for which the geostrophic transparency is valid, or indirectly by taking either a suitable fraction of the speed or a measurement across isobaric or contour intervals which make the necessary corrections, for example, a scale engraved for 60-metre contours on a chart of scale 1 to 7,500,000 can be used on a chart of scale 1 to 15,000,000 if the speeds are halved or if contours at 120-metre intervals are selected.

A selection of suitable scales for use in temperate latitudes on Meteorological Office charts are illustrated in Figures 1 and 2. A direct reading geostrophic wind scale is illustrated in Figure 3.

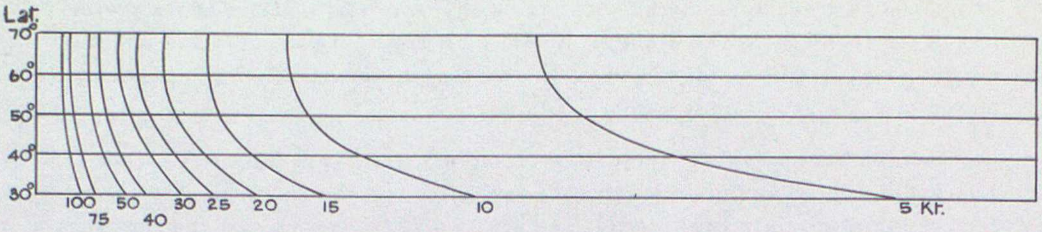


(a) For use with isobars at 2 mb. intervals on a chart where the natural scale is 1:3,000,000.

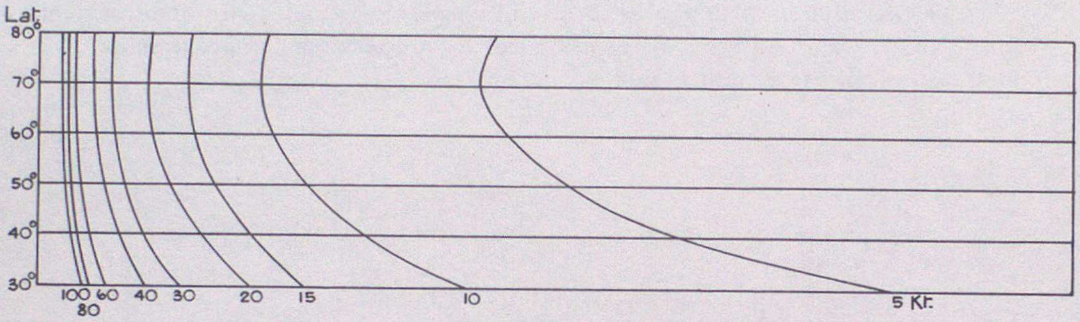


(b) For use with isobars at 1 mb. intervals and air at pressure 1015 mb., temperature 50°F. (For an increase of 10 mb. pressure subtract 1%; for an increase of 5°F. temperature add 1%). This scale is valid for a map constructed on the conical orthomorphic projection with two standard parallels 30° and 60° where the natural scale is 1:5,000,000.

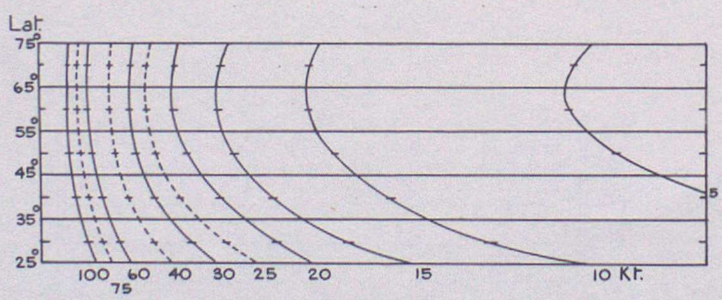
FIGURE 1 *Geostrophic scales for use with isobars on charts drawn at mean sea level*



(*c*) For use with isobars at 2 mb. intervals and air at pressure 1000 mb., temperature 50°F. (For an increase of 10 mb. pressure subtract 1%; for an increase of 5°F. temperature add 1%). This scale is valid for a map constructed on the conical orthomorphic projection with two standard parallels 30° and 60° where the natural scale is 1:7,500,000.



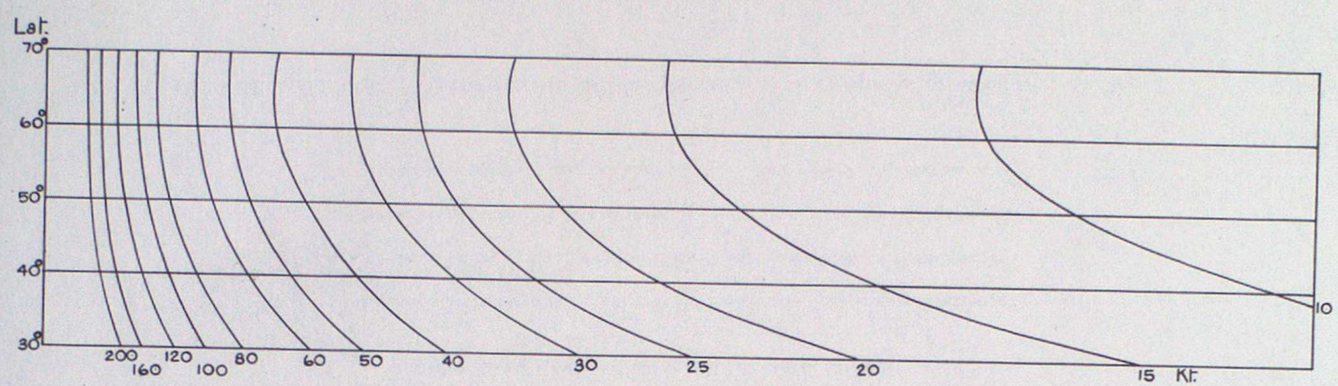
(*d*) For use with isobars at 2 mb. intervals and air at pressure 1000 mb., temperature 50°F. (For an increase of 10 mb. pressure subtract 1%; for an increase of 5°F. temperature add 1%). This scale is valid for a map constructed on the conical orthomorphic projection with two standard parallels 45° and 65° where the natural scale is 1:10,000,000.



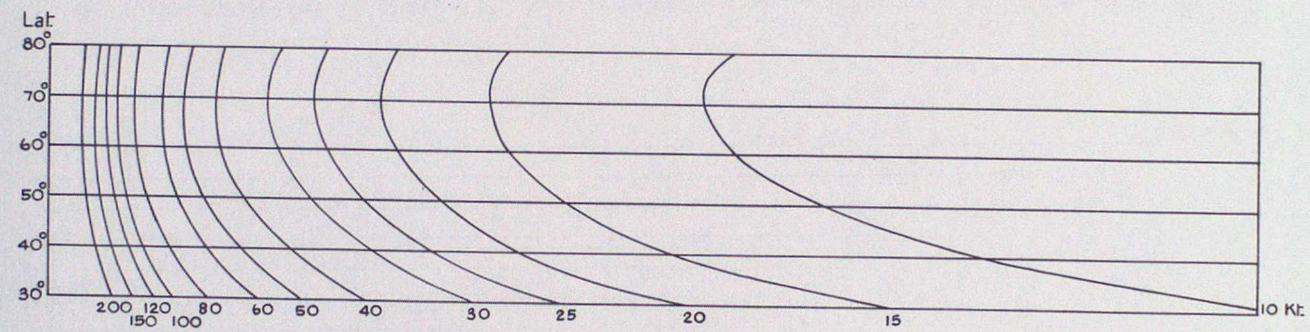
(*e*) For use with isobars at 4 mb. intervals and mean air density at 2,000 ft. elevation of 1·163 kg. per cu.m. This scale is valid for a map constructed on a conformal conic projection with two standard parallels 30° and 60° where the natural scale is 1:15,000,000.

(*f*) For scale for use with charts where the natural scale is 1:30,000,000 see Figure 2 (*c*).

FIGURE 1 (*contd.*) *Geostrophic scales for use with isobars on charts drawn at mean sea level*



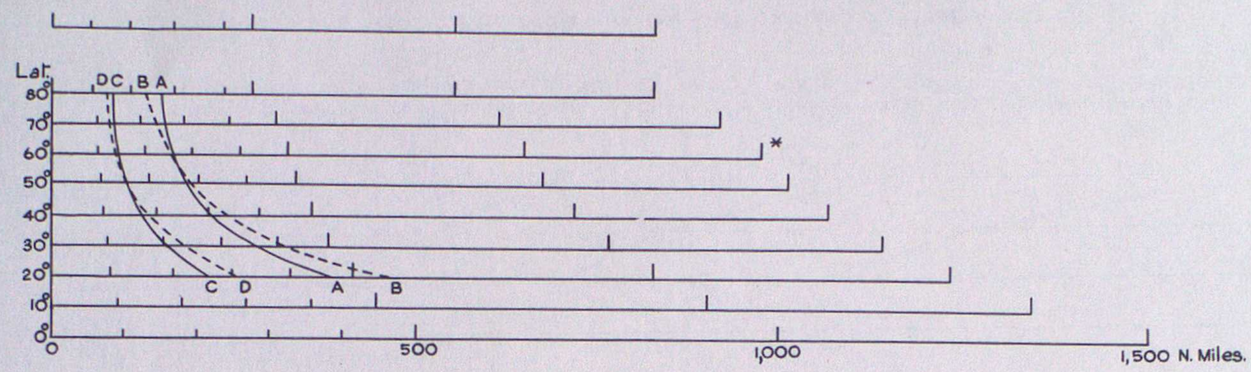
(a) For use with contours at 60 m. intervals. This scale is valid for a map constructed on the conical orthomorphic projection with two standard parallels 30° and 60° where the natural scale is 1:7,500,000.



(b) For use with contours at 60 m. intervals. This scale is valid for a map constructed on the conical orthomorphic projection with two standard parallels 45° and 65° where the natural scale is 1:10,000,000.

FIGURE 2 *Geostrophic scales for use with contours on isobaric charts*

Scale of distances measured along a meridian in latitude 80°-90°N.



(c) Scales for wind speed 20 kt. This scale is valid for a map constructed on equidistant azimuthal projection on the plane of 60°N. where the natural scale is 1:30,000,000.

- Scale of distances measured along a meridian in latitudes 0°-80°N.

A Distance between 60m. contours for E.-W. flow (N.-S. measurements)

B Distance between 60m. contours for N.-S. flow (E.-W. measurements)

C Distance between 4 mb. isobars for E.-W. flow (N.-S. measurements)

D Distance between 4 mb. isobars for N.-S. flow (E.-W. measurements)

FIGURE 2 *(contd.) Geostrophic scales for use with contours on isobaric charts*

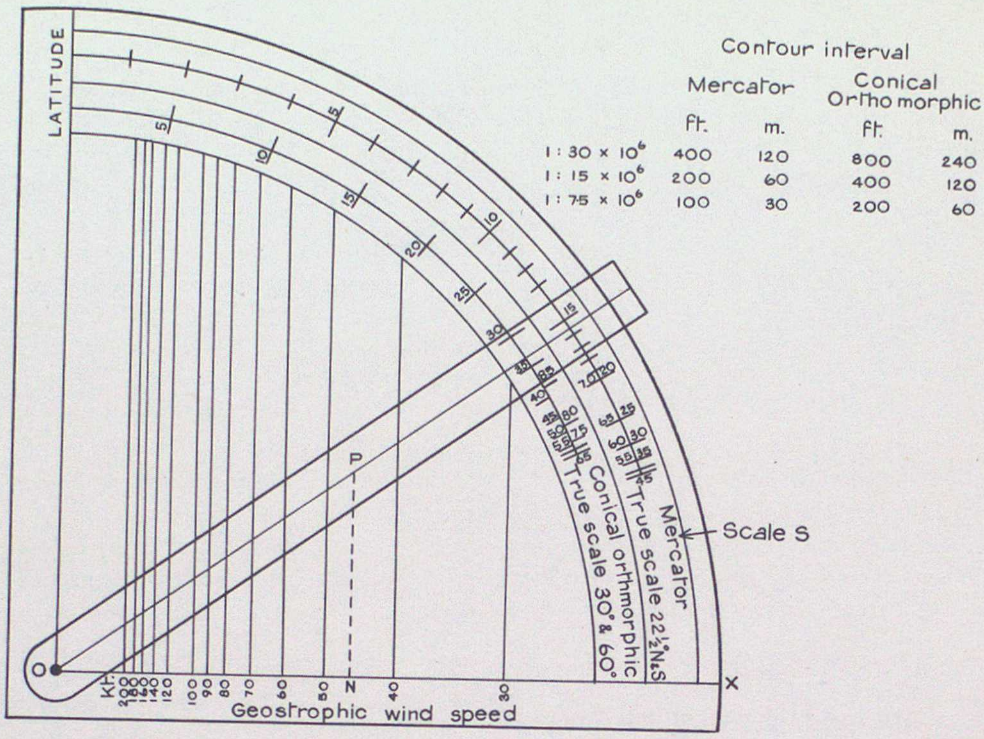


FIGURE 3 *Scale graduated for use on Mercator's projection with standard parallels 22½°N. and 22½°S., and on the conical orthomorphic projection with standard parallels 30° and 60°*

This scale is due to Matthews[3] who has given the theory of the instrument and its method of use which is as follows: Draw the common normal PN to two consecutive contours of the isobaric surface at the desired position. Set the radial cursor to the latitude on scale S and place the scale on the map, adjusting its position so that N lies on OX, PN is perpendicular to OX and P lies on the engraved line of the radial cursor. When this adjustment is complete the geostrophic wind is read off the scale along OX at the point N. With a little practice the scale may be used without actually drawing the normal PN.

13.3.2. *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[4] has shown that the ratio M of the geostrophic wind (V_g) to the gradient wind (V_{gr}) is expressed by the following equation

$$M\left(1 - M + \frac{C\cos\Psi}{fr_i}\right) = \frac{V_g}{fr_i}, \qquad \ldots\ldots (7)$$

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.

Gilbert[5] modified the equation slightly by introducing P as the percentage correction to be applied to the geostrophic wind to determine the gradient wind. Then

$$V_{gr} = V_g \left(1 + \frac{P}{100}\right) . \qquad \ldots . (8)$$

But since $V_g = MV_{gr}$, then $M = 100/(100 + P)$ and substituting for M in equation (7) leads to

$$\frac{100}{100 + P} \left(\frac{P}{100 + P} + \frac{C \cos \Psi}{fr_i}\right) = \frac{V_g}{fr_i} \qquad \ldots . (9)$$

P is positive for anticyclonic curvature and negative for cyclonic curvature. P can be determined from equation (9) since V_g, C and Ψ can be obtained from the working charts and f is readily calculated for a given latitude. From equation (9) Gilbert produced tables from which values of P can be readily determined by interpolation once values for V_g, C and Ψ have been obtained. These tables, reproduced here as Tables 1 and 2 are used as follows:

From the working chart

(i) Obtain the geostrophic wind V_g at the point considered.

(ii) Compute $C \cos \Psi$, the component of the velocity of the pressure system in the direction of the geostrophic wind, in knots ($C \cos \Psi$ is positive or negative as the component has the same sense as or the opposite sense to the geostrophic wind at the point under consideration).

(iii) Obtain the radius of curvature r_i (in nautical miles) of the isobar at the point under consideration.

From Table 1

(iv) Interpolate as necessary for V_g against the latitude in the upper part of Table 1 and then go downwards to find $100\ V_g/fr_i$ in the lower part of Table 1. $100\ C \cos \Psi/fr_i$ is found similarly from $C \cos \Psi$, Ψ and r_i. The radius of curvature, r_i, is taken as positive in both the cyclonic and anticyclonic cases.

From Table 2

(v) Use Table 2*(a)* in the cyclonic and Table 2*(b)* in the anticyclonic case to find P, the percentage correction. Tables 2*(a)* and 2*(b)* contain $100\ V_g/fr_i$ in the left-hand column and $100\ C \cos \Psi/fr_i$ in the body of the table. Use the values found in (iv) for $100\ V_g/fr_i$ and $100\ C \cos \Psi/fr_i$, enter Table 2*(a)* or 2*(b)* as appropriate and read the percentage correction, P, in the row at the head of the Table.

Gilbert also constructed the diagram reproduced in Figure 4. The diagram is used as follows. Determine $100\ V_g/fr_i$ and $100\ C \cos \Psi/fr_i$ from Table 1, enter the appropriate part of the Figure 4 with these values and interpolate for P between the lines of constant percentage correction. The left-hand part of the figure applies to cyclonically curved isobars and the right-hand part to anticyclonically curved isobars. It should be noted that the scale is much larger in the right-hand part of Figure 4. The radius of curvature of the isobar is taken as positive in both cases and the figure allows for both positive and negative values

TABLE 1 *Evaluation of* $\frac{100\,V_g}{f r_i}$ *and* $\frac{100\,C\cos\Psi}{f r_i}$

Φ	V_g or $C\cos\Psi$ for various latitudes (knots)																			
30°	7.0	13.0	20.0	27.0	33.0	40.0	46.0	53.0	60.0	66.0	73.0	79.0	86.0	93.0	99.0	106.0	113.0	119.0	126.0	133.0
40°	9.0	17.0	25.0	34.0	43.0	51.0	60.0	68.0	77.0	85.0	94.0	102.0	111.0	119.0	128.0	136.0	145.0	153.0	162.0	170.0
50°	10.0	20.0	30.0	40.0	50.0	60.0	70.0	80.0	90.0	100.0	110.0	120.0	130.0	140.0	150.0	160.0	170.0	180.0	190.0	200.0
60°	11.0	23.0	34.0	45.0	57.0	68.0	79.0	91.0	102.0	113.0	125.0	136.0	148.0	159.0	170.0	182.0	193.0	204.0	216.0	227.0
70°	12.0	25.0	37.0	49.0	62.0	74.0	87.0	99.0	111.0	124.0	136.0	148.0	161.0	173.0	185.0	198.0	210.0	223.0	235.0	247.0
r_i (n. miles)	$\frac{100\,V_g}{f r_i}$ or $\frac{100\,C\cos\Psi}{f r_i}$ for various values of r_i																			
50	50	99	149	199	249	298	348	398	447	497	-	-	-	-	-	-	-	-	-	-
100	25	50	75	99	124	149	174	199	224	249	273	298	323	348	372	398	422	447	472	497
200	12.4	25	37	50	62	75	87	99	112	124	137	149	162	174	186	199	211	224	236	249
300	8.3	16.6	25	33	41	50	58	66	75	83	91	99	108	116	124	133	141	149	157	166
400	6.2	12.4	18.7	25	31	37	43	50	56	62	68	75	81	87	93	99	106	112	118	124
500	5.0	9.9	14.9	19.9	25	30	35	40	45	50	55	60	65	70	75	79	85	89	94	99
600	4.1	8.3	12.4	16.6	21	25	29	33	37	41	46	50	54	58	62	66	70	75	79	83
700	3.6	7.1	10.7	14.2	17.7	21	25	28	32	35	39	43	46	50	53	57	60	64	67	71
800	3.1	6.2	9.3	12.4	15.5	18.6	22	24	28	31	34	37	40	43	47	50	53	56	59	62
900	2.8	5.5	8.3	11.0	13.8	16.6	19.3	22	25	28	30	33	36	39	41	44	47	50	53	55
1000	2.5	5.0	7.5	9.9	12.4	14.9	17.4	19.9	22	25	27	30	32	35	37	40	42	45	47	50
1250	2.0	4.0	6.0	8.0	9.9	11.9	13.9	15.9	17.9	19.9	22	24	26	28	30	32	34	36	38	40
1500	1.7	3.3	5.0	6.6	8.3	9.9	11.6	13.3	14.9	16.6	18.2	19.9	21	23	25	27	28	30	31	33
1750	1.4	2.8	4.3	5.7	7.1	8.5	9.9	11.4	12.9	14.2	15.7	17.1	18.5	19.9	21	23	24	26	27	28
2000	1.2	2.5	3.7	5.0	6.2	7.4	8.7	9.9	11.2	12.4	13.7	14.9	16.2	17.4	18.6	19.9	21	22	24	25
2500	1.0	2.0	3.0	4.0	5.0	6.0	7.0	8.0	8.9	9.9	10.9	11.9	12.9	13.9	14.9	15.9	16.9	17.9	18.9	19.9

TABLE 2(*a*) *Evaluation of percentage addition (cyclonic curvature)*

$\frac{100\,V_g}{fr_i}$	Percentage correction										
	0	−5	−10	−15	−20	−25	−30	−35	−40	−45	−50
	$\frac{100\,C\cos\Psi}{fr_i}$										
0	0	−5.3	−11.1	−17.6	−25.0	−33.0	−43.0	−54.0	−67.0	−82.0	−100.0
2	2.0	−3.4	− 9.3	−15.9	−23.0	−32.0	−41.0	−53.0	−65.0	−81.0	− 99.0
4	4.0	−1.5	− 7.5	−14.3	−22.0	−30.0	−40.0	−51.0	−64.0	−80.0	− 98.0
6	6.0	0.4	− 5.7	−12.5	−20.0	−29.0	−39.0	−50.0	−63.0	−79.0	− 97.0
8	8.0	2.3	− 3.9	−10.9	−18.6	−27.0	−37.0	−49.0	−62.0	−77.0	− 96.0
10	10.0	4.2	− 2.1	− 9.1	−17.0	−26.0	−36.0	−47.0	−61.0	−76.0	− 95.0
12	12.0	6.1	− 0.3	− 7.5	−15.4	−24.0	−35.0	−46.0	−59.0	−75.0	− 94.0
14	14.0	8.0	1.5	− 5.7	−13.8	−23.0	−33.0	−45.0	−58.0	−74.0	− 93.0
16	16.0	9.9	3.3	− 4.1	−12.2	−21.0	−32.0	−43.0	−57.0	−73.0	− 92.0
18	18.0	11.8	5.1	− 2.3	−10.6	−19.8	−30.0	−42.0	−56.0	−72.0	− 91.0
20	20.0	13.7	6.9	− 0.7	− 9.0	−18.3	−29.0	−41.0	−55.0	−71.0	− 90.0
22	22.0	15.6	8.7	1.1	− 7.4	−16.8	−27.0	−39.0	−53.0	−70.0	− 89.0
24	24.0	17.5	10.5	2.7	− 5.8	−15.3	−26.0	−38.0	−52.0	−69.0	− 88.0
26	26.0	19.4	12.3	4.5	− 4.2	−13.8	−25.0	−37.0	−51.0	−67.0	− 87.0
28	28.0	21.0	14.1	6.1	− 2.6	−12.3	−23.0	−36.0	−50.0	−66.0	− 86.0
30	30.0	23.0	15.9	7.9	− 1.0	−10.8	−22.0	−34.0	−49.0	−65.0	− 85.0
32	32.0	25.0	17.7	9.5	0.6	− 9.3	−21.0	−33.0	−47.0	−64.0	− 84.0
34	34.0	27.0	19.5	11.3	2.2	− 7.8	−19.1	−32.0	−46.0	−63.0	− 83.0
36	36.0	29.0	21.0	12.9	3.8	− 6.3	−17.7	−31.0	−45.0	−62.0	− 82.0
38	38.0	31.0	23.0	14.7	5.4	− 4.8	−16.3	−29.0	−44.0	−61.0	− 81.0
40	40.0	33.0	25.0	16.3	7.0	− 3.3	−14.9	−28.0	−43.0	−60.0	− 80.0
42	42.0	35.0	27.0	18.1	8.6	− 1.8	−13.5	−27.0	−41.0	−59.0	− 79.0
44	44.0	37.0	29.0	19.7	10.2	− 0.3	−12.1	−25.0	−40.0	−58.0	− 78.0
46	46.0	38.0	30.0	21.0	11.8	1.2	−10.7	−24.0	−39.0	−57.0	− 77.0
48	48.0	40.0	32.0	23.0	13.4	2.7	− 9.3	−23.0	−38.0	−55.0	− 76.0
50	50.0	42.0	34.0	25.0	15.0	4.2	− 7.9	−21.0	−37.0	−54.0	− 75.0
55	55.0	47.0	38.0	29.0	19.0	7.9	− 4.4	−18.1	−34.0	−52.0	− 73.0
60	60.0	52.0	43.0	33.0	23.0	11.7	− 0.9	−14.9	−31.0	−49.0	− 70.0
65	65.0	57.0	47.0	38.0	27.0	15.4	2.6	−11.6	−28.0	−46.0	− 67.0
70	70.0	61.0	52.0	42.0	31.0	19.2	6.1	− 8.3	−25.0	−43.0	− 65.0
75	75.0	66.0	56.0	46.0	35.0	23.0	9.6	− 5.1	−22.0	−41.0	− 63.0
80	80.0	71.0	61.0	50.0	39.0	27.0	13.1	− 1.9	−18.7	−38.0	− 60.0
85	85.0	75.0	65.0	55.0	43.0	30.0	16.6	1.4	−15.7	−35.0	− 57.0
90	90.0	80.0	70.0	59.0	47.0	34.0	20.0	4.7	−12.7	−32.0	− 55.0
95	95.0	85.0	74.0	63.0	51.0	38.0	24.0	7.9	− 9.7	−30.0	− 53.0
100	100.0	90.0	79.0	67.0	55.0	42.0	27.0	11.1	− 6.7	−27.0	− 50.0
105	105.0	95.0	83.0	72.0	59.0	45.0	31.0	14.4	− 3.7	−24.0	− 47.0
110	110.0	99.0	88.0	76.0	63.0	49.0	34.0	17.7	− 0.7	−21.0	− 45.0
115	115.0	104.0	92.0	80.0	67.0	53.0	38.0	21.0	2.3	−18.6	− 43.0
120	120.0	109.0	97.0	84.0	71.0	57.0	41.0	24.0	5.3	−15.8	− 40.0
125	125.0	113.0	101.0	89.0	75.0	60.0	45.0	27.0	8.3	−13.1	− 37.0
130	130.0	118.0	106.0	93.0	79.0	64.0	48.0	31.0	11.3	−10.3	− 35.0
135	135.0	123.0	110.0	97.0	83.0	68.0	52.0	34.0	14.3	− 7.6	− 33.0
140	140.0	128.0	115.0	101.0	87.0	72.0	55.0	37.0	17.3	− 4.8	− 30.0
145	145.0	133.0	119.0	106.0	91.0	75.0	59.0	40.0	20.0	− 2.1	− 27.0

Wind

TABLE 2(*a*) (*contd.*) *Evaluation of percentage addition (cyclonic curvature)*

$\frac{100\,V_g}{f r_i}$	Percentage correction 0	−5	−10	−15	−20	−25	−30	−35	−40	−45	−50
	$\frac{100\,C\cos\Psi}{f r_i}$										
150	150.0	137.0	124.0	110.0	95.0	79.0	62.0	44.0	23.0	0.7	− 25.0
155	155.0	142.0	128.0	114.0	99.0	83.0	66.0	47.0	26.0	3.4	− 23.0
160	160.0	147.0	133.0	118.0	103.0	87.0	69.0	50.0	29.0	6.2	− 20.0
165	165.0	151.0	137.0	123.0	107.0	90.0	73.0	53.0	32.0	8.9	− 17.5
170	170.0	156.0	142.0	127.0	111.0	94.0	76.0	57.0	35.0	11.7	− 15.0
175	175.0	161.0	146.0	131.0	115.0	98.0	80.0	60.0	38.0	14.4	− 12.5
180	180.0	166.0	151.0	135.0	119.0	102.0	83.0	63.0	41.0	17.2	− 10.0
185	185.0	171.0	155.0	140.0	123.0	105.0	87.0	66.0	44.0	19.9	− 7.5
190	190.0	175.0	160.0	144.0	127.0	109.0	90.0	70.0	47.0	23.0	− 5.0
195	195.0	180.0	164.0	148.0	131.0	113.0	94.0	73.0	50.0	25.0	− 2.5
200	200.0	185.0	169.0	152.0	135.0	117.0	97.0	76.0	53.0	28.0	0
205	205.0	189.0	173.0	157.0	139.0	120.0	101.0	79.0	56.0	31.0	2.5
210	210.0	194.0	178.0	161.0	143.0	124.0	104.0	83.0	59.0	34.0	5.0
220	220.0	204.0	187.0	169.0	151.0	132.0	111.0	89.0	65.0	39.0	10.0
230	230.0	213.0	196.0	178.0	159.0	139.0	118.0	96.0	71.0	45.0	15.0
240	240.0	223.0	205.0	186.0	167.0	147.0	125.0	102.0	77.0	50.0	20.0
250	250.0	232.0	214.0	195.0	175.0	154.0	132.0	109.0	83.0	56.0	25.0
260	260.0	242.0	223.0	203.0	183.0	162.0	139.0	115.0	89.0	61.0	30.0
270	270.0	251.0	232.0	212.0	191.0	169.0	146.0	122.0	95.0	67.0	35.0
280	280.0	261.0	241.0	220.0	199.0	177.0	153.0	128.0	101.0	72.0	40.0
290	290.0	270.0	250.0	229.0	207.0	184.0	160.0	135.0	107.0	78.0	45.0
300	300.0	280.0	259.0	237.0	215.0	192.0	167.0	141.0	113.0	83.0	50.0
310	310.0	289.0	268.0	246.0	223.0	199.0	174.0	148.0	119.0	89.0	55.0
320	320.0	299.0	277.0	254.0	231.0	207.0	181.0	154.0	125.0	94.0	60.0
330	330.0	308.0	286.0	263.0	239.0	214.0	188.0	161.0	131.0	100.0	65.0
340	340.0	318.0	295.0	271.0	247.0	222.0	195.0	167.0	137.0	105.0	70.0
350	350.0	327.0	304.0	280.0	255.0	229.0	202.0	174.0	143.0	111.0	75.0
360	360.0	337.0	313.0	288.0	263.0	237.0	209.0	180.0	149.0	116.0	80.0
370	370.0	346.0	322.0	297.0	271.0	244.0	216.0	187.0	155.0	122.0	85.0
380	380.0	356.0	331.0	305.0	279.0	252.0	223.0	193.0	161.0	127.0	90.0
390	390.0	365.0	340.0	314.0	287.0	259.0	230.0	200.0	167.0	133.0	95.0
400	400.0	375.0	349.0	322.0	295.0	267.0	237.0	206.0	173.0	138.0	100.0
410	410.0	384.0	358.0	331.0	303.0	274.0	244.0	213.0	179.0	144.0	105.0
420	420.0	394.0	367.0	339.0	311.0	282.0	251.0	219.0	185.0	149.0	110.0
430	430.0	403.0	376.0	348.0	319.0	289.0	258.0	226.0	191.0	155.0	115.0
440	440.0	413.0	385.0	356.0	327.0	297.0	265.0	232.0	197.0	160.0	120.0
450	450.0	422.0	394.0	365.0	335.0	304.0	272.0	239.0	203.0	166.0	125.0
460	460.0	432.0	403.0	373.0	343.0	312.0	279.0	245.0	209.0	171.0	130.0
470	470.0	441.0	412.0	382.0	351.0	319.0	286.0	252.0	215.0	177.0	135.0
480	480.0	451.0	421.0	390.0	359.0	327.0	293.0	258.0	221.0	182.0	140.0
490	490.0	460.0	430.0	399.0	367.0	334.0	300.0	265.0	227.0	188.0	145.0
500	500.0	470.0	439.0	407.0	375.0	342.0	307.0	271.0	233.0	193.0	150.0

TABLE 2(*b*) *Evaluation of percentage addition (anticyclonic curvature)*

$\frac{100\,V_g}{f r_i}$	Percentage correction										
	0	5	10	15	20	25	30	35	40	45	50
	$\frac{100\,C\cos\Psi}{f r_i}$										
0	0	−4.8	−9.1	−13.0	−16.7	−20.0	−23.0	−26.0	−29.0	−31.0	−33.0
2	2.0	−2.7	−6.9	−10.7	−14.3	−17.5	−21.0	−23.0	−26.0	−28.0	−30.0
4	4.0	−0.6	−4.7	− 8.4	−11.9	−15.0	−17.9	−21.0	−23.0	−25.0	−27.0
6	6.0	1.5	−2.5	− 6.1	− 9.5	−12.5	−15.3	−17.8	−20.0	−22.0	−24.0
8	8.0	3.6	−0.3	− 3.8	− 7.1	−10.0	−12.7	−15.1	−17.4	−19.4	−21.0
10	10.0	5.7	1.9	− 1.5	− 4.7	− 7.5	−10.1	−12.4	−14.6	−16.5	−18.3
12	12.0	7.8	4.1	0.8	− 2.3	− 5.0	− 7.5	− 9.7	−11.8	−13.6	−15.3
14	14.0	9.9	6.3	3.1	0.1	− 2.5	− 4.9	− 7.0	− 9.0	−10.7	−12.3
16	16.0	12.0	8.5	5.4	2.5	0	− 2.3	− 4.3	− 6.2	− 7.8	− 9.3
18	18.0	14.1	10.7	7.7	4.9	2.5	0.3	− 1.6	− 3.4	− 4.9	− 6.3
20	20.0	16.2	12.9	10.0	7.3	5.0	2.9	1.1	− 0.6	− 2.0	− 3.3
22	22.0	18.3	15.1	12.3	9.7	7.5	5.5	3.8	2.2	0.9	− 0.3
24	24.0	20.0	17.3	14.6	12.1	10.0	8.1	6.5	5.0	3.8	2.7
26	26.0	23.0	19.5	16.9	14.5	12.5	10.7	9.2	7.8	6.7	5.7
28	28.0	25.0	22.0	19.2	16.9	15.0	13.3	11.9	10.6	9.6	8.7
30	30.0	27.0	24.0	21.0	19.3	17.5	15.9	14.6	13.4	12.5	11.7
32	32.0	29.0	26.0	24.0	22.0	20.0	18.5	17.3	16.2	15.4	14.7
34	34.0	31.0	28.0	26.0	24.0	23.0	21.0	20.0	19.0	18.3	17.7
36	36.0	33.0	31.0	28.0	27.0	25.0	24.0	23.0	22.0	21.0	21.0
38	38.0	35.0	33.0	31.0	29.0	27.0	26.0	25.0	25.0	24.0	24.0
40	40.0	37.0	35.0	33.0	31.0	30.0	29.0	28.0	27.0	27.0	27.0
42	42.0	39.0	37.0	35.0	34.0	33.0	31.0	31.0	30.0	30.0	30.0
44	44.0	41.0	39.0	38.0	36.0	35.0	34.0	33.0	33.0	33.0	33.0
46	46.0	43.0	41.0	40.0	39.0	37.0	37.0	36.0	36.0	36.0	-
48	48.0	46.0	44.0	42.0	41.0	40.0	39.0	39.0	39.0	-	-
50	50.0	48.0	46.0	45.0	43.0	43.0	42.0	42.0	41.0	-	-
55	55.0	53.0	51.0	50.0	49.0	49.0	48.0	-	-	-	-
60	60.0	58.0	57.0	56.0	55.0	55.0	-	-	-	-	-
65	65.0	63.0	62.0	62.0	61.0	-	-	-	-	-	-
70	70.0	69.0	68.0	67.0	-	-	-	-	-	-	-
75	75.0	74.0	73.0	73.0	-	-	-	-	-	-	-
80	80.0	79.0	79.0	-	-	-	-	-	-	-	-
85	85.0	85.0	-	-	-	-	-	-	-	-	-
90	90.0	90.0	-	-	-	-	-	-	-	-	-
95	95.0	-	-	-	-	-	-	-	-	-	-
100	100.0	-	-	-	-	-	-	-	-	-	-

Note.- For all tables r_i is counted positive; the sign is accounted for in the percentage correction.

of $C\cos\Psi$. Only an approximation to P can usually be obtained from Figure 4, and, for accurate work, the tables are to be preferred.

The following example illustrates the practical use of the tables.

Example: V_g = 47 knots, $C\cos\Psi$ = +30 knots, Φ = 40°, r_i = 500 n. miles.

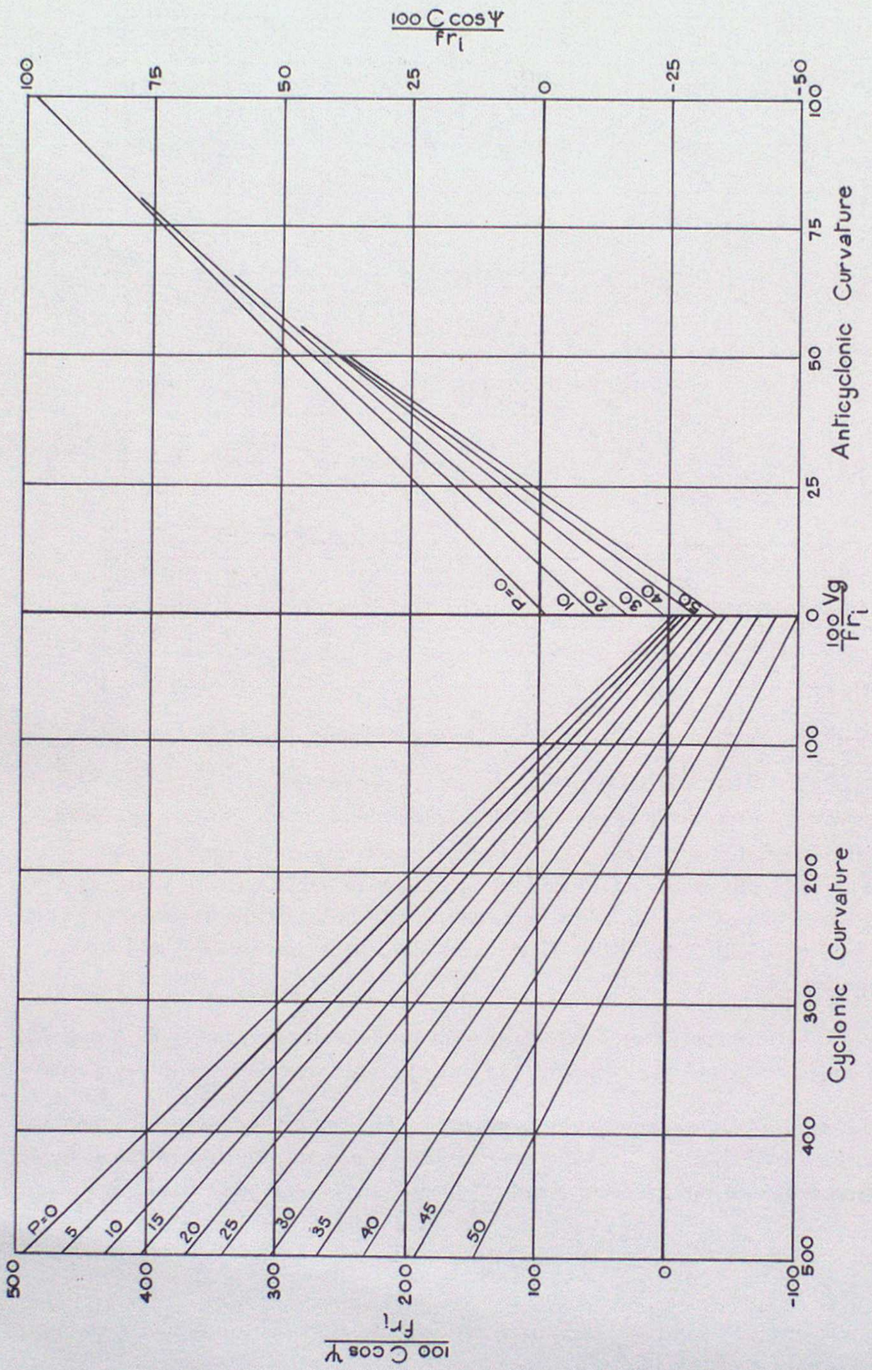


FIGURE 4 ***Diagram for determining the gradient wind***

100 C cos Ψ/fr_i is positive when C cos Ψ has the same sense as the geostrophic wind and negative when C cos Ψ has the opposite sense to the geostrophic wind.

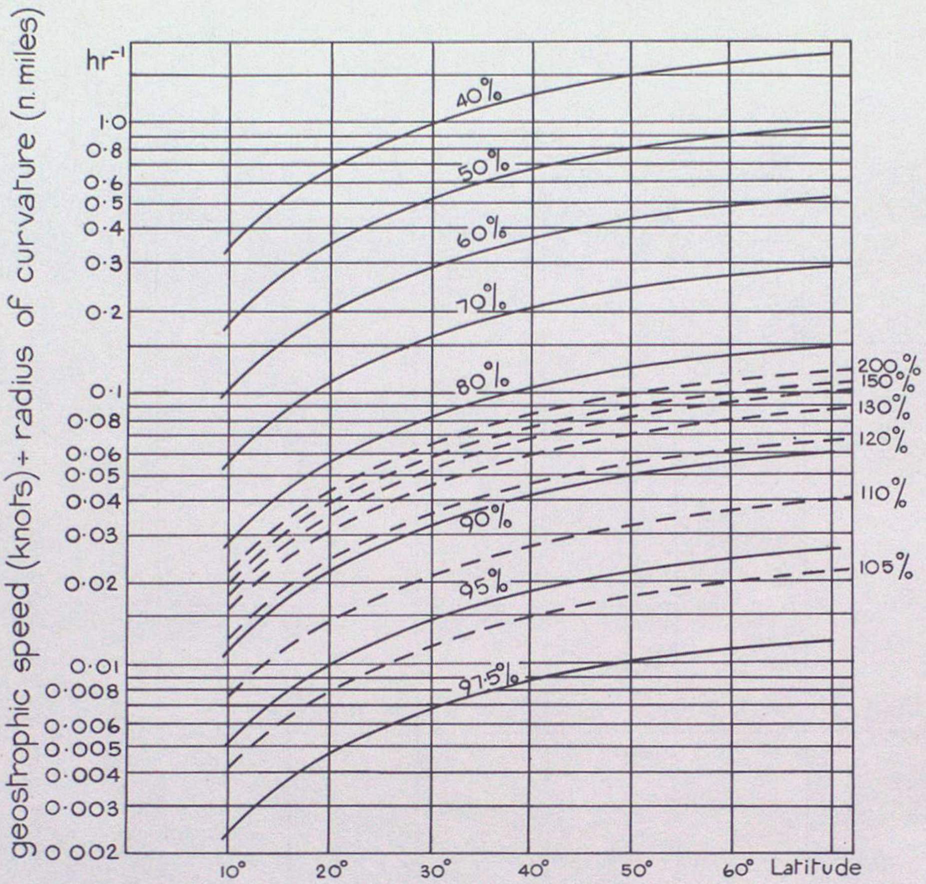


FIGURE 5 *Graph of V_{gr} as a percentage of V_g*

The anticyclonic case is shown by broken lines, the cyclonic case by full lines

Since, in the head table of Table 1, 47 is mid-way between 43 and 51, then $100\ V_g/fr_i$ is mid-way between 25 and 30 in the lower table and its value is 27·5. Similarly, since $C \cos \Psi = 30$ knots is near the mid-point of the interval between 25·0 and 34·0 then $100C \cos \Psi/fr_i$ lies proportionally between 14·9 and 19·9, that is, equals 18.

In the cyclonic case enter Table 2(*a*) with these values, namely 27·5 and 18, then P is found to be −7 per cent; in the anticyclonic case P is +16 per cent.

Another method for obtaining the gradient wind from the geostrophic wind has been described by Silvester.[6] The method rests on a manipulation of the following equation between the gradient wind (V_{gr}) and the geostrophic wind (V_g).

$$V_{gr} = V_g \pm \frac{V_{gr}^2}{fr_i},$$

where V_{gr} is gradient wind, V_g is geostrophic wind, f is $2\,\Omega \sin \Phi$ and r_i is the radius of curvature of the isobars.

The positive sign is taken for anticyclonic curvature and the negative sign for cyclonic curvature. From a consideration of this equation Silvester obtained a diagram from which V_{gr} could be read from values obtainable from the synoptic working chart. The diagram is reproduced in Figure 5. It is used as follows. From the synoptic working chart measure V_g (in knots) and r_i (in nautical miles) at the point under consideration and note its latitude. Calculate V_g/r_i, enter

Figure 5 with this value and the latitude and interpolate as necessary to obtain V_{gr} as a percentage of V_g. The full lines are used for the cyclonic case and the dashed lines for the anticyclonic case.

It should be noted that Silvester's method does not allow for the movement of the pressure system.

13.3.3. *Isallobaric wind*

Brunt and Douglas[2] have shown that 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 mb or 4 mb 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 6.

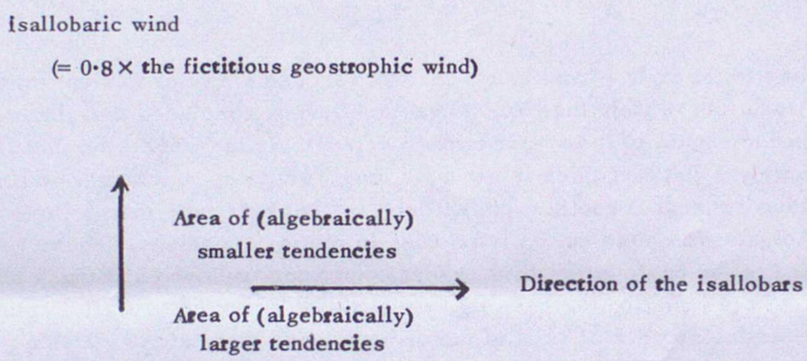


FIGURE 6 *Isallobaric wind*

13.3.4. *Thermal wind*

A measure of the mean thermal wind for a layer is obtained by applying the geostrophic contour scale (appropriate to the scale of the charts) to the thickness isopleths of the layer under consideration.

13.4. INSTRUMENTAL ERRORS OF WIND OBSERVATIONS

13.4.1. *Surface winds (at 10 metres)*

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]

13.4.2. *Upper winds*

In the United Kingdom measurements of upper wind is normally carried out by the radar-tracking of a suitable target carried aloft by a balloon. The radar equipment in use at the time of writing is known as the G.L.III. Bannon[8] has made estimates of the probable instrumental errors in determining the upper winds by G.L.III radar equipment. Some of his results are shown in Table 3.

TABLE 3 *Root mean square vector error (to the nearest knot) as a function of height and mean wind*

Mean Wind	*Height in thousands of feet* 3	10	20	30	40	50	60
				Knots			
10	Less than 1 knot	1	1	1	1	2	2
20		1	1	2	2	2	3
30		1	1	2	3	3	4
40		1	2	3	3	4	5
50		1	2	3	4	5	6
60		1	3	4	5	6	
70		2	3	4	6	7	
80		2	3	5	6	7	
90		2	4	5	7		
100		2	4	6			

Figures to the right of and below the lines in Table 3 refer to slant ranges from the radar set of more than 70,000 yards. Bannon concluded that the order of the root mean square of the vector error in determining the mean wind in a layer approximately 1,200 feet thick is not more than 2 knots up to a height of 10,000 feet, seldom exceeds 3 knots at 20,000 feet and is never more than 5 knots at greater heights which are currently reached in routine soundings. These values are related to the basic errors of the radar equipment employed at British land stations. Some variation on these figures would be expected to arise in practice from the varying degrees of skill of the operators and the state of efficiency of the radar equipment. However some independent work by Johnson[9] indicates that any such variation is probably very small on the average. Johnson estimated the standard vector errors in 100-millibar winds reported by British land stations and found them to be generally less than 6 knots. This value is in very close agreement with the value arrived at by Bannon from considerations of geometry and of the characteristics of the G.L.III radar set.

Radar wind observations from British ocean weather ships do not normally have the same accuracy as those from the land-based radar stations. This arises partly from the radar equipment used and partly from the fact that the radar set is mounted in a ship which may be steaming slowly but, even if hove to, is still subject to pitch and roll. For a number of reasons it is not possible to make reliable estimates of the errors in radar wind observations from ocean weather ships. However, if two ships are close together and make simultaneous observations on the same radar target, it is possible to obtain from the two independent wind measurements their vector differences which are a measure of the errors. Two separate series of such pairs of observations have been examined statistically by Harrison.[10,11] Table 4 shows the mean vector differences for the second series, the results of which were not greatly different from the earlier series.[10]

The mean vector difference for all observations for all heights was 6·9 knots. It will be seen from Table 4 that there were rather large values for 1,000 and 2,000 feet. The reason for this is not known. At levels above 2,000 feet the mean vector difference increased with increasing height and reached a

TABLE 4 *Mean vector differences of pairs of wind measurements*

Height	*Mean vector difference*	*No. of pairs of observations*	*Height*	*Mean vector difference*	*No. of pairs of observations*
ft.	*kt.*		*ft.*	*kt.*	
1,000	6.8	17	23,000	7.0	25
2,000	5.0	18	26,000	9.0	25
4,000	3.1	18	30,000	11.2	25
6,000	3.5	20	35,000	9.8	25
8,000	3.8	22	40,000	10.3	24
10,000	3.5	22	45,000	9.5	23
13,000	5.1	23	50,000	8.3	18
16,000	5.0	24	55,000	6.0	9
18,000	5.7	25	60,000	8.5	4
20,000	7.0	25			

maximum at about 30,000 to 40,000 feet; it tended to decrease above this height. The frequency distribution of the vector differences shows that 46 per cent were less than 4·5 knots and about 90 per cent less than 14·5 knots. In the data examined by Harrison there were a few very large individual vector differences - one was as high as 50 knots. There is no obvious explanation of these large individual differences which, when they occurred, tended to do so in the upper part or very near the end of an ascent. There was no connexion between the mean vector difference for an ascent and the state of the sea.

Since the true mean wind is likely to lie, in the mean, between the values found in pairs of observations of the same radar target, it follows that the probable error of wind observation will be less than the values of the vector differences found by Harrison.

In view of the comparatively small use currently made of pilot balloons for upper air observations in Europe, a discussion of their observational errors is not included in this handbook.

So far as the estimation of upper winds by nephoscopes is concerned, perhaps all that need be said is that the errors in the assumed height of the cloud on which the observation is made have a large effect on the calculated velocity and hence on the usefulness of the measurement. Nevertheless nephoscope observations may be useful on occasions, particularly for the determination of wind directions. Synoptic inferences may sometimes be deduced from time changes in the direction of movements of upper clouds when observed from a single station.

Some remarks on the accuracy of aircraft observations of wind are included in Section 13.12.

13.5. STATISTICS AND OBSERVATIONAL DATA ON SURFACE WIND

13.5.1. *Over land*

Some data regarding the ratio of the surface wind (at anemometer height) to the geostrophic wind for sixteen wind directions at a number of locations have been compiled by Shaw.[12] 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 over a number of years but others on observations for a single year only. The data are therefore not homogeneous. Data for a number of sites are reproduced in Table 5. It is readily seen that there are substantial

TABLE 5 *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° N.N.E.	45° N.E.	67½° E.N.E.	90° E.	112½° E.S.E.	135° S.E.	157½° S.S.E.	180° S.	202½° S.S.W.	225° S.W.	247½° W.S.W.	270° W.	292½° W.N.W.	315° N.W.	337½° N.N.W.	Mean of all
			ft.		ft.																	
Falmouth	1	All	41	Between cliff and harbour	167	.36	.36	.33	.30	.31	.32	.32	.31	.29	.28	.29	.33	.34	.35	.34	.36	.32
Pendennis Castle	1	All	65	Conical headland	256	.58	.66	.74	.83	.85	.83	.74	.69	.63	.58	.54	.52	.53	.56	.54	.54	.65
Pyrton Hill	2	All	98	Slope: hills to plain	500	.39	.38	.48	.48	.40	.37	.37	.42	.48	.40	.42	.43	.47	.49	.44	.40	.43
Southport	5	All	62	Flat sand shore	18	.69	.60	.53	.49	.48	.47	.47	.45	.41	.38	.38	.42	.51	.63	.65	.69	.52
St. Mary's, Scilly	8	4	32	Hilly island	118	.74	.63	.59	.59	.67	.70	.59	.47	.51	.45	.49	.49	.54	.63	.65	.61	.58
Aberdeen	8	4	74	College roof	46	.35	.45	.61	.57	.51	.45	.43	.35	.36	.32	.32	.31	.39	.46	.54	.43	.43
Spurn Head	8	4	40	Spit of sand	26	.80	.85	.64	.72	.65	.72	.64	.54	.49	.49	.49	.53	.59	.76	.76	.73	.65
Yarmouth	8	4	40	Spit of sand	13	.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	15	.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	15	.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	77	.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	27	.25	.29	.29	.29	.24	.13	.11	.12	.13	.23	.24	.22	.18	.18	.24	.19	.21

B: Beaufort force estimated

variations in the ratio from site to site and for different wind directions at each site. The variation of the ratio with wind direction can be readily displayed pictorially by plotting the ratios on a polar diagram with the observing point at the centre. If information on the angle of deviation of the surface wind from the geostrophic wind is available this too can be displayed on the same polar diagram. Figure 7, taken from Shaw, shows one method of displaying all this information for one station.

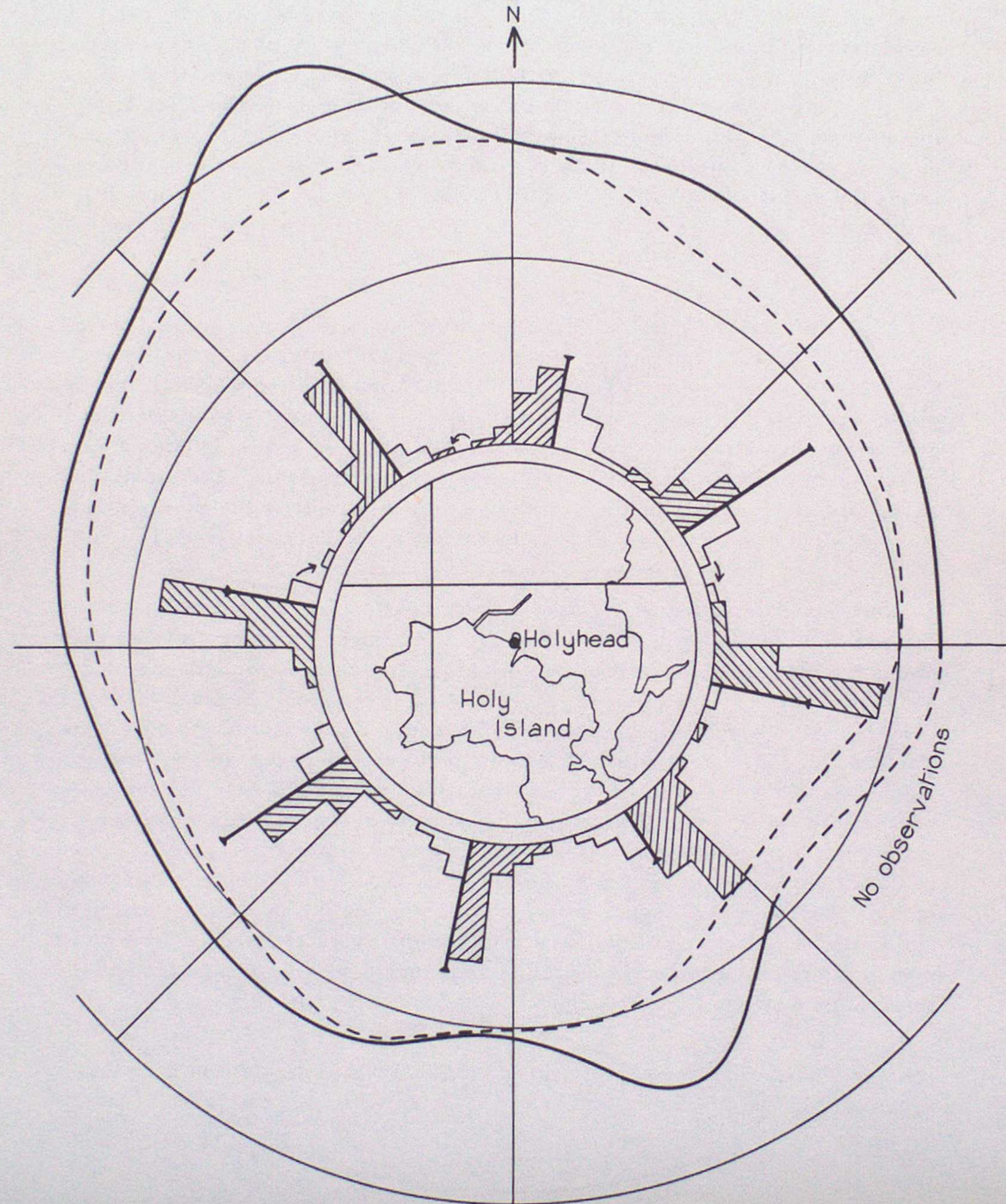


FIGURE 7 *Relation between the geostrophic and observed surface winds at Holyhead*

Radius of map = 4 miles

In the inner circle of the figure is displayed a circular portion (centred at the observing site) of the ordnance survey map on a suitably reduced scale. This illustrates the geographical relief in the immediate neighbourhood of the site. The outermost circle represents a ratio of 1·0 and the next inner circle a ratio of 0·5, that is, surface wind speeds equal to and one-half of the geostrophic value respectively. The full line represents the ratio of surface to geostrophic wind speed for winds of Beaufort force 6 and the dashed line the ratio for winds of force 4. The deviation of the surface wind from the geostrophic wind is shown by columns representing percentage frequencies of deviations of given amount arranged in groups, for steps of two points, on either side of a middle group embracing four points, namely two points of veer and two of back. The percentage number of cases in this central group is indicated by the length of a pin-shaped mark on the scale 1 inch of length (shown by the distance between two consecutive circles in the figure) to 50 per cent. The percentage frequency of the other groups is shown on the same scale by the length of the respective columns, the shaded columns indicating the surface winds which are backed from the geostrophic wind and the unshaded columns those which are veered. In these frequency columns winds of all forces have been included.

A more recent and detailed investigation for one site (Gorleston) has been reported by Durst.[13] Gorleston is a coastal site and coastal effects will exert some influence on the observed surface winds. Thus offshore winds will not be wholly applicable to surface winds well inland nor will onshore winds be completely representative of surface winds over the sea. No similar investigation for an inland station or a maritime station appears to be available. The greater detail of the work reported by Durst appears to make it preferable to the more general and older (but nevertheless still instructive and useful) account given by Shaw.

Gorleston is situated on the East Anglian coast which, in that neighbourhood, extends in a broadly north - south direction. 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 8 and 9.

The diurnal variation of the ratio and the relation of the ratio to the wind speed are shown in Tables 6 and 7.

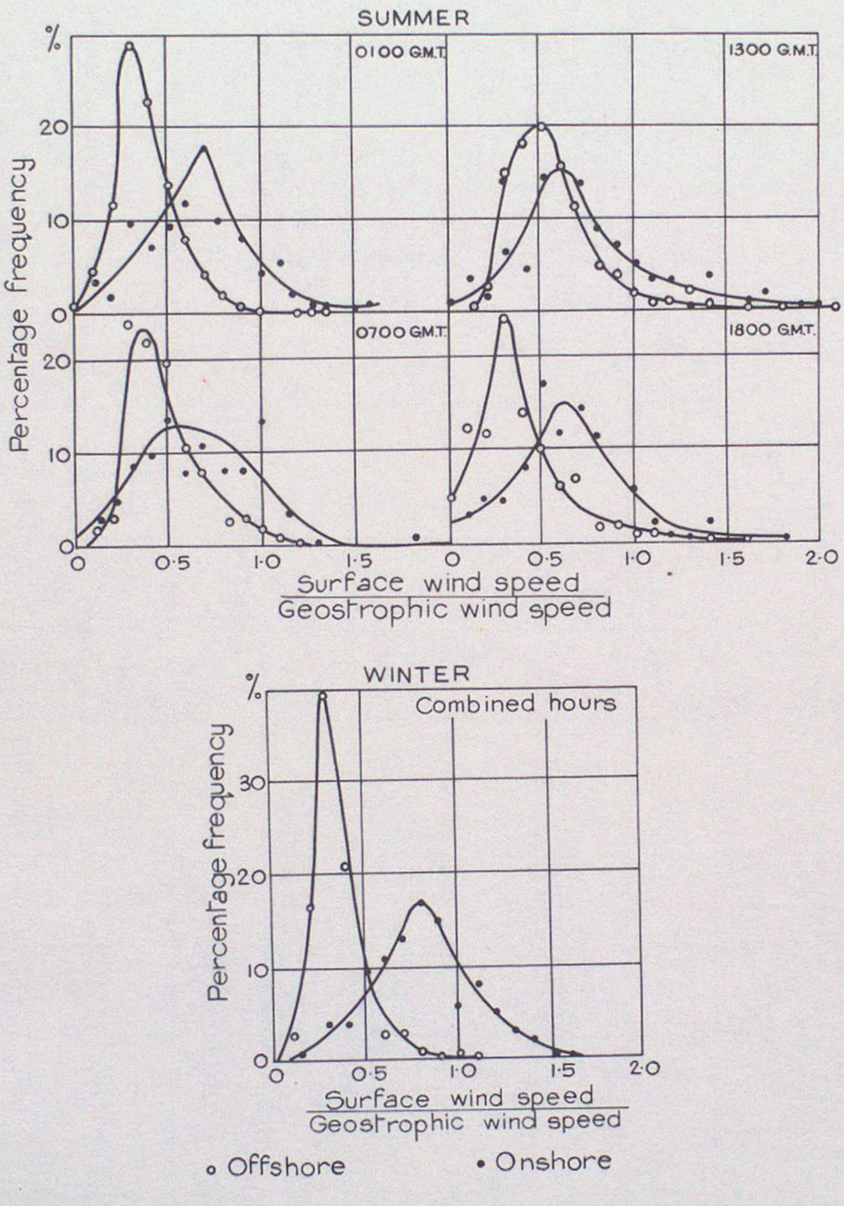


FIGURE 8 *Seasonal and diurnal variation of the ratio of surface to geostrophic wind at Gorleston*

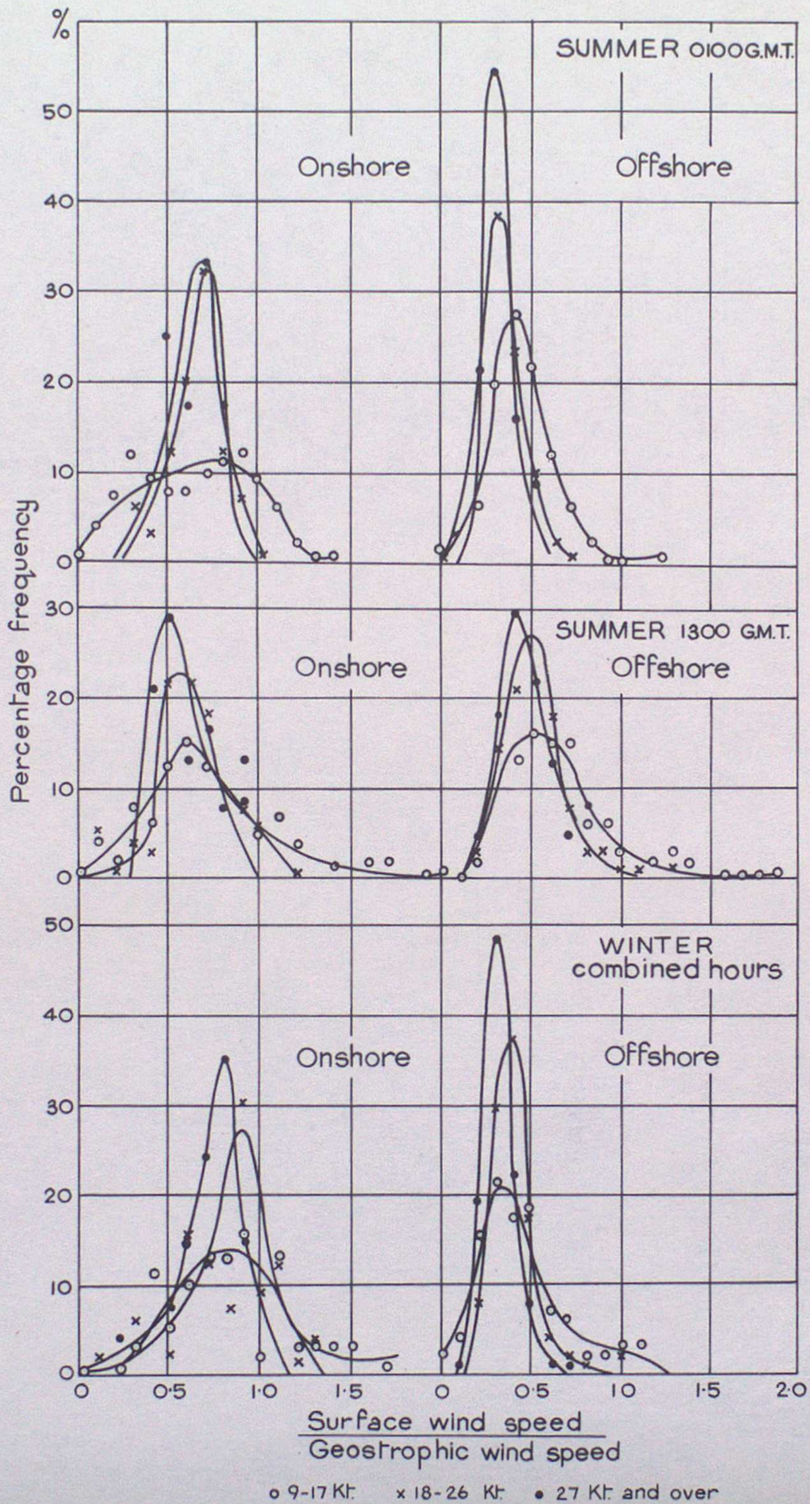


FIGURE 9 *Variation with wind speed of the ratio of surface to geostrophic wind*

Wind

TABLE 6 *Diurnal variation of the ratio of surface to geostrophic wind*

	Time of observation (G.M.T.)							
	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 7 *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*
	G.M.T.						
Winter 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

The broad conclusions are:

(i) Onshore winds are approximately four fifths the geostrophic wind in winter and two thirds in summer.

(ii) There is practically no diurnal variation in the ratio in winter.

(iii) Offshore winds are approximately one third of the geostrophic wind in winter and just over two fifths in summer.

(iv) In summer there is a marked diurnal variation. The offshore winds are relatively weak at night and relatively strong in the day-time.

(v) With a geostrophic wind speed of less than 18 knots the onshore wind has a wide range of speed and may even exceed the geostrophic; the offshore wind is less scattered at night but has about the same scatter in the day-time.

The general state of the sky inland was determined from examination of the Beaufort letters recorded during the six hours preceding each main synoptic hour. Table 8 gives the modes of the variation of the ratio (surface wind/geostrophic wind) when the skies over the land were mainly clear, broken or overcast.

TABLE 8 *Variation of most probable value of the ratio of surface to geostrophic wind with the various states of the sky over the land*

		Mainly clear sky		*Broken skies*		*Overcast*	
	Time	*On-shore*	*Off-shore*	*On-shore*	*Off-shore*	*On-shore*	*Off-shore*
	G.M.T.						
Winter December, January	All hours	0.72	0.32	0.82	0.34	0.80	0.32
Summer May - August	0100	0.65	0.39	0.58	0.34	0.75	0.36
	1300	0.54	0.43	0.63	0.55	0.75	0.49

Table 8 indicates that the state of the sky inland is no great help in deducing the surface wind from the geostrophic wind in winter or offshore winds in summer. It is of doubtful assistance for night-time onshore winds in summer but it does show that the more cloudy the midday sky the more nearly will the onshore wind at that time approach the geostrophic value.

Information on the direction of the surface wind in relation to the pressure gradient is contained in Table 9 which sets out the mean values of the geostrophic wind together with its standard scalar deviation and the vector mean values of the surface wind together with its standard vector deviation.

TABLE 9 *Mean values of surface and geostrophic winds*

	Time	*No. of obs.*	*Geostrophic wind*	*Standard deviation*	*Surface wind**		*Standard vector deviation*
	G.M.T.		*kt.*	*kt.*	°	*kt.*	*kt.*
Winter							
Offshore	0100	76	27	12	33	8	8
	0700	67	28	17	33	8	5
	1300	83	29	14	30	10	7
	1800	68	31	10	36	9	8
	All hours	294	28	14	32	9	7
Onshore	0100	53	20	10	24	14	10
	0700	51	20	10	24	14	10
	1300	48	20	8	25	15	8
	1800	53	20	10	24	16	9
	All hours	205	20	10	24	14	10
Summer							
Offshore	0100	204	17	8	37	6	4
	1300	187	18	9	39	7	8
Onshore	0100	81	15	7	39	9	6
	1300	100	15	7	45	8	7

* Angle given is that between geostrophic and surface wind directions.

Table 9 shows that the surface wind on the average lies at an angle of 30° to 40° to the isobars with offshore winds but in winter with onshore winds it is only 24° to the isobars. However, the correlation coefficient of the surface wind with the geostrophic wind is comparatively small, being no more than about 0·4 or 0·5 with both onshore and offshore winds.

The results of a comparison of winds recorded by anemograph and the geostrophic wind during the year 1946 at Stornoway, Bell Rock, Scilly Isles and Kew have been reported by Marshall.[14] Ratios of the surface to geostrophic wind speeds are given in Table 10. The deviation of the surface from the geostrophic wind is contained in Table 11.

From his investigation Marshall concluded that "winds were backed about 10° more at night than by day at Stornoway and Kew Observatory on the average. At Bell Rock winds were backed rather more at 1800 G.M.T. than at the other main synoptic hours but were close to the isobars at both 0600 G.M.T. and 1800 G.M.T., and seemed often to be veered a little from the geostrophic wind direction at midday and midnight. At the Scilly Isles the wind deviated little from the general run of the isobars at night and was backed 10-20° by day."

Wind

TABLE 10 *Ratio of surface wind speed to geostrophic wind speed*

Location	Geostrophic wind direction	No. of cases	Time (G.M.T.) 0000	0600	1200	1800	All hours
Stornoway	Northerly	218	.38	.41	.54	.50	.46
	Easterly	184	.53	.54	.66	.64	.60
	Southerly	392	.53	.49	.54	.53	.52
	Westerly	462	.47	.48	.58	.53	.52
Bell Rock	Northerly	198	.62	.77	.66	.64	.67
	Easterly	188	.90	.85	.85	.80	.85
	Southerly	327	.58	.58	.60	.66	.61
	Westerly	504	.69	.64	.67	.72	.68
Scilly Isles	Northerly	193	.80	.79	.72	.86	.79
	Easterly	269	.66	.63	.62	.58	.62
	Southerly	259	.56	.58	.61	.54	.57
	Westerly	515	.72	.68	.68	.71	.70
Kew Observatory	Northerly	174	.27	.26	.48	.38	.35
	Easterly	255	.38	.37	.56	.50	.45
	Southerly	200	.27	.28	.42	.33	.32
	Westerly	563	.28	.29	.41	.36	.34

TABLE 11 *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

Marshall found that, at the four locations included in his investigation, the diurnal variation of surface wind speed over the year as a whole was greatest with northerly winds. The least variation occurred with geostrophic winds from a southerly point at Stornoway, with southerlies and westerlies at Bell Rock and with westerlies at the Scilly Isles and Kew Observatory.

Some very detailed values for the diurnal variation of wind speeds observed at Cardington were given by Giblett and others[15] and a condensed version of one of the tables is given in Table 12. It can be seen that the ratios of the wind near the surface to the geostrophic wind are greater by day than by night and that there is a definite seasonal variation. In summer the ratios by day are decidedly greater than in winter. Table 13 shows the frequency of the deviation of the wind from the isobars at Cardington as computed from one year's records. It indicates that there is both a seasonal and diurnal variation. For the whole year, the backing of the wind had a mode of 30 to 34° in the early hours and 20 to 24° in the early afternoon. Table 12 shows that at Cardington the wind at 150 feet was occasionally very light in relation to the geostrophic wind but, on a few occasions, it reached

TABLE 12 *Frequency of ratios of mean hourly wind at 150 feet to geostrophic wind at Cardington, March 1929 - February 1930*

Months	*Time G.M.T.*	*Ratio of mean hourly wind at 150 feet to geostrophic wind* 0.2	0.3	0.4	0.5	0.6	0.7	0.8	0.9	1.0	1.1	1.2	1.3	1.4	1.5	1.6	1.7	1.8	1.9	2.0	*Indeterminate*
		Number of occasions																			
December 1929	0100	1	11	15	24	15	7	3	4	..	1	..	..	..	..	..	..	..	..	..	9
January 1930	0700	1	6	15	18	17	10	8	2	3	1	2	1	..	..	..	..	..	..	..	6
February 1930	1300	..	5	7	13	16	18	10	11	..	2	1	..	..	..	..	..	..	..	..	7
	1800	1	2	13	21	21	9	12	4	1	..	2	..	..	..	..	..	..	..	..	4
March 1929	0100	1	3	10	20	11	11	5	7	3	..	..	1	..	1	..	..	..	..	..	19
April 1929	0700	..	5	6	18	22	11	5	2	3	..	..	..	..	..	..	..	..	..	..	20
May 1929	1300	2	2	7	10	7	17	10	6	5	7	4	1	1	..	..	..	..	..	..	12
	1800	1	3	5	19	7	13	13	4	2	5	3	4	2	..	..	..	..	..	..	11
June 1929	0100	3	3	13	10	15	13	3	5	3	3	3	..	..	..	..	..	..	..	..	18
July 1929	0700	..	3	11	8	14	9	17	10	5	2	..	..	..	..	..	..	..	..	..	13
August 1929	1300	..	3	2	7	11	16	10	11	6	9	5	..	1	..	1	2	1	..	..	7
	1800	..	2	1	6	12	19	8	7	2	11	..	3	2	2	2	1	1	..	1	12
September 1929	0100	..	7	19	18	13	11	2	2	6	3	..	..	..	..	..	..	..	..	..	10
October 1929	0700	2	15	11	22	13	11	6	..	1	1	..	..	..	..	..	..	..	..	..	9
November 1929	1300	1	4	9	10	18	10	11	12	3	3	1	..	..	2	..	..	..	..	1	6
	1800	1	4	10	18	13	9	11	7	2	2	1	1	..	1	..	..	1	..	..	10
Year:	0100	5	24	57	72	54	42	13	18	12	7	3	1	..	1	..	..	..	..	..	56
March 1929 to	0700	3	29	43	66	66	41	36	14	12	4	2	1	..	..	..	..	..	..	..	48
February 1930	1300	3	14	25	40	52	61	41	40	14	21	11	1	2	2	1	2	1	..	1	32
	1800	3	11	29	64	53	50	44	22	7	18	6	8	4	3	2	1	2	..	1	37

TABLE 13 *Frequency of backing of wind at 150 feet from direction of isobars at Cardington, March 1929 - February 1930*

Months	Time G.M.T.	Veered	0°-4°	5°-9°	10°-14°	15°-19°	20°-24°	25°-29°	30°-34°	35°-39°	40°-44°	45°-49°	50°-54°	55° and over
		Deviation of surface wind from isobars — *Number of occasions*												
December 1929, January 1930, February 1930	0100	3	6	3	4	6	7	12	17	6	6	6	4	5
	0700	1	1	..	6	6	13	10	17	13	7	4	1	2
	1300	1	5	4	10	9	19	11	9	4	3	2	1	1
	1800	1	..	1	5	11	6	11	18	12	4	5	7	3
March 1929, April 1929, May 1929	0100	5	..	1	3	2	7	3	6	11	11	2	2	14
	0700	6	1	2	2	3	10	7	11	5	4	5	4	8
	1300	6	5	5	5	11	3	9	5	6	4	3	..	12
	1800	11	7	1	2	6	5	8	4	3	5	5	3	13
June 1929, July 1929, August 1929	0100	4	1	6	8	7	6	9	4	7	2	4	4	5
	0700	8	4	8	6	8	7	6	11	2	6	2	2	3
	1300	13	2	6	8	1	12	6	6	3	6	6	1	4
	1800	9	4	4	1	7	7	11	5	3	6	3	1	6
September 1929, October 1929, November 1929	0100	6	1	2	4	5	10	9	14	9	4	3	4	5
	0700	1	..	1	3	9	9	13	9	8	9	4	7	4
	1300	6	6	8	12	9	10	7	7	8	2	2	..	2
	1800	3	2	3	6	8	7	9	7	7	6	4	7	5
Year: March 1929 to February 1930	0100	18	8	12	19	20	30	33	41	33	23	15	14	29
	0700	16	6	11	17	26	39	36	48	28	26	15	14	17
	1300	26	18	23	35	30	44	33	27	21	15	13	2	19
	1800	24	13	9	14	32	25	39	34	25	21	17	18	27

or even exceeded the geostrophic value. In his investigation Marshall[14] also commented on surface winds which were abnormally strong or light in relation to the geostrophic wind. Values equal to the geostrophic speed occurred fairly frequently at Stornoway, Bell Rock and the Scilly Isles. Table 14 shows some percentage frequencies.

TABLE 14 *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

It will be seen that these occurrences were fairly frequent with geostrophic winds from the easterly quadrant at Stornoway, from the northerly, easterly and westerly quadrants at Bell Rock and from the northerly and westerly quadrants at the Scilly Isles. The occurrences were generally associated with marked low-level instability. Marked anticyclonic curvature of the isobars was also apparent on some of these occasions.

The percentage occurrences of surface winds which were abnormally light in relation to the geostrophic wind is given in Table 15, also due to Marshall.[14]

TABLE 15 *Percentage frequency of occurrence of surface wind less than a quarter of the geostrophic wind*

Geostrophic wind	*Stornoway*+	*Bell Rock**	*Scilly Isles**	*Kew Observatory*+
	%	%	%	%
Northerly	13	12	3	4
Easterly	10	1	5	5
Southerly	20	6	1	2
Westerly	4	3	2	6

+ 1200 G.M.T. only. * All main synoptic hours.

Mean hourly surface wind speeds less than a quarter of the geostrophic value were infrequent at the open exposure at the Scilly Isles and at Bell Rock (except for northerly geostrophic winds at Bell Rock). The rather high frequency for northerlies at Bell Rock is ascribed by Marshall to the presence of small lee depressions in the Bell Rock area formed by the passage of the north-westerly airstream over the mountains of Scotland. (Marshall also found well marked abnormalities in wind direction at Bell Rock with geostrophic winds between north-west and north.) Abnormally light surface winds at Kew Observatory were common at night. At midday, however, they were more common at Stornoway than at Kew Observatory especially with geostrophic winds from the south-south-east. There was a tendency for abnormally light surface winds to be associated with thermal stability in the lower layers or cyclonic curvature of the isobars but this was not always the case.

Some very detailed information of surface winds observed at Cardington is contained in *Geophysical Memoirs* No. 54[15] 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 worthwhile. Outstations are strongly recommended to devise and undertake such work when resources are available.

13.5.2. *Over sea*

Gordon has examined the relation between low-level winds and winds at 2,000 and 4,000 feet observed at sea. The surface isobars over the sea can seldom be determined with sufficient accuracy for use in investigation but the forecasting of surface winds must of course depend on a forecast of the isobaric distribution. By comparing winds observed at various heights above the sea surface Gordon[16] obtained results which are useful in estimating the ratio of the surface to the geostrophic wind.

The mean value of the ratio of the wind speed observed by anemometer to the wind speed at 2,000 feet observed (or interpolated) on upper wind soundings from ocean weather ships in the North Atlantic was 0·71. In another paper Gordon[17] computed the variation in the ratio of the wind speed at 50 feet to the wind speed at 2,000 feet for various lapse rates. His results are shown in Figure 10.

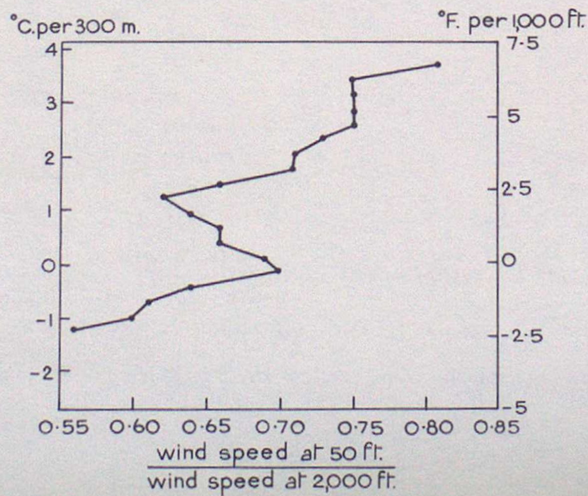


FIGURE 10 *Variation between the ratio* $\dfrac{\textit{wind speed at 50 feet}}{\textit{wind speed at 2,000 feet}}$ *and lapse rate*

Forecasters should note that the lapse rate used by Gordon was obtained in the following way:

$$\frac{\text{Temperature at surface minus temperature at 2,000 feet,}}{2}$$

that is, it is a mean value over the lowest 2,000 feet of the atmosphere and this is the value which forecasters must determine or estimate if the results are used in practical forecasting.

It is seen in Figure 10 that, from a minimum of 0·56 in strong inversion conditions, the ratio increased as the strength of the inversion decreased. A value of 0·75 was reached when the lapse rate was close to the dry adiabatic rate.

Figure 11 shows the variation of the ratio with the wind speed at 2,000 feet and indicates that there was a fall in the ratio as the speed increased from the range 10 to 19 knots to the range 20 to 29 knots. The ratio remained within the range 0·65 to 0·70 for further increases in wind speed.

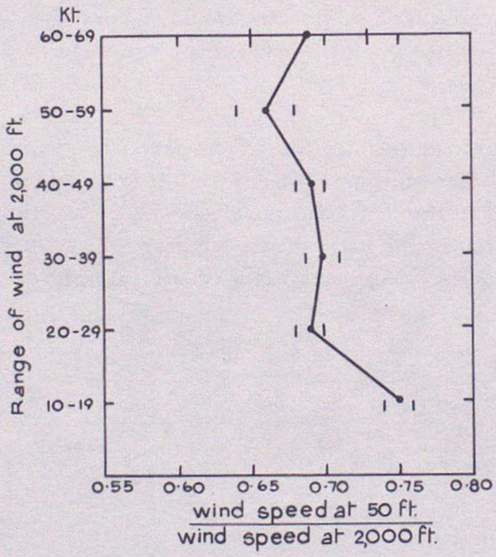


FIGURE 11 *Variation of the ratio* $\frac{\textit{wind speed at 50 feet}}{\textit{wind speed at 2,000 feet}}$ *for certain ranges of the wind speed at 2,000 feet*

The distance between the short vertical lines is twice the standard deviation of the mean at each point.

Information on the direction of the wind at 50 feet in relation to the direction of the wind at 2,000 feet for various lapse rates was obtained by Gordon[18] and is reproduced in Table 16.

Wind

TABLE 16 *Angle by which the wind at 50 feet is backed from the wind at 2,000 feet as a function of lapse rate*

	Lapse rate		*Mean angle*	*No. of obs.*	*Notes*
	°F. per 2000 ft.	*°C. per 600m. approx.*	*degrees*		
	−9 to 0	−5 to 0	17.4	64	Isothermal or inversion
	1 to 3	.5 to 1.5	15.9	99	-
	4 to 6	2 to 3.5	11.8	178	Approx. saturated adiabatic
	7 to 9	4 to 5	7.2	237	-
	⩾10	⩾5.5	7.0	121	Approx. dry adiabatic
Total				699	
Mean			10.5		

Table 16 shows that the mean angle decreases fairly steadily with increasing lapse rate, decreasing from about 17° in inversion or isothermal conditions to about 7° in conditions of marked instability. Gordon considered that the decrease levelled off for lapse rates greater than the dry adiabatic. The scatter of the individual observations was large but standard deviations or extremes are not available.

Gordon[16] also obtained some data on the diurnal variation during daylight hours of the ratio between the low-level wind speed and that at 2,000 feet. In this investigation data from H.M. Ships when in Marsden square 146 (that is, between 40° to 50°N. and 10° to 20°W.) were used. The height of the anemometer above the sea surface might have varied according to the vessel but it was probably of the order of 100 feet in most cases. The results are given in Table 17 obtained when the mean values were grouped at intervals between 0700 and 1900 local time.

TABLE 17 *Variation of the ratio* $\dfrac{\textit{wind speed at 100 feet}}{\textit{wind speed at 2,000 feet}}$ *during daylight hours*

	local time					
	0700-0800	*0900-1000*	*1100-1200*	*1300-1400*	*1500-1600*	*1700-1800*
$\dfrac{\text{wind speed at 100 feet}}{\text{wind speed at 2,000 feet}}$	0.92	0.99	1.01	0.97	0.82	0.82

If it is assumed that the wind speed at 2,000 feet has no diurnal variation, then Table 17 indicates that the wind speed at 100 feet during daylight hours has a maximum around noon when it is very close to the wind speed at 2,000 feet. In the mid afternoon and early evening there is a noticeable decrease to about four fifths. Information is not available for the hours of darkness.

The ratio of wind speed at 100 feet to wind speed at 2,000 feet is rather higher than the normally accepted ratio of wind speed at surface (or 50 feet) to wind speed at 2,000 feet. Gordon thought that, in this investigation, this might be due to the fact that wind speeds were measured at about 100 feet and that the days on which the observations of upper winds were available were limited to days of good visibility when the height of the base of low cloud was not below 2,000 feet.

13.5.3. *The diurnal variation of wind near the ground*

The diurnal variation of wind in the lower layers of the atmosphere can be explained by the diurnal variation in temperature distribution. During the day incoming radiation frequently establishes in the lower layer a lapse of temperature of sufficient magnitude to create unstable conditions. This instability causes the lower layer to be mixed vertically. Consequently the momentum of the wind from any one layer tends to be mixed into adjacent layers through the influence of this vertical stirring. In this way vertical variations in wind velocity are reduced and the vertical stirring leads to fairly smooth changes through the mixed layer. Near the ground, wind energy is destroyed by the turbulent flow due to friction from the ground so that the lower layers, even at the time of greatest vertical stirring, tend to have lower wind speeds than layers a few hundred feet above the ground. At night the lower layers of the atmosphere normally cool under the influence of outgoing long-wave radiation (see Chapter 14) and a stable distribution of temperature is established. This stability inhibits vertical mixing. In these conditions there is very little vertical mixing and consequently only very small amounts of momentum are transferred to the layer near the ground from levels above the stable layer. Thus the energy of the wind which is destroyed by frictional effects near the ground is no longer partially made good by the downward transfer of momentum from upper levels. Consequently surface winds decrease. As momentum is not transferred from upper levels immediately above the stable layer, wind speeds at about 1,000 to 2,000 feet above the ground tend to be a maximum when the lower wind speeds are a minimum and *vice versa.*

The depths of the vertical layer in which there is a diurnal variation of wind due to diurnal variations in the lapse rate near the ground will vary with the season of the year and with the synoptic situation. It is not practicable to give mean values but estimates can usually be made by considering the probable variation of temperature near the ground (see Chapter 14).

Some detailed information on the vertical diurnal variation of wind at specific sites has been obtained in research investigations and reported by a number of workers.[19-21] Summaries of these papers are not included as, in general, the numerical results seemed unlikely to be of direct value in forecasting. However, in an investigation of the lower layers of the atmosphere when radiation fog was likely at Cardington, Stewart[22] obtained some measurements of wind in the lowest

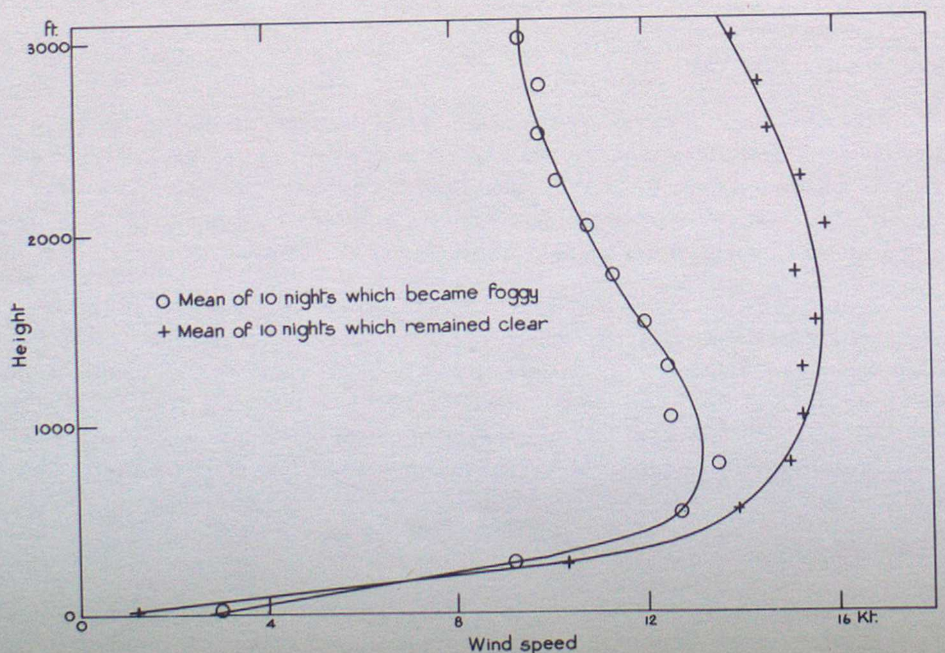


FIGURE 12 *Profile of wind speed at Cardington at 1800 G.M.T. on clear evenings*

3,000 feet of the atmosphere. Some mean values observed at 1800 G.M.T. on clear evenings are shown in Figure 12. The curves show a difference in wind structure above 750 feet between nights on which fog subsequently formed and those on which it did not. The figure has been included, however, primarily to illustrate the profile of wind near the ground on radiation nights and, in this layer, there are only small differences between the foggy and non-foggy nights. There was much variation in behaviour from night to night so that Figure 12 should be interpreted as giving only an approximate profile of wind speed.

13.5.4. *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 recording gauges of some cup-type anemometers yield 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 then 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 and others[15] showed, however, that gusts were very heterogeneous. The time interval between gust and ensuing lull was very varied. Giblett and others 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 200 feet 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.

A selection of typical traces recorded at Cardington is shown in Figure 13. These illustrate the different types of eddies which are found on normal anemograms under differing conditions of the lapse rate.

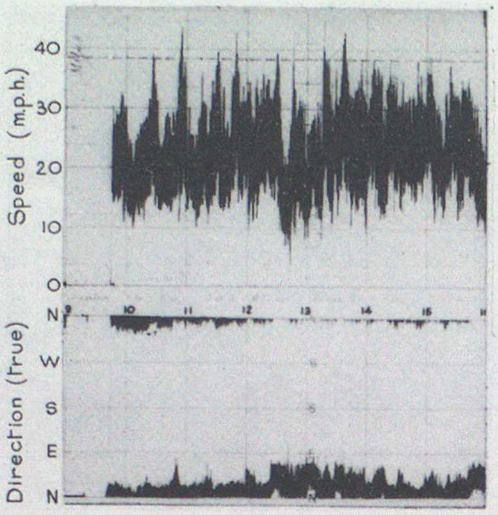
As stated earlier it is not possible to forecast individual gusts but it is possible to indicate the probable value of maximum gusts which may occur. These values may be related to the mean hourly wind or the geostrophic wind at 2,000 feet and the temperature lapse. Table 18 contains statistics based on Giblett's paper and should be of value.

TABLE 18 *Frequency of the ratios of maximum gust to mean wind at Cardington*

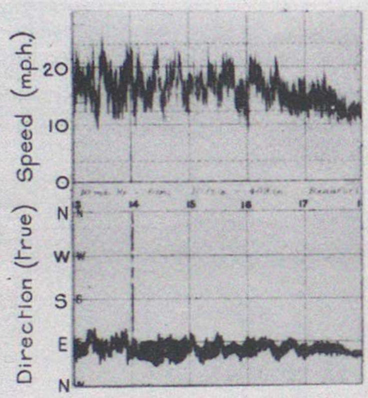
Time G.M.T.	*Wind speed at 150 ft.*	*Ratio* 1.2	1.3	1.4	1.5	1.6	1.7	1.8	1.9	2.0	2.1	2.2	2.3
	mi.hr.$^{-1}$												
						Height 50 ft. - June 1929							
0100 to 0300	10 - 19	2	4	8	13	11	6	1	1	..	..	..	..
	20 - 29	..	..	2	1	..	..	1	..	..	..	..	..
1300 to 1500	10 - 19	..	2	1	7	16	7	12	6	2	2	..	..
	20 - 29	..	..	1	9	6	2	..	..	..	..	..	..
	30 - 39	..	..	1	1	1	..	..	..	..	..	..	..
						Height 50 ft. - December 1929							
0100 to 0300	< 10	1	..	2	..	2	..	..	1	1	..	..	..
	10 - 19	..	3	4	5	2	4	1	..	1	..	..	..
	20 - 29	..	..	7	14	2	1	1	..	..	..	..	..
	30 - 39	..	..	2	6	9	1	..	..	..	..	..	..
	40 - 49	..	..	..	1	..	..	..	..	..	..	..	..
	50 - 59	..	..	1	2	..	..	..	..	..	..	..	..
1300 to 1500	10 - 19	1	1	8	10	11	1	3	1	..	..	..	1
	20 - 29	..	..	2	7	8	2	..	1	..	..	..	..
	30 - 39	..	..	2	6	2	1	1	..	..	..	..	2
	40 - 49	..	..	..	1	1	3	..	..	..	..	..	..
	50 - 59	..	..	..	1	1	..	..	..	..	..	..	..
						Height 150 ft. - June 1929							
0100 to 0300	10 - 19	8	6	10	12	6	2	1	..	..	..	..	..
	20 - 29	..	..	..	1	3	..	..	..	..	..	..	..
1300 to 1500	10 - 19	..	..	3	9	7	12	8	5	5	3	2	1
	20 - 29	..	1	3	5	3	4	1	1	..	..	..	..
	30 - 39	..	..	..	2	1	..	..	..	..	..	..	..
						Height 150 ft. - December 1929							
0100 to 0300	< 10	..	..	..	..	..	..	..	..	..	..	..	..
	10 - 19	2	2	7	3	2	3	1	..	..	..	..	..
	20 - 29	..	1	10	7	4	1	2	..	..	..	..	..
	30 - 39	..	..	8	9	1	..	..	..	..	..	..	..
	40 - 49	..	..	1	1	..	..	..	..	..	..	..	..
	50 - 59	..	..	3	..	..	..	..	..	..	..	..	..
1300 to 1500	10 - 19	..	1	6	9	10	6	2	..	1	..	1	..
	20 - 29	..	..	4	7	4	2	1	1	..	..	..	1
	30 - 39	..	..	..	4	7	1	1	1	..	..	..	..
	40 - 49	..	..	1	1	1	1	1	..	..	..	..	..
	50 - 59	..	..	2	..	..	..	..	..	..	..	..	..

The general conclusion to be drawn from Table 18 is that the gusts at 50 feet have a greater ratio to the mean speed by day than those at 150 feet but a smaller ratio

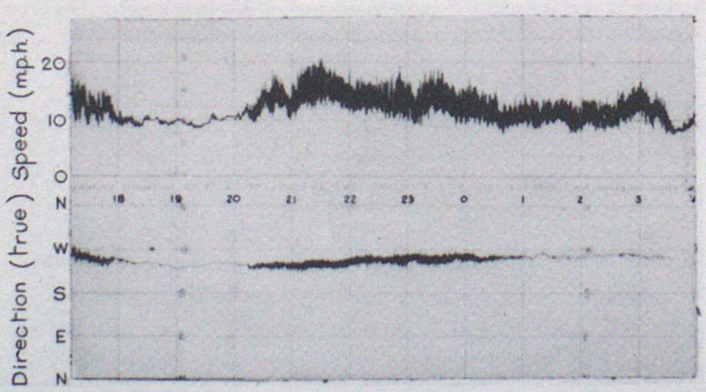
Wind



(a) Typical gustiness during a period of deep convection. During the period 1000 to 1500 G.M.T. cumulonimbus cloud was present and convection was taking place through a depth of several thousands of feet in the lower troposphere. High gusts were spaced out at comparatively wide intervals but the whole trace was irregular both in direction and speed.



(b) Typical gustiness during a period of limited convection. During the period 1500 to 1700 G.M.T. instability was taking place in the lower 2000 to 3000 feet or so and the vertical temperature gradient was generally approximately adiabatic. Stratocumulus cloud was often present. The major fluctuations in speed and direction were still considerable but are much less than in Figure 13*(a)*.



(c) Typical gustiness (or absence) in stable (or inversion) conditions. During the period 2100 to 0300 G.M.T. the vertical temperature gradient was less than the adiabatic and might show an inversion. The trace on the anemogram was relatively broad. The fluctuations appeared to be very rapid, more so than with either of the types illustrated in Figures 13*(a)* and 13*(b)*. During the period 1830 to 2000 G.M.T. there was a decided inversion of temperature in the lowest 150 feet of the atmosphere. Fluctuations had almost entirely disappeared from the trace.

FIGURE 13 *Examples of anemograph traces illustrating gustiness* by night. Generally the maximum gust to the mean wind lies within the ratios 1·4 to 1·6 but occasionally the ratio may exceed 2. The higher gusts are more likely to occur on a day with unstable air, vigorous convection clouds and some showers. The lower values are likely to occur with stable lapse rates and at night.

Gold[23] has 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 19.

TABLE 19 *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

For coastal stations the ratios were mainly between 1·4 and 1·5; inland country stations between 1·6 and 1·8 but for inland town exposures the ratios were mainly between 1·9 and 2·1.

The seasonal and diurnal variation of the ratio of the maximum gust at 150 feet to the geostrophic wind at Cardington is apparent in Table 20. The figures in Table 20 refer to all geostrophic wind speeds but there are important differences for different ranges of wind speed. For strong geostrophic winds (say ⩾ 35 knots) the geostrophic wind provides a close estimate of the maximum gust but for lighter winds, especially in summer and by day, the gusts may be greatly in excess of the geostrophic speed. Some further statistics are given in Table 21.

TABLE 21 *Frequency of occasions on which the ratio between the maximum gust at 150 feet and the geostrophic wind fell within certain values at Cardington*

Geostrophic wind speed	*Ratio of maximum gust to geostrophic wind* 0.5	0.6	0.7	0.8	0.9	1.0	1.1	1.2	1.3	1.4	1.5	1.6	1.7	1.8
kt.	*Number of occasions*													
26 - 34	14	14	28	26	22	26	18	17	9	5	2	2	0	1
35 - 43	9	14	12	14	20	13	5	5	2	3	0	0	0	0
44 - 51	8	12	10	10	8	6	3	0	0	1	0	0	0	0
52 - 60	4	2	11	8	4	2	1	0	0	0	0	0	0	0
⩾ 60	6	10	12	6	3	0	0	0	0	0	0	0	0	0
Total	41	52	73	64	57	47	27	22	11	9	2	2	0	1
	per cent													
	10	13	18	16	14	11	7	5	3	2	0.5	0.5	0	0

13.5.5. *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.

TABLE 20 *Frequency of the ratio of maximum gusts at 150 feet to geostrophic wind speed at Cardington*

Months	Time G.M.T.	<0.5	0.5	0.6	0.7	0.8	0.9	1.0	1.1	1.2	1.3	1.4	1.5.	1.6	1.7	1.8	1.9	2.0	>2.0	Indeterminate
		Ratio of maximum gust to geostrophic wind speed																		
		number of occasions																		
December, January, February	0100	10	11	13	14	16	4	7	5	4	1	2	1	..	..	..	..	..	..	2
	0700	3	2	13	19	11	12	5	7	4	5	1	..	3	..	..	..	..	2	3
	1300	2	..	12	8	7	13	13	7	6	5	6	3	3	1	1	..	..	..	3
	1800	6	4	7	10	16	8	11	8	7	4	4	2	1	..	1	..	..	..	1
March, April, May	0100	1	6	13	12	16	3	8	4	5	1	2	2	1	1	..	..	..	2	13
	0700	8	2	3	10	9	22	8	3	7	3	1	3	1	..	..	1	..	..	9
	1300	..	3	3	4	9	5	3	15	2	6	3	6	6	3	2	1	2	12	7
	1800	1	3	3	14	4	9	9	7	6	5	..	5	6	3	2	1	1	5	8
June, July, August	0100	8	4	9	15	6	12	8	5	4	3	3	1	..	1	1	..	1	..	11
	0700	6	3	2	10	6	11	6	7	11	7	5	5	3	2	1	..	..	..	7
	1300	..	1	2	..	2	7	5	6	10	6	8	6	5	7	1	5	4	10	7
	1800	2	..	1	1	5	5	12	7	9	5	3	4	4	2	6	3	1	11	10
September, October, November	0100	5	9	11	14	10	13	6	4	1	4	2	2	..	1	..	..	1	..	6
	0700	9	14	6	12	16	12	9	3	1	3	..	..	2	..	..	..	..	..	4
	1300	1	2	5	6	5	16	8	7	8	3	6	4	3	4	2	1	1	4	4
	1800	6	2	10	10	7	10	13	5	5	4	2	3	2	..	1	1	..	3	7

13.6. 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.

13.6.1. *Sea-breeze*

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. It is clear that the surface sea-breeze must be regarded as one leg of a relatively small local-wind circulation consisting of inflowing air at sea level, an ascending current over the land, outflowing air at a height of a few hundred feet and a descending column of air over the sea. It should be noted that the sea-breeze commences to flow down the pressure gradient and approximately at right-angles to a flat coastline. As the sea-breeze continues to flow the effect of the rotation of the earth (the Coriolis force) causes a veering of the wind so that with increasing time the sea-breeze tends to blow at larger angles to the thermally-induced pressure gradient and by evening may be blowing almost parallel to the coast. It is difficult to give a value for the vertical extent of the sea-breeze but in this country it is probably in the order of a few hundred feet. The upper current from land to sea is usually much lighter (often barely detectable) and is spread over a depth of several thousand feet.

It was convenient to presuppose calm conditions in this simplified description of the sea-breeze. In most cases there is a general wind due to the general pressure distribution and to obtain the actual wind this general wind must be compounded vectorially with the sea-breeze (and any other "local" winds which may be blowing). It is quite clear that with a general onshore wind the effect of the sea-breeze may be scarcely noticeable - perhaps an almost imperceptible increase in wind speed. On the other hand a sufficiently strong offshore wind may completely inhibit the sea-breeze. When, however, a sea-breeze replaces a land-breeze, there is often a noticeable change in the air mass. On a hot day the sea-breeze may bring a marked moderation in temperature and an increase in bodily comfort or there may be important differences in visibility. It is not therefore purely an academic matter to forecast correctly the onset of the sea-breeze. The forecasting of sea-breezes is considered in Section 13.9.

Forecasters who wish to study the theoretical aspects of the sea-breeze may consult references 24 to 30.

13.6.2. *Land-breeze*

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. At night 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. A very moderate onshore component of the gradient wind will completely inhibit the land-breeze. However, in some coastal areas with high ground situated only a few miles 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[31] has 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 14, 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 "the isopleths of the magnitude of the land-breeze over the sea are roughly a "mirror" of those over land."

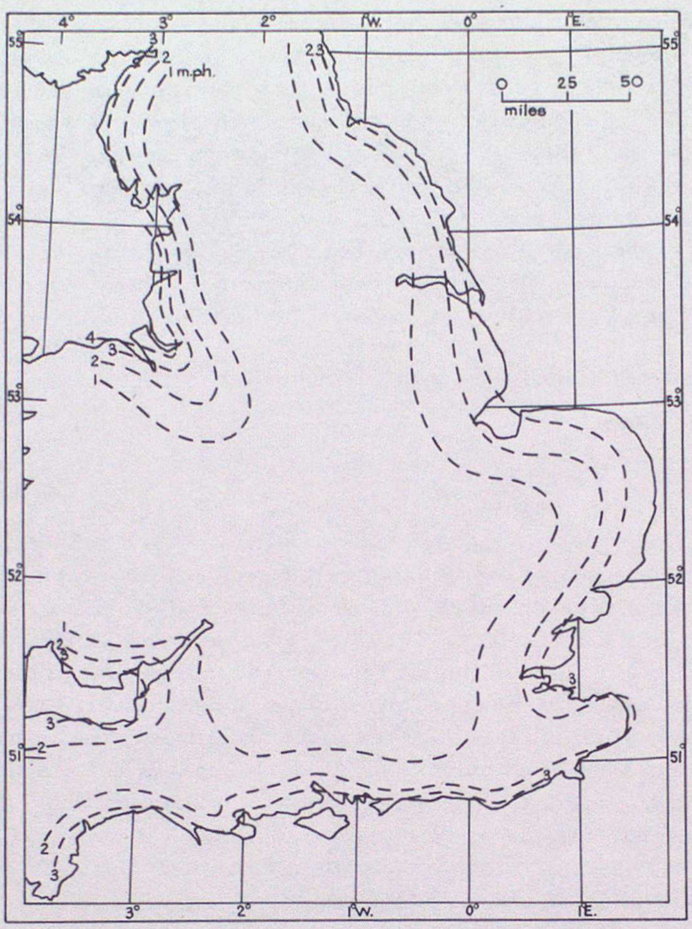


FIGURE 14 *Lines of equal magnitude of land-breeze (March, April, May)*

13.6.3. *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 and 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 15.

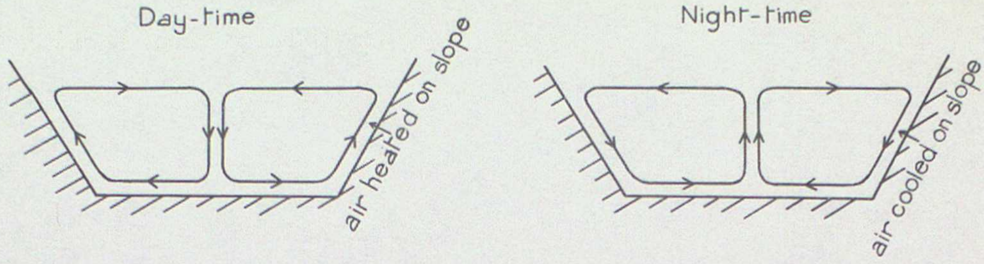


FIGURE 15 *Schematic illustration of local circulations due to heated and cooled slopes*

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 15. 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 (see reference 24 for a number of original papers on this topic). Defant[32] has produced a series of schematic diagrams illustrating both these types of winds. These are reproduced in Figure 16.

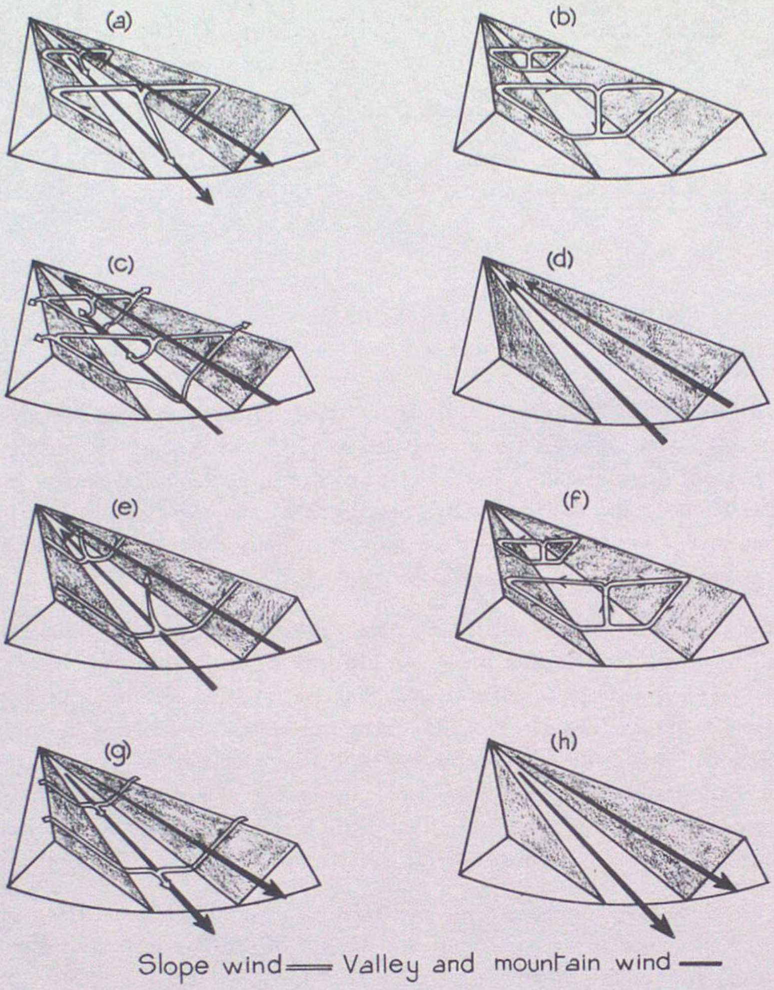


FIGURE 16 *Schematic illustration of the normal diurnal variations of the air currents in a valley (after Defant[32])*

(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.

Mountain and valley winds occur most frequently during persistent high pressure situations, notably in summer. They are thus typical fair-weather phenomena. The down-slope wind occurs frequently on clear radiation nights in winter also. These winds can, however, also occur in cloudy weather but they are then generally feeble in character and cause only a minor modification to the general wind.

Lawrence[31] has 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 17. These curves satisfy the following formula:

$$\text{Speed of down-slope wind} = \sqrt{\left(\frac{2 \times \text{distance of donor site from the sea}}{1{\cdot}4 + 100 \tan\theta}\right)}$$

where θ is the angle of slope of the ground.

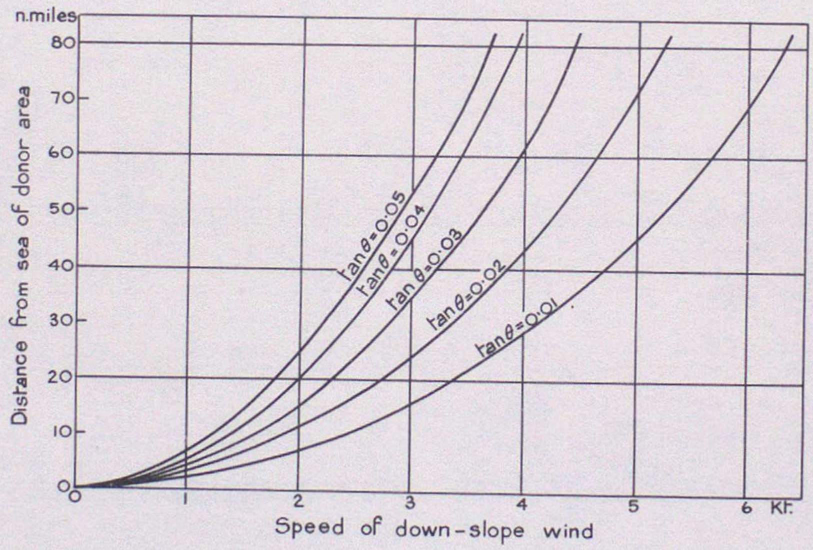


FIGURE 17 *Relation of speed of down-slope wind (on a broad slope) to distance from sea and angle of slope of land*

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.

13.6.4. *Effect of topography*

Subsections 13.6.1, 13.6.2 and 13.6.3 have described effects of topography on wind which arise through horizontal differences in temperature. However, when moderate or strong winds blow across hilly country the simple mechanical deflection of the wind by the ground is more important and is best studied by a systematic comparison of observed surface winds with the geostrophic wind. The effects are complicated and depend not only on the wind in the lowest 1,000 feet or so but

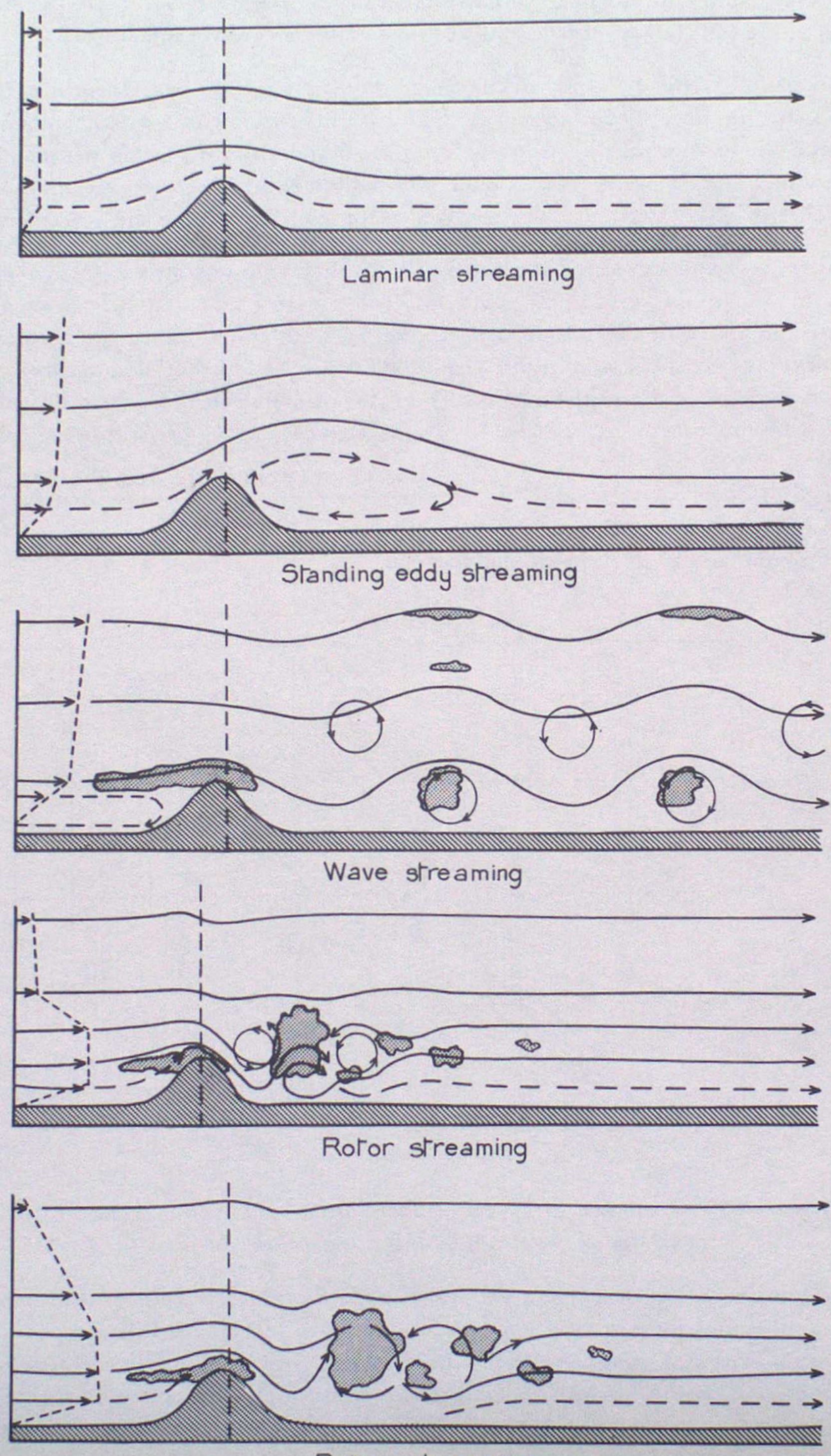


FIGURE 18 *Types of air flow over ridges (after Förchtgott[34])*

The variation of the wind speed with height is indicated by the wind profile at the left in each case.

upon the variation of wind and static stability up to heights of 10,000 feet and above. The flow of air over hills has been studied both theoretically and observationally mainly in regard to the vertical currents produced and their effect on aviation, but the results also give some idea of the distribution of surface wind which may be expected. Reference may be made to a summary by Corby[33] in which diagrams of typical flow patterns are quoted from Förchtgott[34] and are again reproduced here in Figure 18. On encountering a hill the air may break away from the ground and leave a stationary eddy with light winds in the lee (sheltering by the hill from the wind) but on other occasions, particularly when lee waves are formed in the free atmosphere, the wind down the lee slope may be intensified and be stronger than over level country. The difference in behaviour depends on the airstream characteristics and the reader is referred to *Meteorological Reports* No. 18[33] for further information. Strong winds on lee slopes are particularly likely at night when there is an inversion of temperature or a stable layer near the ground.

Strong winds in valleys may also arise as "gap" winds in specific areas (mainly in larger mountain masses than occur in the British Isles). These arise when a mountain range separates air masses which have markedly different temperatures in the levels below the mountain tops. Pressure is approximately the same level for level in the two air masses above the crest level of the ridge but below this the colder air mass has higher pressure than the warmer. If a suitable valley traverses the mountain range a pressure difference of several millibars may exist between the ends of the valley and air is accelerated down the pressure gradient, leading to strong winds along the valley which under particularly suitable conditions may reach gale force. The "Bora" in the northern Adriatic is an example of a wind set up in this manner. References 35 and 36 may be consulted by forecasters who wish to make a closer study of this type of wind.

The configurations of topography, when considered in combination with synoptic types, are so numerous that detailed consideration of each combination is not practicable in this handbook. Each site should be considered on its merits and deductions should be made about the probable abnormalities of surface wind in the area. Wherever possible, past records of wind observations should be scrutinized and analysed so that the abnormalities of surface winds can be related to the topography and the synoptic situation. Local studies of this nature are often of great value to forecasters.

13.7. STATISTICS AND OBSERVATIONAL DATA ON UPPER WINDS

The short-term variability of surface winds has been considered in Section 13.5 which contains quantitative values based on a number of observations. Some of these observations were of a routine character using standard instruments. Other observations were made for specific researches and employed special anemometers. Instrumental observations of the short-period variability of upper winds are more difficult to obtain and the amount of quantitative data available is far smaller than that for surface winds.

Actual winds are always fluctuating. The larger eddies (for example, circulations around well marked depressions or anticyclones) take many hours to pass a station and their effect on the upper wind structure is normally detected by the synoptic network and allowance for them can be made by orthodox forecasting techniques. Rather smaller eddies may be revealed by a dense network but there are always very much smaller-scale eddies which cause turbulent wind fluctuations on a short time and distance scale - say a period measured in minutes and over distances of a few miles. It is never possible to extract direct reliable information on any one of these from normal synoptic reports. Although no direct account of them can be taken in normal forecasting it is useful if forecasters are aware of some broad estimates of these fluctuations in upper winds.

Durst[37] has examined a number of diverse observations of the short-period variation of upper winds over England. The individual results are rather too detailed for inclusion in this handbook but Figure 19, taken from Durst's work, presents the gist of the results in a concise and concentrated form. The reader who wishes to examine the data and reasoning on which Figure 19 is based should consult the original paper.[37]

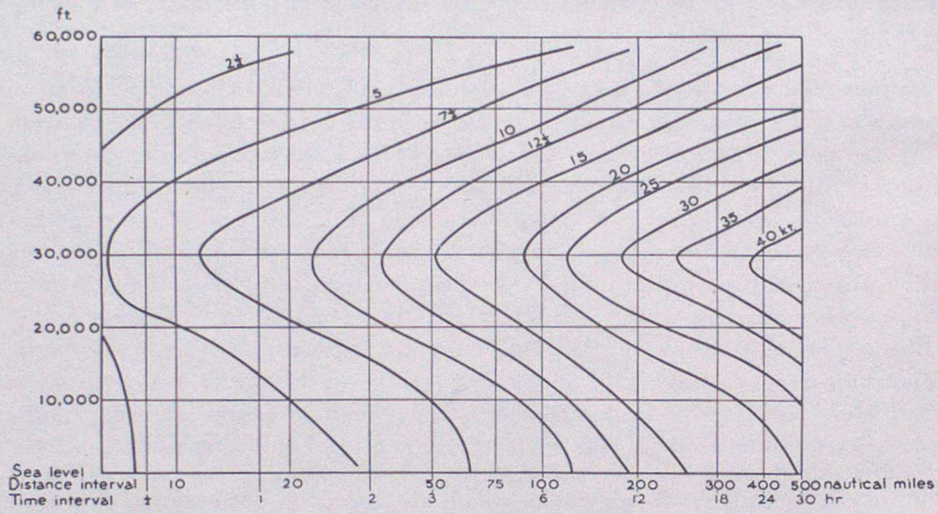


FIGURE 19 *Probable vector variation of wind with time and place over England (after Durst[37])*

No very high degree of accuracy can be claimed for Figure 19 but it certainly shows the general trend of variation of the 50 per cent errors which will arise by the assumption that a wind measured at one time is a forecast of a wind at a later time.

Upper winds are affected by orographic features. The larger masses of high ground may produce distortions in the air flow which are recognizable on the synoptic scale, for example, lee depressions. On some occasions smaller orographic features cause variations in the winds on a scale which is too small to be revealed by normal methods of analysis of routine observations. Normally nothing can be done to allow for this in the analysis and prognosis of horizontal (isobaric) winds. The vertical components of these disturbed winds sometimes become organized into a system of "standing" waves. These vertical air currents are of importance to aviation and are more fully discussed in Chapter 20 - Turbulence (as affecting aircraft). The reader may also consult the summary of air flow over mountains, prepared by Corby.[33]

Some comments on various aspects of the accuracy of forecasts of upper winds and of the relation between upper winds and contours (or prontours) are included in Sections 13.8 and 13.10.

Wind

13.8. FORECASTING WINDS - GENERAL

The established method of forecasting winds, as practised in the United Kingdom, rests primarily on the measurement of geostrophic (or gradient) winds from isobars or contours on forecast charts. Considerable interpolation is often necessary and some extrapolation may also be required in time and space. Further, when observational data become available which are more recent than that on which the forecasts were prepared, some adjustment to the values measured from the forecast charts may be justifiable. These adjustments must often be made in an *ad hoc* manner since time seldom permits the reconstruction of fresh prebaratic/prontour charts. Care must be taken in assessing the amount by which wind velocities, measured from forecast charts based on careful, methodical and reasonably leisurely consideration of a sequence of charts, should be varied on receipt of a fresh, single observation which seems to throw doubt upon the validity of the formerly expected developments.

Experience indicates that "panic" amendments to winds sometimes lead to errors subsequently shown to be at least as great as those which would have been made if the original forecast had remained unaltered. Where the upper air observing network is very open - as over the sea - amending action may have to be taken on a single subsequent observation, but even in these regions forecasters should scan all other recent data to see whether any supporting evidence to justify the amendment can be found. When surface observations show that a rapid development which was not allowed for in the prebaratic is occurring, it is generally fairly easy to estimate the changes to the prontours for periods of up to about 12 hours but, for longer periods, the problem is usually more difficult. This arises from a number of reasons. For example: Will a sudden fall (or surge) of pressure be maintained and if so, in what way will the thermal pattern be distorted, strengthened or weakened? Will there be a discontinuous change in direction and speed of movement of the system and may the sudden change be the herald of a major change in the synoptic type?

These problems are not peculiar to wind forecasting since they affect the whole forecast. They are mentioned here as a warning to forecasters against making too precipitate amendments to forecasts on the basis of an ill-considered and inadequate assessment of an isolated observation. However, in a region with a dense network of observations, other reports, close in time or space, will usually corroborate or shed doubt on the validity of the single observation and the inferences drawn from it. In such regions little is normally lost by waiting for the next observation but it may well be desirable to issue some precautionary warning to recipients of forecasts which may be in error. On the other hand where there is an open spatial and temporal observing network the forecaster may have no reasonable expectation of further observational data in the near future and cannot afford for operational reasons to defer a decision. He must make a judgement on whether to accept the observation at its face value and introduce a major change in the forecast. There is no invariant rule of thumb for this and experience is a principal guide. In all cases such isolated observations should be most carefully scrutinized, for example, forecasters should take the elementary precaution of checking the accuracy of the plotting and the accuracy of transmission (that is, with the version received at a neighbouring station, at the teleprinter centre, or at the meteorological communications centre). In addition the observation(s) should be checked, as far as possible, for consistency, for example: Are pressure and tendency in reasonable agreement? Is a ship plotted in the correct location, that is, in relation to earlier reports from the ship? Does the upper air sounding (rawin- and/or radio-sonde) lead to a plausible vertical and thermal structure and is it a possible - even if unexpected - development from recent analyses, making due allowance for uncertainties of analysis and for continuity? Whenever possible the isolated observation and its interpretation should be discussed with the superior forecasting centre before taking major amending action.

The whole concept of forecasting winds from isopleths on constant-level or constant-pressure charts rests on the assumption that the winds and the pressures are in balance (either geostrophic or gradient). The ageostrophic winds are of vital importance for development but there is no currently available, practical, comprehensive method for assessing them quantitatively on the forecast bench.

A method of assessing the horizontal component of the isallobaric wind at low levels was given in Section 13.3. No practical technique is available for the computation of ageostrophic winds at upper levels. In regions of well marked divergence or convergence it may be possible to deduce from the observations that the wind flow across the isopleths amounts to 5 or 10 knots and some ageostrophic effects are implicitly included in very short-period forecasts of winds which are made by modifying the winds deduced from the contour charts by means of the latest available, observed, upper winds.

Although orthodox forecasting of winds, as practised in the United Kingdom, rests primarily on forecast charts, much reliance is placed on extrapolation and modification of recent actuals for short-period forecasts - say 6 to 12 hours ahead. This arises partly from the fact that forecast charts are usually made up for a time rather farther ahead (that is, around 24 hours) but also from the fact that for short periods persistence is often a good forecast of the upper wind. For 6 hours ahead of the observations this is true generally at all levels above about 700 millibars, but some variation would be justified when marked changes of the wind régime were clearly indicated (for example, frontal passage, upper troughs). For

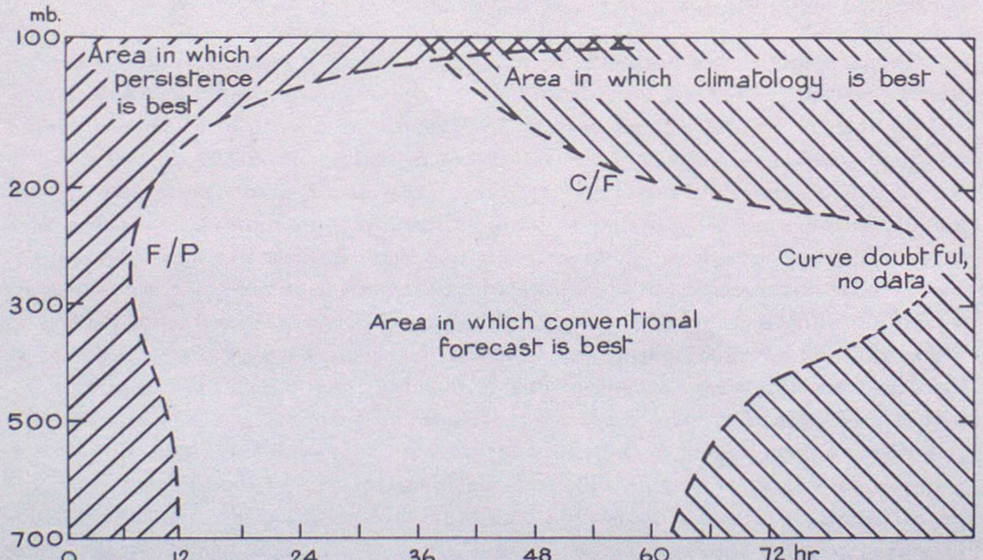


FIGURE 20 *Map of time - altitude plane divided into areas in which contour-technique forecasts, persistence and climatology can be expected to give the best estimate of the future wind*

F/P = time after which contour-technique forecasts become better than persistence

C/F = time after which climatology becomes better than contour-technique forecasts

12 hours ahead orthodox forecasting is probably superior to persistence. For long periods ahead, errors in orthodox forecasting become so great that a climatological wind becomes a better forecast, that is, one with a smaller error. It is not easy to estimate the duration of forecast beyond which climatology improves on contour-chart technique but a limit between 2 and 3 days would probably be appropriate; it might be greater in very persistent synoptic situations. Ellsaesser[38] has surveyed a number of papers on the accuracy of upper wind forecasts and has produced a diagram which indicates the variation with altitude and time of the relative accuracy between wind forecasts based on persistence, contour-chart forecasting and climatology. Figure 20 shows the results for areas near the British Isles. The diagram is very tentative and forecasters should not interpret it too rigidly.

13.9. FORECASTING THE SURFACE WIND

The geostrophic wind measured from isobars on a surface chart or contours on the 1000-millibar chart provides, for most synoptic situations, a close estimate of the mean wind in the free atmosphere at a height of about 2,000 feet above level ground which is near mean sea level. With strong isobaric curvature or large pressure changes, allowance for the curvature of the isopleths or the isallobaric component of the wind may improve the estimate of the wind at 2,000 feet. The surface wind is then forecast by modifying both direction and speed to take account of surface friction and topographic effects.

Over land topography often exerts a marked effect on surface winds, both in direction and speed. No treatment of direct practical value to forecasters could possibly be exhaustive and include all possible topographies. It is strongly recommended that, wherever possible, local investigations of past records be undertaken with the object of determining in a factual and systematic way the peculiarities of the surface winds in areas or districts for which forecasts are issued.

13.9.1. *Over land*

The deviation of surface winds from their geostrophic value is indicated for a number of sites by the numerical values included in Section 13.5.1. It is quite clear from those values 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 value and backed by 30°. During strong convection, speeds might be increased to one half of the geostrophic value and the backing reduced to 20 to 25°. At night with clear skies winds will fall below one third of the geostrophic wind and the backing may be 40° or more. Much greater variations may be caused by katabatic effects on clear nights and by sea-breezes on some days. (Sea- and land-breezes are considered in Sections 13.9.7 and 13.9.8.)

The above suggestions are broad-scale values. Nevertheless if the rules are applied systematically and modified by amounts within the limits implied by the various statistics in Section 13.5.1, the forecaster should soon gain useful experience and a practical working knowledge of the behaviour of the surface wind in the forecast locality for a number of synoptic patterns. Such experience should normally enable the forecaster to estimate and judge with fair accuracy the factor to apply to geostrophic values to produce good forecasts of the surface winds for the more common synoptic situations.

13.9.2. *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 Section 13.5.2 will normally enable a good forecast of the surface wind to be made. It should be noted, however, that over sea areas from which meteorological reports are sparse the determination of the actual isobaric pattern may be rather more imprecise than over land. Errors in the isobaric patterns will result in errors of rather similar magnitude in estimates of the surface wind.

It is suggested that, in stable conditions, surface winds will be about two thirds of the geostrophic value and backed by some 15 to 20°. In unstable conditions winds will be backed about 10° and speeds will be about four fifths of the geostrophic value.

13.9.3. *Diurnal variation*

Some account of this has been taken in the suggested rules given in Sections 13.9.1 and 13.9.2. It is very difficult to quote representative values for the diurnal variation of surface winds which can be applied directly to forecasting. Nevertheless the statistics on diurnal variation which are included in several tables in Section 13.5 should form a useful basis for forecasting. It is stressed that the average values determined for any one site should be applied to other sites with great caution. In addition diurnal variations may depart widely from the mean in conditions of marked thermal stability or instability in the lower layers.

13.9.4. *Gales*

The forecasting of gales is primarily a problem for the prebaratic, namely, the correct forecasting of pressure gradients. Given a correct prebaratic forecasters will normally have little difficulty in deciding when gales should be forecast and gale warnings issued.

Where very large positive or negative pressure tendencies (for example, 10 millibars per 3 hours) are reported from a fairly limited area (for example, 200 miles or so in horizontal dimensions) and seem likely to be maintained for a few hours the possibility of gales should be most carefully considered. Such large tendencies can produce large changes in pressure gradients and may result in gales developing quite quickly in areas where the actual or previously expected winds were relatively light.

13.9.5. *Gusts*

Section 13.5.3 described the variability and short-lived characteristics of gusts and it is quite clear that it is impossible to forecast the speed and direction of individual gusts. Nevertheless the values included in Tables 18 to 21 of Section 13.5.3 should enable forecasters to make reasonable estimates of the probable upper limit to gusts which may occur. When applying these values in forecasting it should be noted that Tables 18 and 19 relate the maximum gust to the mean wind at anemometer height and that Tables 20 and 21 relate the maximum gusts to the geostrophic wind.

13.9.6. *Squalls*

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 16).

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 miles 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 10,000 to 15,000 feet 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

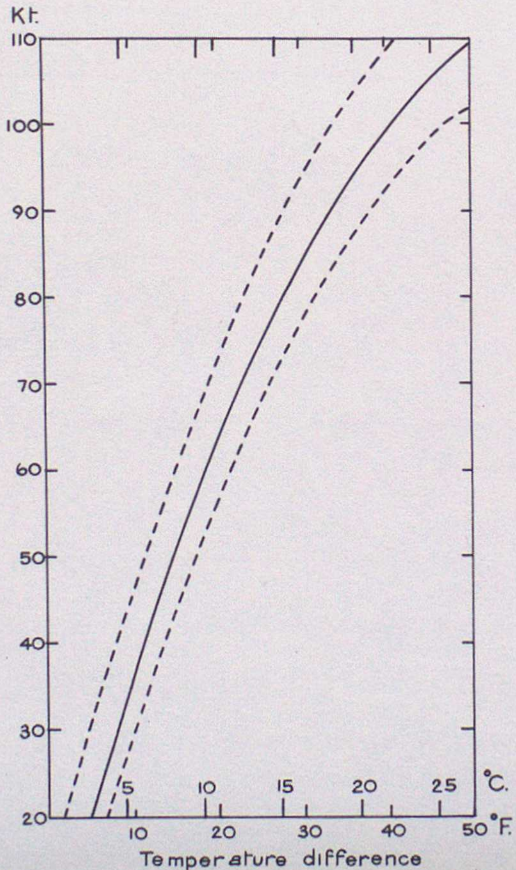


FIGURE 21 *Peak wind speed and temperature differences in thunderstorms in the United States of America*

Full line = regression curve
Broken lines = standard error of estimate

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[39] 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 21.

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°C. isotherm and the wet-bulb curve of a representative tephigram. This is illustrated in Figure 22. (Note.- The wet-bulb temperature is readily obtained on a tephigram from dry-bulb and dew-point temperatures by the method described in Chapter 3.) Thus a forecast of dry-bulb temperature and an estimate of the "downrush" temperature together with Figure 21 yield an estimate of the peak wind speed (with a standard error of estimate of 8 knots) likely to occur in non-frontal thunderstorms in the United States. There is no record of this method having been systematically tested for the United Kingdom.

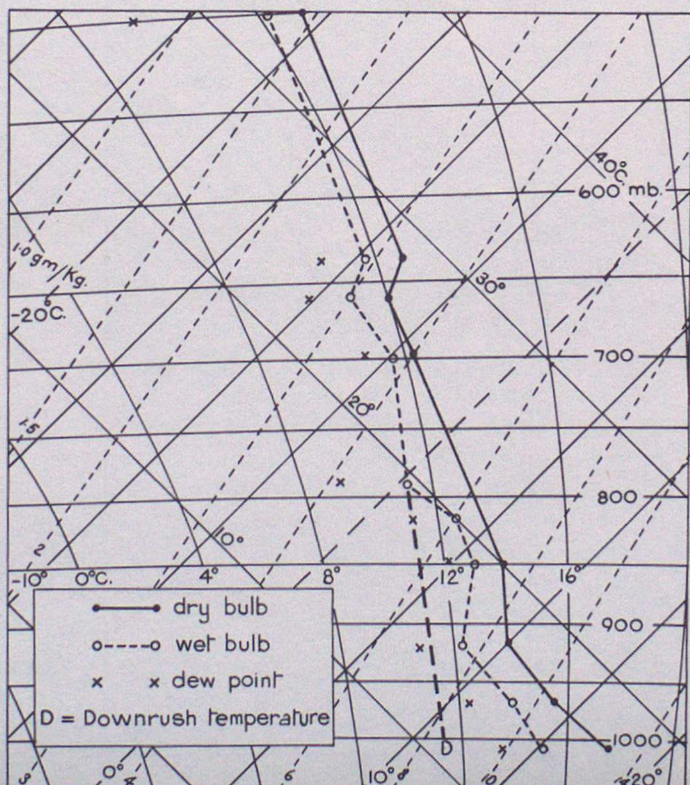


FIGURE 22 *Example of tephigram to illustrate the method recommended by Fawbush and Miller[39] for computing the "downrush" temperature in non-frontal thunderstorms in the United States of America*

13.9.7. *Sea-breezes*

There are few general rules for forecasting sea-breezes but clearly the difference in temperatures over the land and sea and the strength and direction of the general gradient wind must be important parameters. Local topography is undoubtedly important also and cannot satisfactorily be dealt with generally. Each site or area must be considered by the forecaster. However, a "local" investigation can often be of value in forecasting the sea-breeze and an account of such an investigation has been given by Watts.[40] Watts investigated the occurrence or non-occurrence of sea-breezes at Thorney Island (between Portsmouth and Chichester) for offshore winds of various wind speeds and for five groups of wind direction at 3,000 feet in combination with the temperature excess (in a screen) over the land over that at a depth of about 2 feet in the nearby Chichester harbour. Figure 23 shows the results.

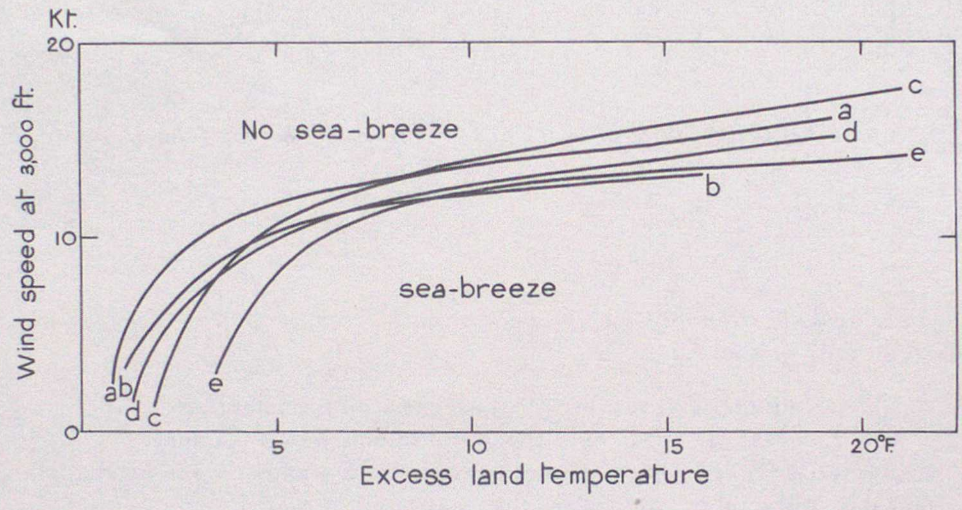


FIGURE 23 *Occurrence of sea-breeze at Thorney Island*
Direction of 3,000 ft. wind:
(a) 280 - 319° *(d)* 020 - 059°
(b) 320 - 349° *(e)* 060 - 089°
(c) 350 - 019°

The curves shown in Figure 23 represented a fairly good separation of occurrences from non-occurrences.

Watts also examined the relationships between the times of onset of sea-breeze and the excess land temperature, the speed of the offshore wind and its direction at 3,000 feet. Rather surprisingly, so far as could be judged from the data, the magnitude of the excess land temperature had no bearing on the time of onset of the sea-breeze. The relation between times of onset and wind speed and direction at 3,000 feet for Thorney Island is shown in Figure 24.

Watts found that the manner of onset of sea-breeze at Thorney Island bore a relation to the thermal stability of the air. The arrival of the sea-breeze is gradual in conditions of thermal stability but is sudden in unstable conditions. Out of 58 occasions of sea-breeze at Thorney Island in 1952, 26 occurred when convection was possible to at least 10,000 feet and the average time taken to establish the sea-breeze was 21 minutes; in 16 of these cases the time taken was 10 minutes or less. On the remaining 32 occasions when convection was not possible the average time was 117 minutes.

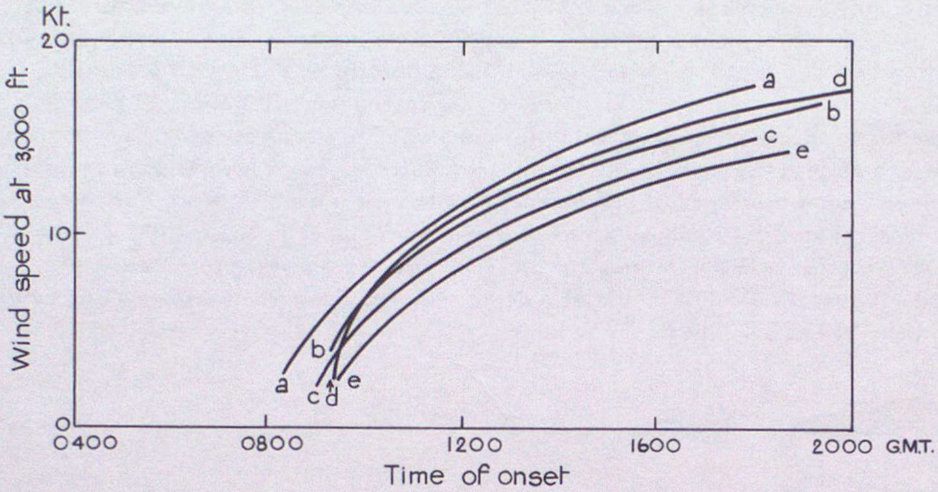


FIGURE 24 *Time of onset of sea-breeze at Thorney Island*
Direction of 3,000 ft. wind:
(a) 280 - 319° *(d)* 020 - 059°
(b) 320 - 349° *(e)* 060 - 089°
(c) 350 - 019°

Watts describes the following procedure for forecasting the sea-breeze with offshore winds at Thorney Island.

(i) Obtain the maximum expected excess land temperature.

(ii) Forecast the upper wind speed and direction at 3,000 feet.

(iii) From (i) and (ii) use Figure 23 to estimate whether a sea-breeze is probable, marginal or improbable.

(iv) If the sea-breeze is probable or marginal use (ii) above and Figure 24 to determine the time of onset.

(v) Estimate the stability of the air. If stable, then expect an average period of transition from gradient to sea-breeze direction from about two hours *before* the forecast time. If unstable, expect a sharp change from gradient to sea-breeze direction at about the forecast time.

It must be emphasized that these results were determined from actual occurrences at Thorney Island and they are not necessarily applicable for other sites. They have been included to illustrate the value of this type of investigation and to stimulate local initiative to undertake investigations for other sites.

Experience shows that the strength of the sea-breeze usually rises to a maximum in the early afternoon and tends to decay gradually in the late afternoon or early evening by which time it has often decayed almost entirely. The horizontal extent of penetration inland is quite variable but 10 to 25 miles would probably be typical of many days. On some occasions and with favourable wind directions or aided by topography the extent of penetration inland may be much greater. Marshall[41] has described a case in which he was able to detect the sea-breeze in the Thames Estuary at about 1130 G.M.T. and to obtain isochrones of the onset of sea-breeze as it moved westwards through the Thames Valley. The sea-breeze achieved its maximum penetration inland at 2040 G.M.T. when it reached Abingdon (approximately 90 miles from Southend). On this occasion the geostrophic wind was in the same direction as the sea-breeze and undoubtedly assisted in the deep penetration inland.

13.9.8. *Land-breezes*

The synoptic conditions favourable for the development of land-breezes are clear skies at night and slack pressure gradients. In the United Kingdom, where the ground is orographically featureless for several miles inland from the coast, land-breezes are normally very feeble. However in localities where a line of hills is located more or less parallel to the coastline a down-slope wind at night may reinforce a weak land-breeze so that a noticeable wind from land to sea develops. For example, the land-breezes of the Cheshire Plain and of the lower-lying parts of Lancashire are reinforced by winds which originate as down-slope winds on the western slopes of the Pennines. The more gentle slopes of the South Downs also produce down-slope winds under suitable conditions so that the land-breezes on the coasts of Sussex are reinforced. When both a land-breeze and a down-slope wind are occurring it is difficult to determine which part of the surface wind is due to the one or other cause. In many cases and from the point of view of the practical forecaster it is probably better not to try to separate these two components but to regard them as jointly producing the "nocturnal wind". Many of the values, due to Lawrence,[31] which were given in Section 13.6.2 refer to "nocturnal winds". A local study of the nocturnal wind at Thorney Island has been made by Moffitt.[42] His results are of direct practical value for forecasting winds in that locality. Thorney Island is situated near a land-locked harbour on the south coast of England. The southern slopes of the South Downs in this area lie roughly parallel to the coastline with their ridge rising to some 600 feet above mean sea level at a point about 8 miles north-north-east of the station. At Thorney Island the nocturnal wind is from the north-north-east and may occur at any time of the year when clear skies and light winds prevail. If there is a light onshore gradient the nocturnal wind sets in with a sudden change of wind direction to about 030° (and a sudden temperature fall of about 2°F.).

Moffitt considered that the nocturnal wind at Thorney Island is primarily of katabatic origin but that there was an additional land-breeze effect when the sea was warmer than the land. All occasions during 1952-54 with gradient wind speed less than 30 knots with an onshore component were examined. The occurrences, non-occurrences or intermittent occurrences of nocturnal winds were then plotted on a scatter diagram using the difference between day maximum and subsequent night minimum temperatures as abscissae and wind speed as ordinates. Moffitt was able to draw a curve separating most of the occurrences from the non-occurrences. The curve is reproduced in Figure 25.

The occasions of nocturnal wind which were above the line were mainly of two types:

(a) occasions with a large temperature difference between sea and land (sometimes as much as 20°F.),
(b) occasions of fairly strong gradient wind.

Moffitt observed that occasions *(b)* were mostly partial occurrences in which the nocturnal wind set in for a few minutes several times during the night but was continually dispersed by turbulence created by the relatively strong gradient wind.

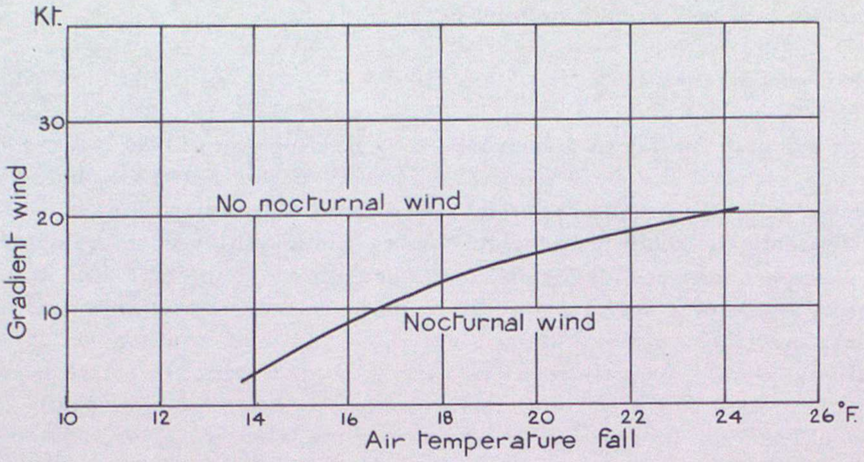


FIGURE 25 *Graph of gradient wind and temperature fall at Thorney Island*

Figure 25 clearly had forecasting value but it was also desirable to be able to forecast the time of occurrence of a nocturnal wind. It was found that 90 per cent of all occurrences arrived when the temperature at Thorney Island had fallen by 14 to 16°F. from the day maximum. During 1952-54 on clear nights, 68 per cent of occurrences arrived between 2 and 4 hours after sunset but with variable cloud cover some 90 per cent of occurrences began more than 4 hours after sunset.

From these investigations Moffitt deduced the following rules for forecasting the nocturnal wind at Thorney Island:

(i) Forecast the night minimum temperature (see Chapter 14, Section 14.7.2) and hence deduce the expected fall of temperature from the day maximum.

(ii) Forecast the onshore gradient wind speed.

(iii) Enter Figure 25 and estimate whether a nocturnal wind is probable, improbable or marginal.

(iv) Consider whether the nocturnal wind may arrive earlier than 2 hours or later than 4 hours after sunset, bearing in mind the expected cloudiness and the coastal sea temperature.

Land-breezes in the United Kingdom are normally very shallow and probably do not exceed 200 feet in depth.

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 a maximum, namely, in anticyclonic conditions in early autumn.

13.9.9. *Katabatic and anabatic winds*

In the description of local winds in Section 13.6 the terms katabatic and anabatic were avoided for the following reason. The definitions of katabatic and anabatic winds specifically include the down-slope and up-slope winds but they do not appear to exclude the mountain and valley winds. It appeared desirable therefore, in the descriptive account, to treat each wind separately. For the rest of this section the term katabatic or anabatic wind should be taken as implying both slope and mountain or valley winds.

The synoptic conditions favourable for the development of katabatic winds are clear nights, either a very slack or no pressure gradient, and low humidities - conditions in which there will be strong outgoing long-wave radiation (see Chapter 14). This leads to strong cooling which causes katabatic winds to flow down slopes and also leads to or increases a surface inversion of temperature. Further cooling does not necessarily lead to an increase in katabatic flow. 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 100 feet. The depth of mountain winds may be rather greater and, on occasions, may extend to 500 feet.

Katabatic (and nocturnal) winds are usually less gusty than day-time winds. However, katabatic winds sometimes set in with short bursts of a periodic nature. Heywood[43] illustrates such a case when the katabatic wind took some 2 hours to become established and, during this period, the wind showed periodicities of the order of 20 minutes. Heywood also found that the katabatic wind was not specially frequent at any particular season of the year. It would seem that a clear night is just as likely to lead to a katabatic flow in summer as in winter.

Eldridge[45] has given a useful account of katabatic wind at Driffield. From his account it was not possible to produce a precise forecasting rule but with a geostrophic gradient of 10 miles per hour or less (whatever the direction), less than a quarter of the sky covered for three or four hours after sunset and a high day maximum temperature a katabatic wind at Driffield is almost certain. There is a good chance of it with gradients up to 20 miles per hour. According to Eldridge the katabatic wind at Driffield may be expected to start about 2 to 3 hours after sunset and last till about dawn.

A search of the literature has failed to reveal any work on anabatic winds which is of direct practical value for forecasting these winds in the United Kingdom.

It is readily apparent that there is no systematic simple procedure by which a forecaster can make estimates of anabatic or katabatic winds. The data included in Section 13.6 should form a useful background. Topography is vital and the only sound advice is to consider each site individually, and draw conclusions based on physical reasoning. Subsequent forecasts can then be modified in the light of experience. The value of investigations of past records of surface winds can hardly be overstressed. Actual winds are, of course, the resultant of all forces operative. They include the effects of pressure gradient, slope and mountain and valley winds and any "funnelling" or sheltering effects of topography. A systematic analysis of past records will often produce results which can be directly applied to forecasting; at the least they provide sound climatological information of a localized nature which is of great value to forecasters fresh to the district.

13.10. FORECASTING UPPER WINDS

For forecasts up to about 12 hours ahead it is usually preferable to start with the recent observations and modify them in the light of the trends indicated by the prontour. It was shown in Section 13.8 that persistence is often a good forecast of the wind for a few hours ahead. In spite of this forecasters should always consider carefully the changes which have taken and are expected to take place. This will involve an examination of the wind observations and the changes which have occurred in time and space during the preceding 12 hours or so over and around the forecast area. It will also involve a consideration of the general synoptic situation and the rather longer-term developments which are expected to occur. From such examinations it is usually possible to infer in a rather *ad hoc* manner the changes in the wind régime which are likely to occur in the following few hours. Wind hodographs are often useful for this type of wind forecasting. Hodographs show in a very clear way the wind shear in the vertical and the levels in which the advection of warm or cold air is taking place. This knowledge is often of great value for accurate short-period forecasting. In many cases hodographs enable systematic and mutually consistent short-period upper wind forecasts to be made by extrapolation from the latest observed values in a way which is scarcely possible by a more casual modification to the latest observation from the area concerned or from some suitable upwind site.

The forecasting of upper winds from the top of the friction layer (about 2,000 feet or 60 metres) to about 100 millibars for periods of about 12 to 36 hours ahead rests primarily on the measurement of geostrophic or gradient values from suitable forecast charts. Where these forecast charts are valid for the time of the forecast a direct measurement is adequate for a first approximation. In many cases, however, the forecast period will differ somewhat from that for which the forecast charts were prepared. It is then necessary to interpolate or extrapolate. Generally, straightforward linear interpolation or extrapolation will be satisfactory but for periods beyond 24 hours where intense development is expected, allowance for some non-linear effects may seem desirable. In such cases the accuracy of the forecast charts may be doubtful since rapid development and/or movement are difficult to forecast accurately with confidence. Errors in the charts may nullify the extra accuracy which it was hoped to achieve by non-linear extrapolation. In practice it is probably difficult to improve consistently on linear interpolation or extrapolation.

On some occasions the deformation of the thickness pattern is important and the speed and extent of this deformation is often difficult to estimate accurately. Although this is primarily a problem for the preparation of prebaratic and prontour charts it cannot wholly be ignored when interpreting prontour charts in terms of wind and estimating in broad terms the extent to which the actual wind may deviate from the forecast given.

When interpreting upper air charts in terms of wind it is important to take account of the probable accuracy or errors involved. This subject has been considered by several workers.

Bannon[46] has estimated the deviation of the wind from the contours at Liverpool. He concluded that at 700 and 500 millibars the deviation of the wind from the contours was small, the mean deviation regardless of sign being of the order of 9°. On the edge of anticyclones the deviation appeared to be significantly one of a wind blowing in towards high pressure at those levels. There was also an apparent systematic deviation across the contours associated with travelling troughs and wedges at those upper levels, being towards lower pressure ahead of troughs and towards higher pressure behind troughs. At 700 millibars the mean absolute value of the vector deviation of the wind from the geostrophic was about 9 knots. At 500 millibars the value was about 12 knots, but the determination at the upper level was much less accurate.

In the evaluation of winds from actual contour charts there is some uncertainty whether the geostrophic or gradient value provides a better estimate of the actual wind. It is not possible to be specific on this point. Zobel[47] has given details of a comparison of geostrophic and gradient winds as deduced from 200-millibar contour charts with actual winds observed at Crawley over a period of 1 year. The comparisons indicated that:

(i) There is a general overall increase of accuracy of wind estimation from contour charts at 200 millibars when the geostrophic wind is corrected to gradient wind by means of a correction for curvature of the contours.

(ii) The increase of accuracy applies to both cyclonic and anticyclonic types and to all degrees of curvature.

(iii) Accuracy is most markedly improved if the curvature is anticyclonic and the wind speed in excess of 60 knots.

(iv) The improvement of the gradient approximation over the geostrophic is appreciable for wind direction of either curvature between south-east and north-east through west. For other directions the accuracy of both is approximately the same.

In view of this the gradient wind evaluated from a 200-millibar contour chart is likely to be a closer estimate of the actual wind than the geostrophic value. Similar comparisons are not available for other levels.

Murray[48] has made a broader survey of the relation of winds to contour charts. Actual contour charts are not perfect representations of the wind field. The errors depend on the errors of radio-sonde and radio-wind observations, on the personal element of chart construction, on the small-scale wind fluctuations and on errors introduced by assuming and/or estimating geostrophic winds. Murray has made estimates of all these for pressure levels 700, 500, 300 and 200 millibars. He has also estimated the errors introduced by the erroneous forecasting of contours. A summary of his results is given in Table 22.

Among the main conclusions Murray drew from his study were:

(i) A considerable contribution to the errors of forecasting upper winds by the contour technique is made by errors inherent in the technique, which cannot, by any obvious means, be avoided.

(ii) Errors owing to geostrophic departures account for only a rather small fraction (of the order of 10 per cent) of the total root mean square error of forecasts. The geostrophic approximation must therefore be regarded as reasonably satisfactory in practical work on upper air forecasting by contour methods.

(iii) Errors caused by erroneous forecasting of the contour patterns are definitely the greatest up to 300 millibars at least, accounting for some 70 per cent of the total forecast error at 700 and 500 millibars and about 60 per cent at 300 millibars. At 200 millibars this error is about equal to the combined effect of errors of observation and technique.

(iv) The total forecast errors are substantially less than the 24-hour wind variations except at 200 millibars where the difference is small. However the errors due to erroneous forecasting of the contour patterns at all levels are much less than the 24-hour wind variations.

TABLE 22 *Magnitude of errors in 24-hour forecasts of wind (at a point) from forecast charts analysed according to the contribution of various factors and compared with the variability of the wind at selected levels*

	Pressure level (millibars)			
	700	*500*	*300*	*200*
		knots		
(1) Error owing to errors of observations of contour height	3	4	10	14
(2) Error produced by inaccurate estimation of geostrophic wind	5	7	8	6
(3) Error arising from personal element in chart construction	4	5	6	4
(4) Error in measurement of wind*	2	3	4	4
(5) Error caused by small-scale fluctuations	4	4	4	4
(6) Overall inherent technique error* (errors (1) to (5) combined)	8	11	15	16
(7) Apparent departure from geostrophic value on working charts	10	14	20	18
(8) Geostrophic departure (error (7) less (6))	6	9	13	8
(9) 24-hour forecast error	20	27	33	25
(10) Error caused by erroneous forecasting of contours (error (9) less (7))	17	23	26	17
(11) 24-hour wind variation	25	35	48	27
(12) Standard vector deviation from seasonal mean wind	28	36	50	40

All figures are root mean square vector errors or departures or variations.

* Errors in wind measurement affect any comparison between forecast and observed winds and are therefore included here. They should not strictly be regarded as part of the forecast error, but their omission does not materially affect the figures in row (6).

Ellsaesser[38] has examined a number of studies on the accuracy of forecasts of upper winds. Some adjustments to the various results had to be applied to enable comparisons to be made. From the heterogeneous data of some four separate studies Ellsaesser obtained Table 23 which shows the average vector error in 24-hour forecasts of wind.

TABLE 23 *Average vector error in 24-hour forecasts*

	Pressure level (millibars)				
	700	*500*	*300*	*200*	*100*
			knots		
(1) Average vector error in 24-hour forecasts	16	20	28	29	18
			per cent		
(2) Percentage of mean wind speed at these levels represented by (1)	67	58	59	59	70

Ellsaesser found that the forecast error varied with the season, being less in summer than in winter.

By a judicious combination of available prontours prepared by and disseminated from certain major forecasting centres and the latest wind soundings, wind forecasts at any level up to about 100 millibars can be made for areas around north-west Europe. Forecasts of winds up to 100 millibars are, at the time of writing, sufficient to meet the normal needs of manned aircraft. If winds at slightly lower pressures (higher levels) are occasionally required they can generally be adequately estimated by extrapolating upward from 100 millibars using available observations of winds. It is fortunate that, in temperate regions, wind patterns at these levels usually change shape only slowly and the patterns themselves move quite slowly. Further, wind speeds are often much less than at lower levels near the tropopause because the thermal gradients in the lower stratosphere are often in reverse to those in the upper troposphere. Thus a rather *ad hoc* procedure probably suffices to meet occasional enquiries for winds above 100 millibars although it is probable that regular demands for such information would require a more systematic approach.

Near the tropopause the winds are often strongest and in situations where the height of the tropopause varies or is likely to vary substantially either in space or time, maps of the level(s) of the tropopause may be helpful in improving the accuracy of wind forecasts near these levels. The more general discussions of the jet stream and of winds at the levels near the tropopause which were given in Chapter 7 should provide some useful background information on which to base a judgement.

The current techniques of analysis and prognosis by systematic graphical additions of thicknesses to a lower-level chart to deduce a higher-level chart are well suited to lower and mid tropospheric levels. At high tropospheric levels and low stratospheric levels they have some drawbacks since the stratum between two pressure surfaces may include the tropopause in one area of the chart but not in another. Because of the change in thermal gradients near the tropopause the thickness isopleths for the layer may smooth out thermal differences which are much more marked near the tropopause. Near the British Isles the tropopause will usually be between about the 300- and 100-millibar pressure levels (but it may occasionally be below the 300-millibar level) so that the problem is not unimportant particularly as these are the pressure levels between which the majority of jet aircraft operate.

No systematic form of analysis and prognosis has yet been devised which is effective, practical and theoretically satisfying so far as the thermal discontinuity at the tropopause is concerned. A method of displaying the variation of winds near the tropopause when it lies between 300 and 200 millibars has been described by Harmantas and Simplicio.[49] The method requires the use of 300- and 200-millibar charts and the 500-300-millibar thickness chart. In addition the contours of the tropopause and estimates of the vertical wind shear between the tropopause and the constant pressure surfaces above and below are required. In obtaining these estimates Harmantas and Simplicio make the assumptions that:

(i) the core of the jet stream lies somewhat below the tropopause and vertically above the zone of maximum wind at 300 millibars,

(ii) the horizontal temperature gradient at 300 millibars is conservative for all levels from 300 millibars upward to the tropopause,

(iii) the horizontal temperature gradient in the stratosphere is conservative for all levels below 200 millibars to the base of the stratosphere,

(iv) a tropopause break can be replaced by a continuous tropopause with uniform slope.

The contours of the tropopause are determined and delineated on a chart. The examples given by the authors indicate that the height is shown by isobars, that is, where the tropopause is at 300 millibars, 250 millibars and 200 millibars. As an aid to determining the geographical distribution between the locations for which direct upper air observations are available Harmantas and Simplicio suggest that:

(v) the position of the zone of maximum winds at the 300-millibar level appears to be fairly close in the horizontal to the position in the vertical of the tropopause break,

(vi) time continuity may be used.

Isotherms at 5°C. intervals are next drawn on the 300-millibar map for those areas where the tropopause lies above that level (that is, at lower pressure). Isotherms are determined from actual observations and by using the 500-300-millibar thickness field as a guide. This procedure gives the tropospheric isobaric temperature field at 300 millibars. Using assumption (ii) above these isotherms may be converted into a vertical wind shear. A geostrophic scale for 200-foot intervals, when applied to isotherms at 5°C. intervals measures the vertical wind shear in knots per 10,000 feet. These values can be broken down into knots per 1,000 or 2,000 feet, whichever is more convenient. Alternatively isotherms at smaller temperature intervals may be constructed. From the values of wind shear so obtained isopleths of the vertical wind shear from 300 millibars up to the tropopause may be drawn on the contour chart of the tropopause; the directions of these vertical shears are given by the direction of the isotherms.

In a similar way the 200-millibar stratospheric temperature field is determined using direct observational data together with a knowledge of the position where the tropopause is at 200 millibars. The vertical wind shears in convenient units are computed from the isotherms. Isopleths of the vertical wind shear from 200 millibars down to the tropopause are drawn on a duplicate copy of the contour chart of the tropopause and the directions of these shears are obtained from the direction of the isotherms.

Then, to determine the wind for a flight on intermediate level between 300 and 200 millibars, the following procedure is used:

(i) Determine, from the map of the tropopause contour, those parts of the flight which are in the troposphere and in the stratosphere.

(ii) For those parts of flight in the troposphere, use the 300-millibar wind to which is added vectorially the appropriate value of the shear vector for the distance which the flight is above the 300-millibar surface.

(iii) For those parts of flight in the stratosphere use the 200-millibar wind to which is added vectorially the appropriate value of the shear vector for the distance which the flight is below the 200-millibar surface.

The authors make no mention of the techniques needed for forecasting the various isopleths required. Even the analytical processes will require considerable labour and it is clear that this method could only be attempted in a fairly large office with adequate upper air data and the staff to deal with it. The writer of this handbook has had no practical experience of this technique. Nevertheless the method has been included because it does represent an approach to the determination of tropospheric winds from tropospheric values of temperatures and winds and of stratospheric winds by extrapolating downwards from stratospheric values. As mentioned earlier, thickness techniques in layers which contain the tropopause are not very reliable and a technique which attempted to get over this difficulty of the discontinuity in temperature fields at the tropopause, even though a number of assumptions were made, seemed worthy of inclusion. However, since considerable labour in applying the technique is involved and the problem of forecasting the isopleths may be severe, it is possible that estimates of the wind variations from measurements of the winds at the standard pressure levels, modified qualitatively to take into account variations of the tropopause levels and backed up by careful inspection of latest rawinds and/or aircraft reports, may, with considerably less labour, yield results which are as good.

13.11. SOME OTHER METHODS OF FORECASTING WINDS

Although conventional United Kingdom techniques rest on measurements of isopleths on isobaric or contour forecast charts it should not be thought that no other technique is available. For example, charts of stream-lines and isotachs may be used and statistical techniques are available for forecasting either for longer periods ahead or with very limited data. Statistical forecasts appear to have a smaller error than those arrived at by orthodox techniques when very limited data are inadequate for satisfactory contour charts to be drawn.

13.11.1. *Use of isotachs*

An isotach is a line along which the wind speed is constant. Thus a family of isotachs will completely determine the speed of the winds in a horizontal or isobaric plane. The wind direction can be determined by a family of isogons which are lines along which the direction is constant. Thus sets of isotachs and isogons will completely determine the horizontal or isobaric wind field. Isogons are, however, seldom used in practical forecasting. A set of stream-lines would also provide the second quantity required to determine the wind field but stream-lines are seldom drawn in temperate latitudes. However, a set of contours provides an approximation to the stream-lines but it is important to remember that a contour is only an approximation to an instantaneous stream-line and that it does not yield a close estimate of the trajectory of a parcel of air over an interval of several hours.

By using isotachs and contours a practical approximation to the wind field is obtained. The technique is more useful at the higher levels (for example, 300-millibar to 200-millibar levels) in association with strong winds. Some papers which contain remarks on the construction and use of isotachs are listed in references 50 to 54.

Experience shows that, except in regions of light winds, the patterns of isotachs present a coherent picture which persists and can be followed with reasonable continuity on a sequent of synoptic charts. Isotach patterns usually move in orderly fashion in conformity with the general synoptic developments. Accordingly, once the general development has been estimated, isotach patterns can often be extrapolated and modified in a qualitative manner. In the cores of jet streams the areas where winds are strongest are often clearly shown by a system of closed isotachs (usually elongated along the jet stream). When using isotachs it is a useful practice to indicate the axis (which is usually curved) of the jet stream by means of an arrow aligned along the jet maximum. When drawing such axes it should be noted that the axes of maximum wind are not necessarily parallel to contours. This is particularly often the case in the region a few hundred miles upwind and downwind of the wind maximum. Such maxima move much more slowly than the wind speeds in the maxima so that the air moves through the maxima. Thus air is accelerated when moving towards a maximum and decelerated when moving downstream from the maximum. These accelerations produce a flow across the contours. The cross-contour flow is from high to low when the air accelerates (that is, increases downstream) and from low to high when the air decelerates (that is, decreases downstream). At the centre of the wind maximum the acceleration along the flow is zero and the jet-stream axis will be very close to the direction of the contours. Figure 26 shows schematically this inclination of a jet-stream axis to the contours.

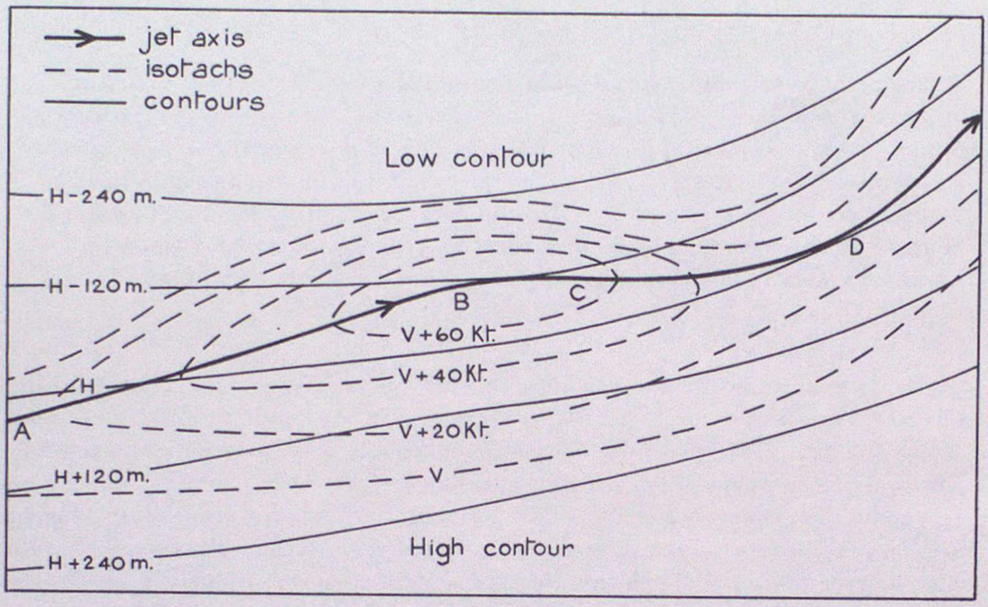


FIGURE 26 *Schematic illustration of inclination of jet axis to contours*

It will be seen that between *A* and *B* speeds along the jet axis increase by some 40 knots leading to substantial acceleration and that the jet axis is inclined towards lower contour heights. Conversely from *C* to *D* where there is deceleration the jet axis crosses the contours towards greater heights. Between *B* and *C*, a region of almost uniform speed, the jet axis osculates the contour.

Simla[53] has made some approximate calculations of the angle between the jet axis and contours for a number of latitudes for changes of wind speeds of 40 knots in various distances along the jet axis. The calculations involve a number of assumptions and approximations and the detailed results are not reproduced here. In latitudes between 45 and 60°N. the calculations show that the inclination of the jet axis to the contours is about 15 to 20° when the speed changes by 40 knots in about 300 miles or about 10° where a 40-knot change occurs in 600 miles - measured along the jet axis. For lower latitudes and more rapid changes of wind speed the calculations indicate that the inclination may be much greater.

When drawing and extrapolating isotachs it should be noted that the horizontal shear of wind is often much greater on the cold side of a jet than on the warm side. The corresponding isotachs are therefore often very closely packed on the cold side of a jet maximum. The more general description of wind profiles in jet streams which were included in Chapter 7 may be helpful when using isotachs.

13.11.2. *Statistical methods*

Durst[37] has investigated the variation of wind with time and distance and, from a consideration of the statistical results, has devised methods by which forecasts of probable winds at a subsequent time, either at a place or along a route, can be made by using the latest available measurements of upper winds together with some predetermined correlation coefficients.

Wind

For periods of 24 to 36 hours ahead and in areas where data are fairly plentiful it seems likely that the orthodox methods of forecasting upper winds will yield results which are, on average, better than simple statistical techniques. However, for longer periods or in areas where data are sparse, statistical techniques may well provide the best possible forecast with current methods.

The crux of the method lies in the predetermination of correlation coefficients from a statistical examination of observed data for a given location. The determination of these correlation coefficients is laborious and time-consuming and they must be determined for each location, for various pressure levels (or heights) and for suitable seasons of the year. Statistical forecasts of spot winds can be improved by taking account of the wind observations from another station. This requires the compilation of further correlation coefficients. However, although the labour of compilation is great, once the coefficients are available, their application on the forecasting bench is very simple. This can be seen from the following equation which indicates the type of relation which is used for wind forecasts at a point.

Forecast wind at A (at time t) = $\overline{V}_A + a(V_A - \overline{V}_A) + b(wV_B - w\overline{V}_B)$ where, for given t and stations A and B, a, b and w are constants which can be predetermined, $\overline{V}_A$ is the climatological average wind at A, $w\overline{V}_B$ the weighted climatological average wind at B (these constants can also be predetermined) and V_A and V_B are the observed wind at A and B respectively.

Thus on the right-hand side of the equation the only variables for given t and stations A and B are the observed wind at A and B.

The equation may be expressed in words as follows:

Forecast wind at station A = Climatological average wind at station A
+ a × (the actual observed wind minus climatological average wind)
+ b × (the difference between a weighted value of the actual observed wind at the station B minus the weighted climatological average wind at station B).

The method may be extended to yield a forecast of wind over a route, either as component winds along and across the route or in the form of a mean component along the route. The method permits an estimate to be made from only one observation at one end of the route but for long routes the use of several observations along the route is preferable (and improves the accuracy). For example, the probable future wind along the route London to Rome may be deduced from a single wind sounding in southern England and the probable future winds on the route Ireland to Newfoundland may be calculated from wind soundings in Ireland, over ocean weather stations J (53°N.20°W.) and C (52°30′N.35°W.) and in Newfoundland.

These forecasting methods are not generally in use in the United Kingdom and data for their application are not included in this Handbook. For details of the theory the reader should refer to the paper by Durst.[37] Some information on the technique of preparation of statistical wind forecasts and on the accuracy of such wind forecasts, mainly in comparison with the accuracy of wind forecasts made by synoptic techniques, are contained in papers by Johnson[55] and by Durst and Johnson.[56]

At some stations information is sometimes requested for probable winds at various seasons of the year along various routes for specified altitudes. It is appropriate in this section to refer to the technique of calculating such information which must be obtained by statistical techniques. Full details of the method and its practical application are contained in references 57 and 58.

13.12. USE OF AIRCRAFT REPORTS OF WIND

Reliable weather reports from aircraft are often of great value to forecasters in many aspects of their work, and certainly not least in importance are those related to winds. It is, however, important that these reports should be applied with care and discretion. Reports may be received from aircraft whilst in flight or after landing. In the latter case the reports are received by direct debriefing of a member of the aircrew through an intermediary or from completed forms delivered to the Meteorological Office.

So far as in-flight reports are concerned their value is enhanced by the up-to-date nature of the information. If transmission errors are suspected it may be possible to have the message checked but, if not, the forecaster must make a judgement based on synoptic experience and all other available information. If the reported wind seems definitely out of line the forecaster may have to reject it but, if further reports are expected either from the same aircraft or from other aircraft in the locality, it may sound practice to defer a final judgement until further reports become available. If "*D*" factors (that is, the difference between readings on a radio altimeter and a pressure altimeter set to standard pressure level - 1013·2 millibars) are reported the difference in "*D*" factors reported by one particular aircraft at successive positions are equivalent to contour differences (within the limits of accuracy of observation) and so yield an indication of the component of wind speed perpendicular to the track of the aircraft. A succession of "*D*" factors may assist in determining the spacing of the contours in the pressure surface in which the aircraft is flying. The errors of aircraft altimeters are such that "*D*" factors reported by different aircraft sometimes show large differences and considerable care must be exercised in using differences in "*D*" factors. The difference between successive "*D*" values from the same aircraft may however be expected to more accurate than individual "*D*" values.

Leaving aside observational or transmission difficulties it is important that the locality to which the wind applies should be known. The codes in use in 1960 make provision for the type and reliability of the wind reports to be specified. In reports from meteorological reconnaissance aircraft this information can be specified with considerable detail but in reports from transport aircraft the "mean" wind is reported and it is not precisely clear what the "mean" is. Further the method by which the mean wind has been determined is not specified but some indirect assessment of the reliability of the wind may be obtained by considering the method used for determining the position of the aircraft. Provision for reporting this is made in the code and consideration of this data (where given) will indicate, in a general way, the confidence to be placed in the accuracy of the reported position of the aircraft and therefore, indirectly, of the accuracy of the mean wind. Some transport aircraft, whilst in flight, transmit some messages which contain, primarily, information of operational significance but the messages may also contain some weather information in "semi-plain" language. This type of message is usually compiled in a rigid order and, in the meteorological section, there may be included a value for the mean wind. In some cases this mean wind is immediately followed by either a position or a time which specifies the location, in space or time, about which the mean wind was taken. In some plain-language reports the wind is given as the equivalent tailwind. Reports of tailwinds (or headwinds) are of less value to the forecaster than reports of wind direction and speed.

From the foregoing paragraph it is quite clear that winds reported from aircraft are not "spot" winds. They are almost always mean winds and the accuracy of the mean will vary. In these respects they must be interpreted with circumspection. When winds are strong, reasonably steady in direction and there is considerable horizontal shear, it may generally be assumed that the horizontal shear in the direction of the current is much smaller than that at right-angles to it. In such circumstances mean winds reported by aircraft flying across the current may give little indication of the variation of wind across the current. Reported winds from aircraft flying in a direction with or against the current may, on the other hand, be quite representative. Where the upper air patterns indicate quite sharp variations in wind direction and speeds (for example, a sharp upper trough) it may not be possible to derive much specific information from a reported mean wind whilst the aircraft is traversing the sharp feature but preceding and succeeding reports often indicate very clearly the different wind régimes on either side and enable the forecaster to determine with fair precision the location of the axis of the upper air feature. Over areas where upper air data are sparse such information is often of great value not only for the benefit of other aircraft but also for general synoptic work. Provided the time is available there are few occasions when a close study of in-flight reports from aircraft will not be amply repaid.

When aircrew visit the meteorological office for debriefing after a flight, useful information on winds, which is of direct practical value, is sometimes obtained. Where the wind distribution is relatively flat and featureless a debriefing will seldom achieve little more than a confirmation of the existing analysis and knowledge. However, on those occasions when the wind field in the area of the flight has well marked features, much valuable information can often be obtained from a careful debriefing of the crew, preferably with the analysed charts available for inspection and reference. It may be possible to reposition the features rather more accurately or to vary the gradients slightly particularly where neither interpolation between available routine instrumental meteorological observations nor historical continuity enables such features to be located with much accuracy. In such cases the weight of reliance to be placed on aircraft observations depends on the nature of the flight, the accuracy and frequency of the aircraft fixes and also the competence and experience of the crews. Where aircraft observations are at variance with the historical synoptic patterns and routine meteorological reports judgement is needed in deciding the credence to be given to such reports. Such conflicting aircraft reports must not be rejected out of hand. They should be carefully considered to see if they are feasible in the light of the synoptic situation. If the wind observations from the aircraft indicate a possible development which seems to be supported by other visual observations by the crew (for example, cloud development) the other meteorological data should be carefully re-examined by the meteorologist for any indication (possibly hitherto overlooked) which either confirm or refute the possible development. In some cases a modification to the analysis may be justified but, even where the meteorologist decides that the weight of current evidence is against the aircraft report, subsequent reports should be carefully scrutinized, at least for several hours, in case later developments justify a re-appraisal of the situation.

Handbook of Weather Forecasting

13.13. SELECTED SYNOPTIC EXAMPLES

The variation in time and space of winds associated with typical synoptic situations (for example, cold and warm fronts, waves, cold and warm anticyclones, jet streams etc.) has been described in Chapters 5, 6 and 7. Particular examples illustrating these variations have therefore been excluded from this Chapter.

The synoptic examples which have been chosen illustrate the onset of sea-breezes, the diurnal variation of wind, severe squalls and the occurrence of widespread gales.

13.13.1. *Sea-breezes*

The following two situations were chosen by Watts[40] to illustrate his account of the sea-breeze at Thorney Island.

13.13.1.1. *Gradual change of wind at Thorney Island, 11 June 1952* Plate I shows the synoptic situation over southern England at 0900 G.M.T., 11 June 1952. A ridge extended from the Azores anticyclone to southern England and the English Channel. Within this ridge a small anticyclone was centred near St. George's Channel.

From the tephigram for Larkhill shown in Figure 27 maximum temperatures in the upper sixties or low seventies seemed likely and, with a wind at 3,000 feet estimated at 310° 9 knots at Thorney Island, Figure 23 indicated that a sea-breeze was likely. The inversion of temperature just above the 850-millibar level at Larkhill would restrict convection from the ground at or near 850 millibars. Thus, according to the classification suggested by Watts, conditions were thermally stable on 11 June and a sea-breeze could be expected to arrive around 1100 G.M.T. (obtained from Figure 24) with a gradual change of wind.

The anemogram shown in Figure 27 indicates that between 1030 and 1130 G.M.T. there was a gradual change of wind direction from about 290 to 190 to 200° which is the normal sea-breeze direction for Thorney Island. The thermogram shows a corresponding decrease in temperature.

Plate II shows the detailed synoptic chart for southern England at 1500 G.M.T., 11 June 1952. By comparison with Plate I it will be seen that a sea-breeze had set in at many stations along the south coast from North Foreland to Devon. The sea-breeze had penetrated inland to South Farnborough, but elsewhere the penetration inland at 1500 G.M.T. seemed to have been much less.

13.13.1.2. *Sudden change of wind, 23 July 1952.* Plate III shows the synoptic situation over southern England at 1200 G.M.T., 23 July 1952. The weather over the British Isles was dominated by a ridge which extended north-eastward from the Azores anticyclone. A weak cold front lying from Manchester to the north Kent coast was moving slowly west-south-westward.

The tephigrams for Larkhill are shown in Figure 28. Day-time maximum temperatures in the upper seventies were likely and with a wind at 3,000 feet of about 060°13 knots a sea-breeze was just possible in mid-afternoon. With day-time temperatures in the upper seventies the 0300 G.M.T. Larkhill ascent would show instability to heights above 700 millibars indicating that a sudden onset of sea-breeze might be expected. However, the 1500 G.M.T. Larkhill ascent showed that substantial warming had occurred in the middle troposphere down to about 800 millibars and the accompanying increase in stability would limit convection to heights below 700 millibars. This might have led to a gradual onset of sea-breeze but, in

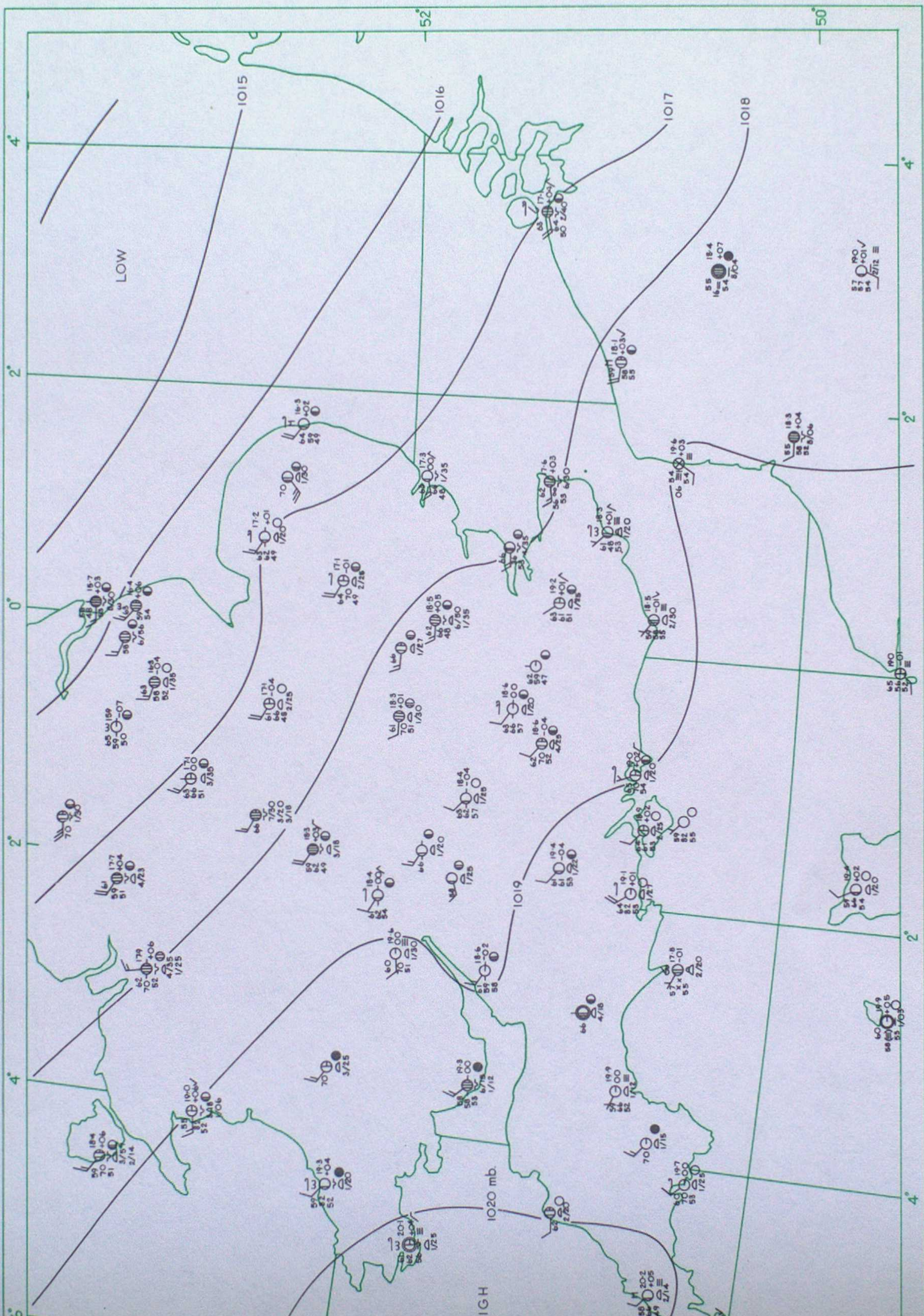


PLATE I Synoptic situation, 0900 G.M.T., 11 June 1952

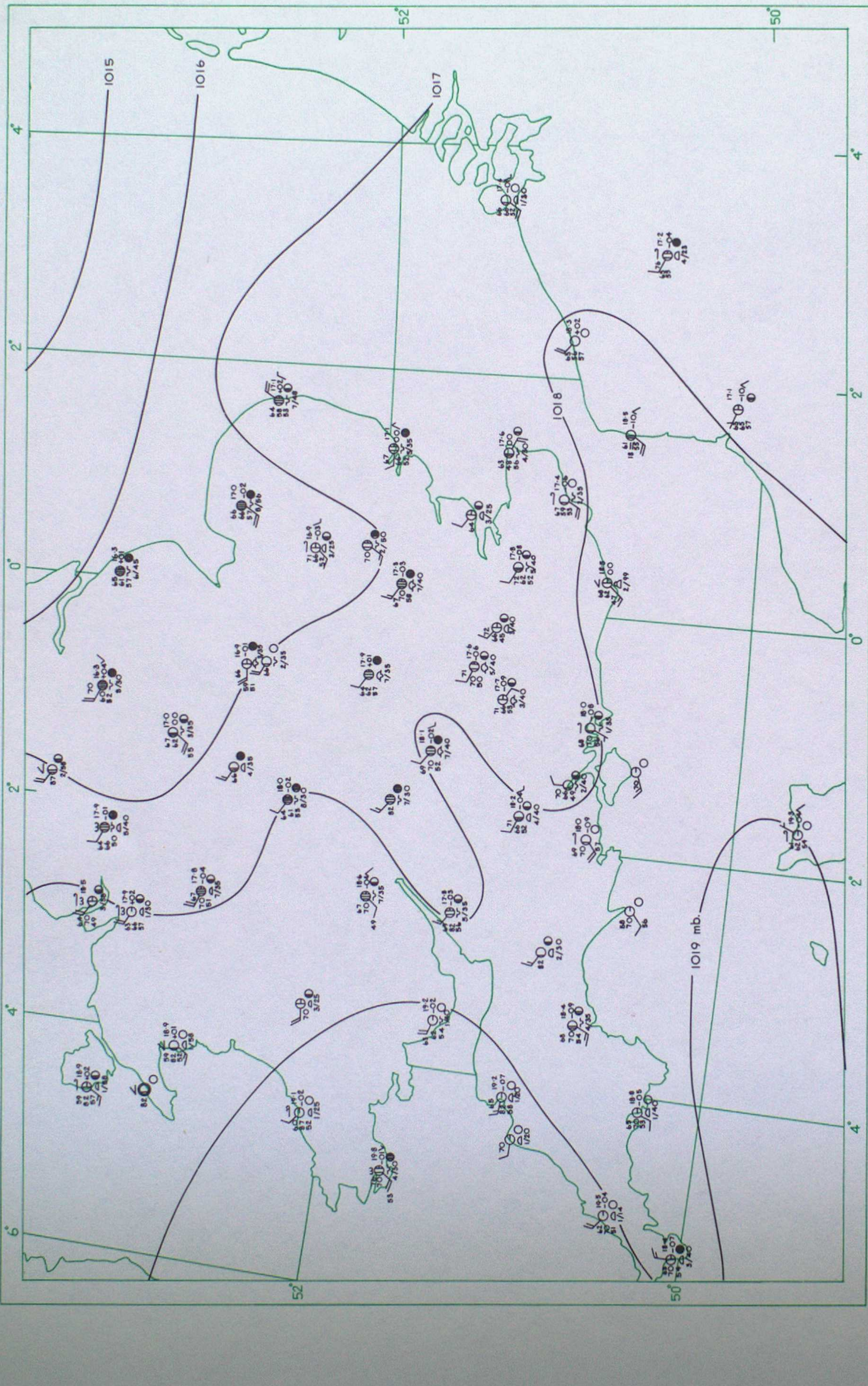


PLATE II Synoptic situation, 1500 G.M.T., 11 June 1952

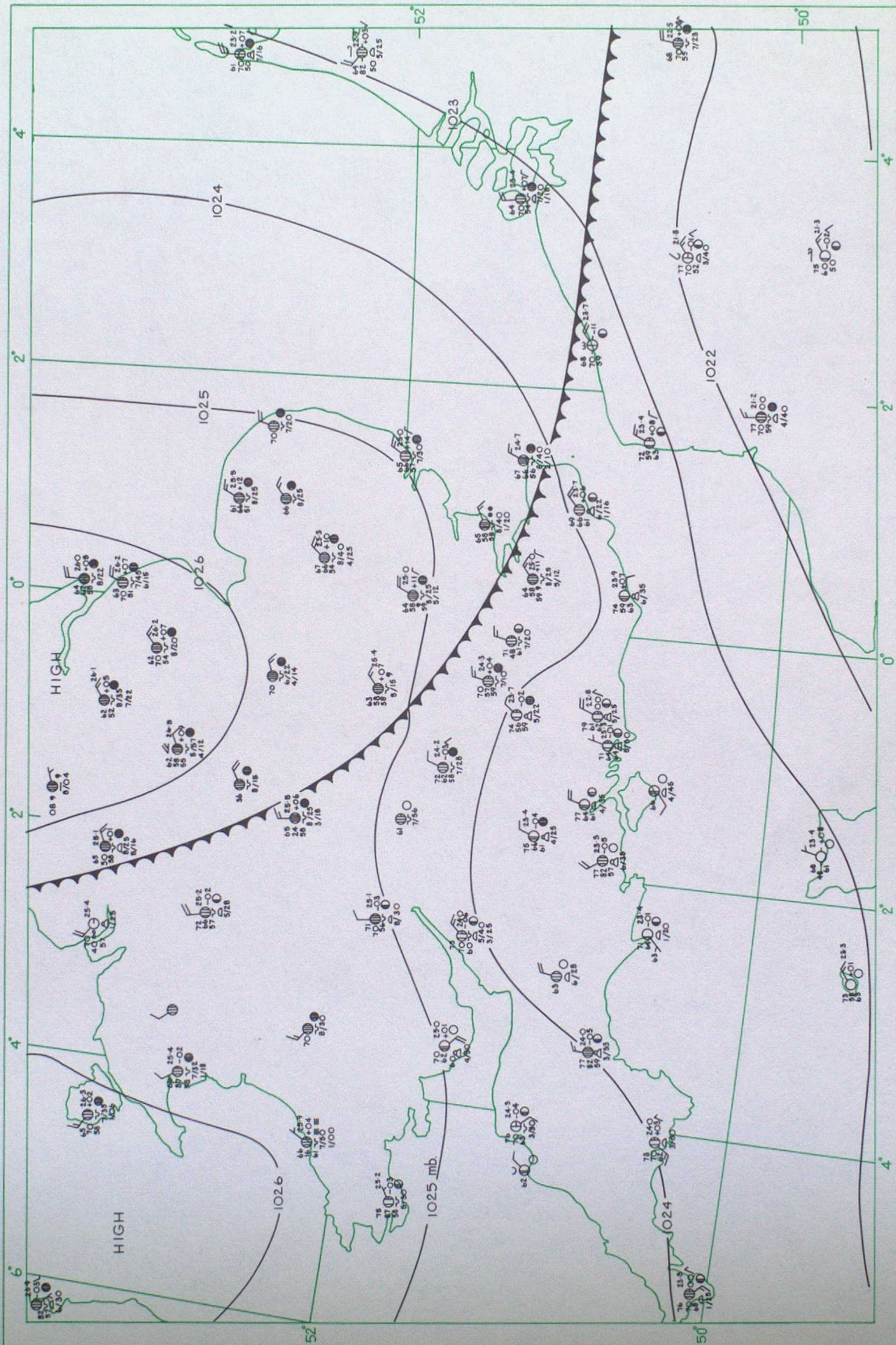


PLATE III Synoptic situation, 1200 GMT, 23 July 1952

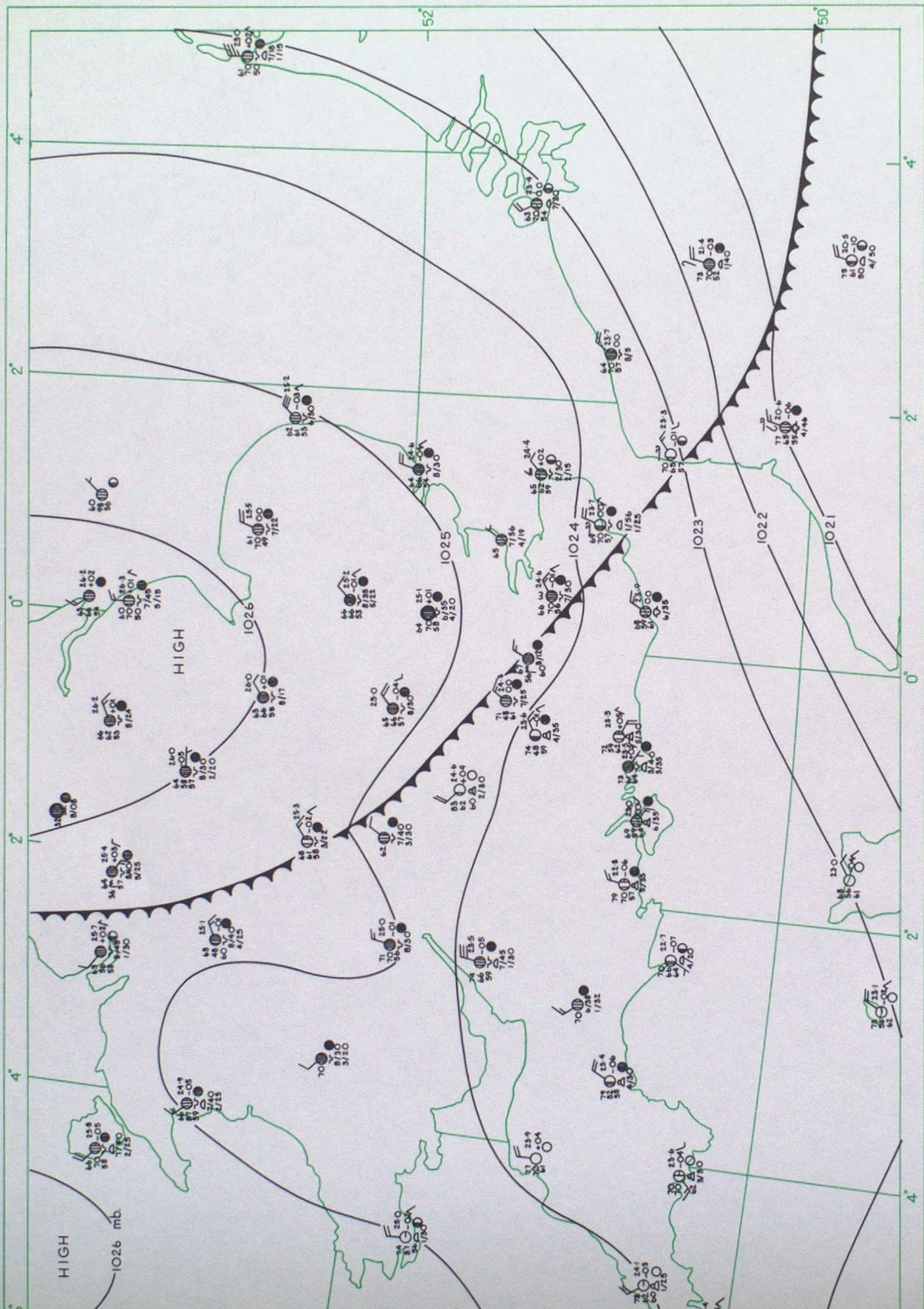


PLATE IV Synoptic situation, 1500 G.M.T, 23 July 1952

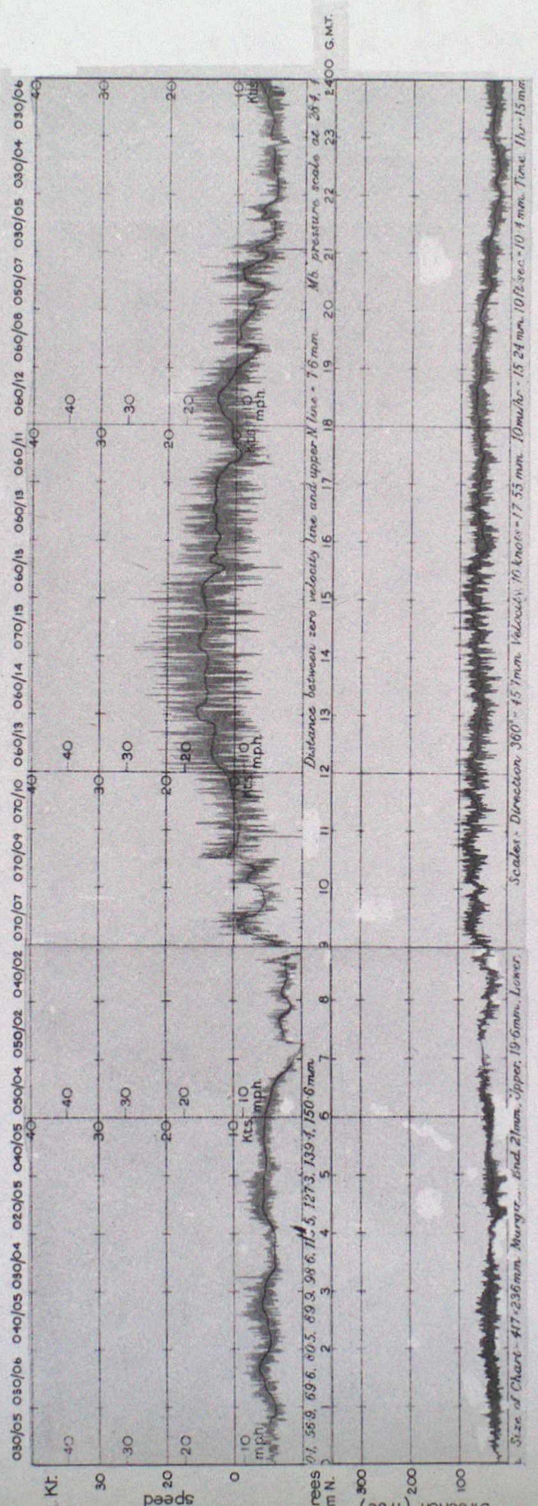


PLATE V Anemogram for Mildenhall, 13 April 1958

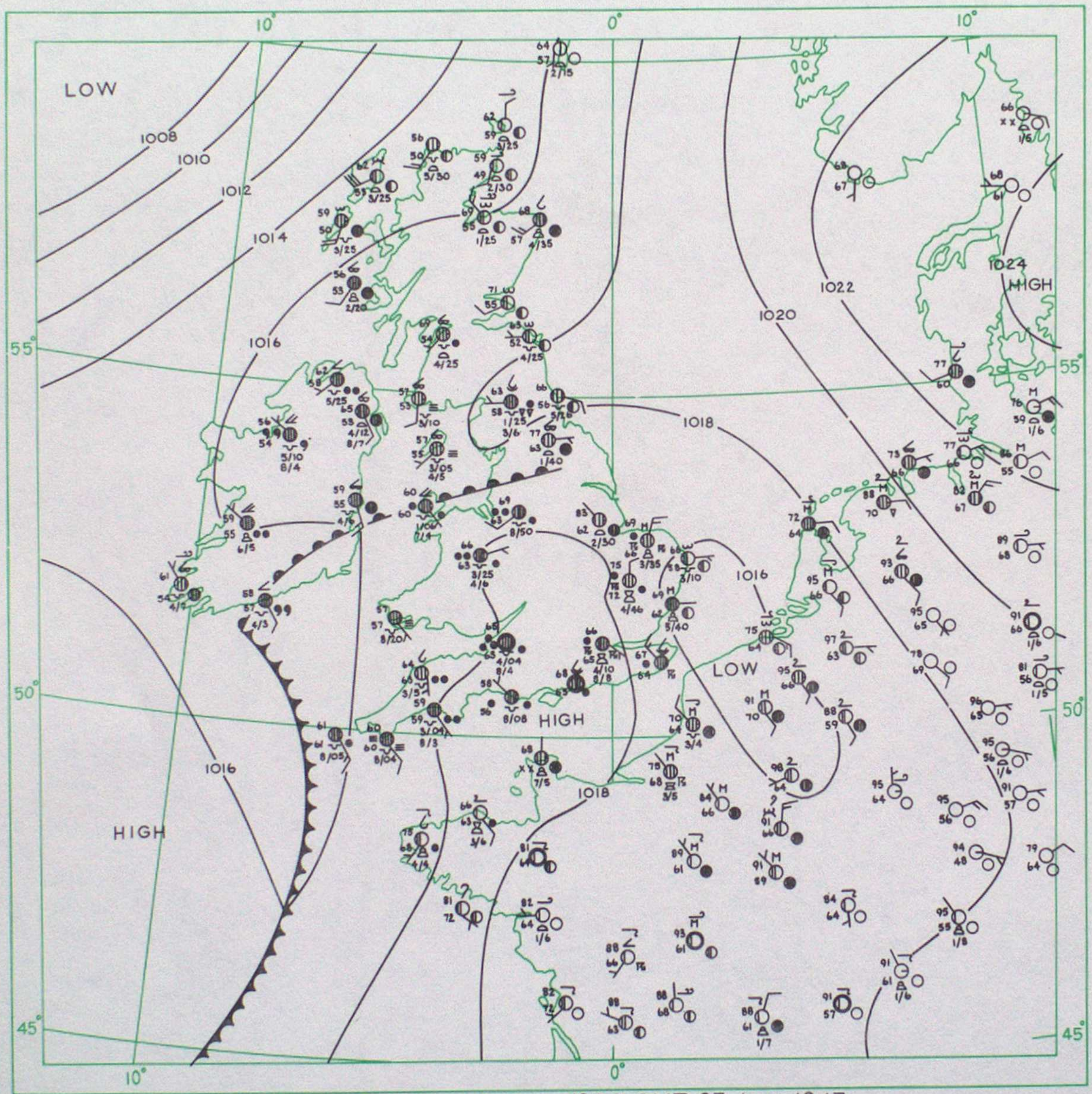


PLATE VI Synoptic situation, 1200 G.M.T., 27 June 1947

PLATE VII Synoptic situation, 1800 G.M.T., 4 February 1959

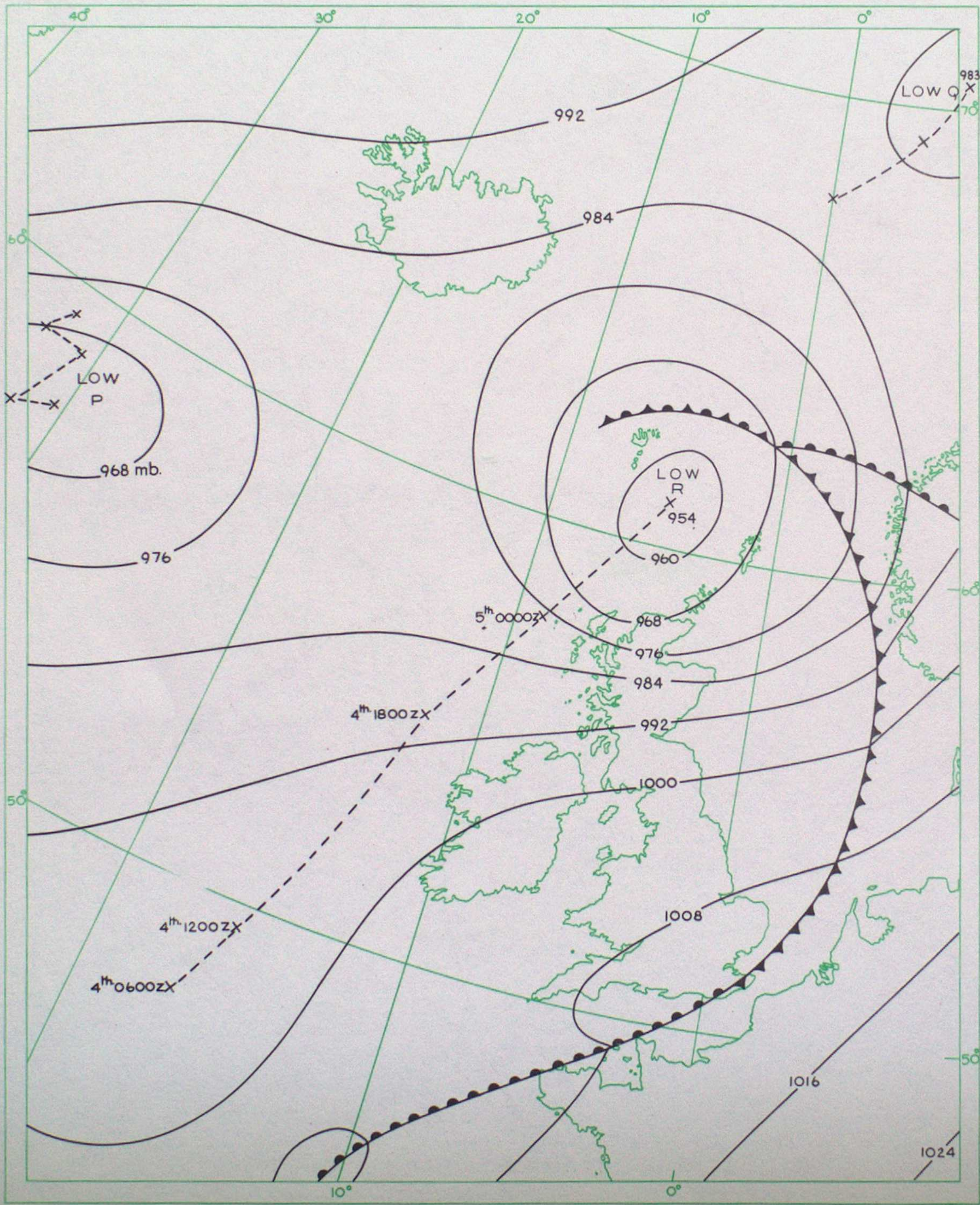


PLATE VIII Synoptic situation, 0600 G.M.T., 5 February 1959

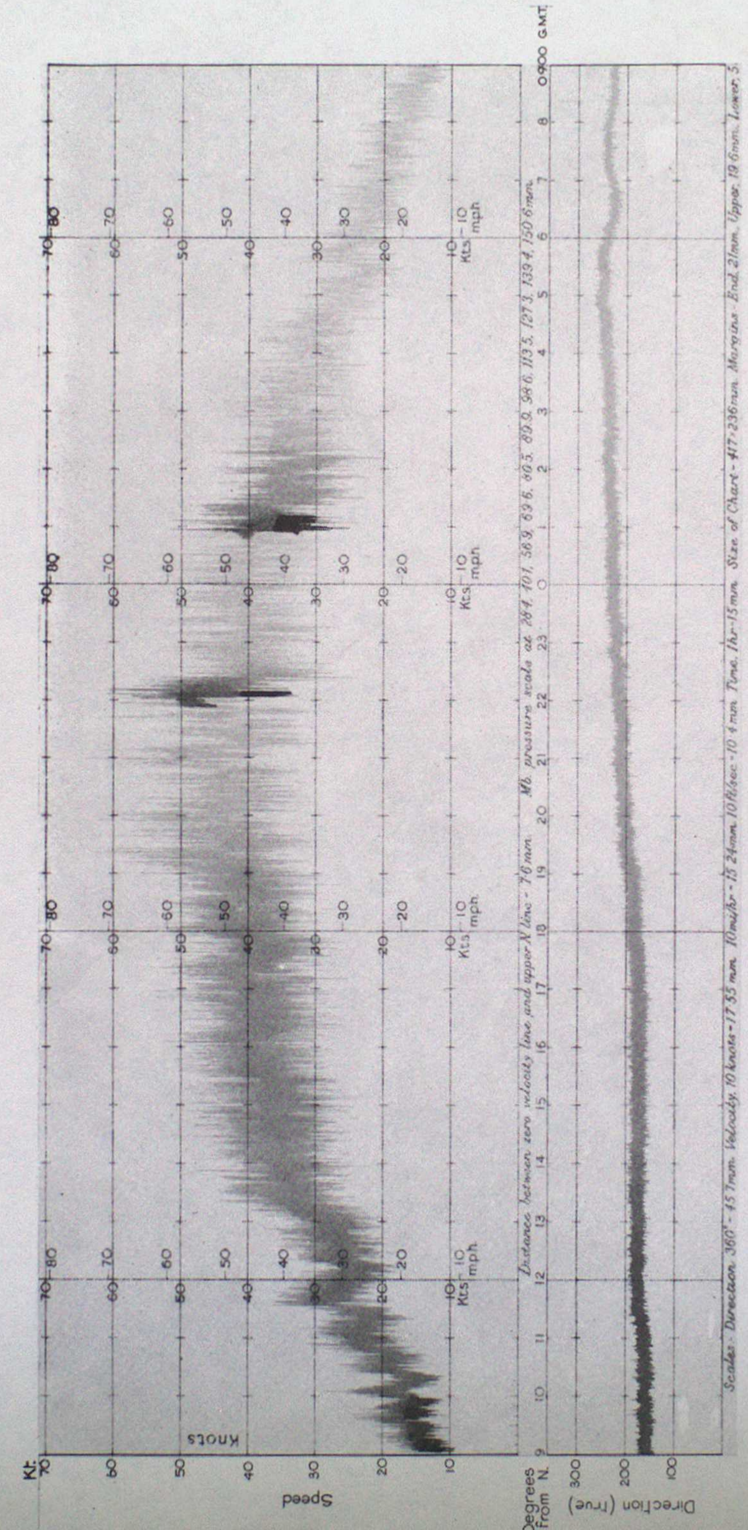


PLATE IX Anemogram for Valley, 0900 G.M.T, 4 February to 0900 G.M.T, 5 February 1957

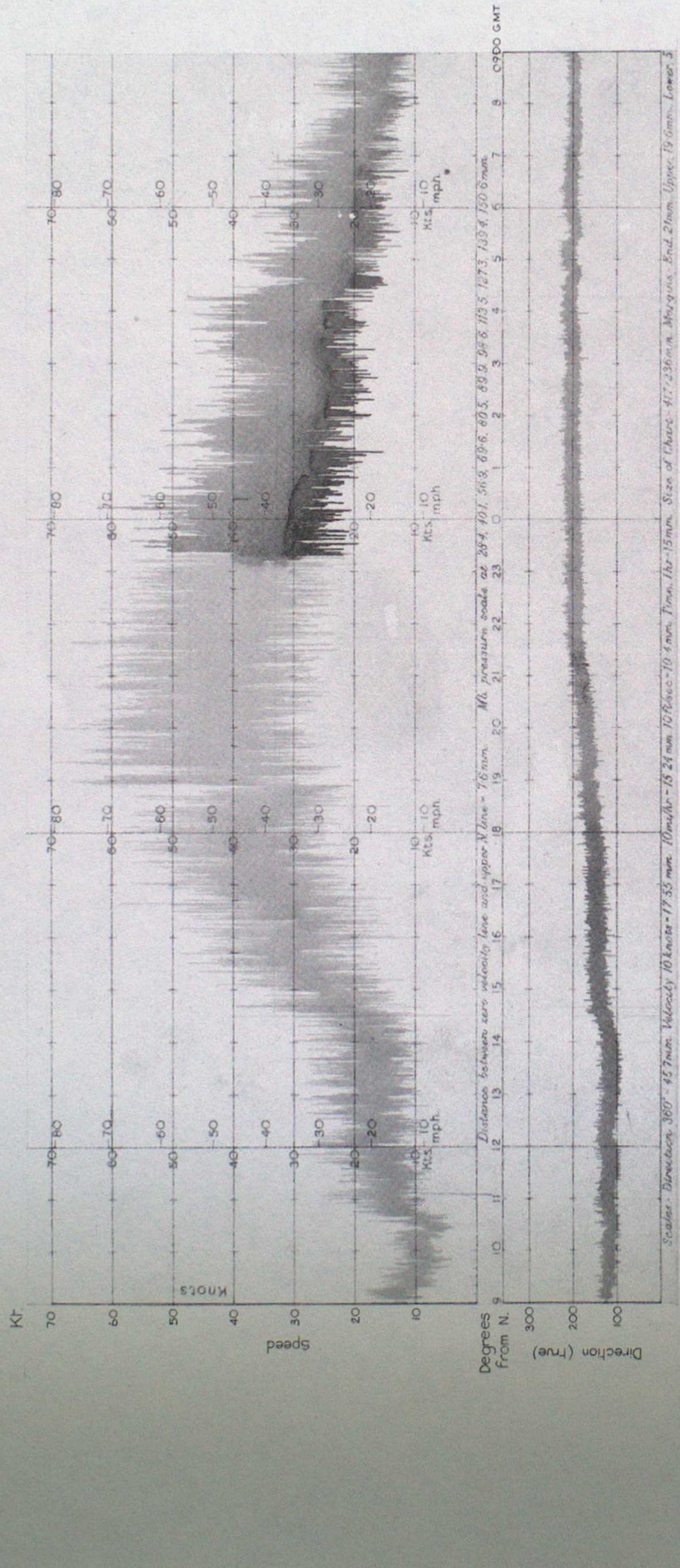


PLATE X Anemogram for Aldergrove, 0900 G.M.T,
4 February to 0900 G.M.T, 5 February 1957.

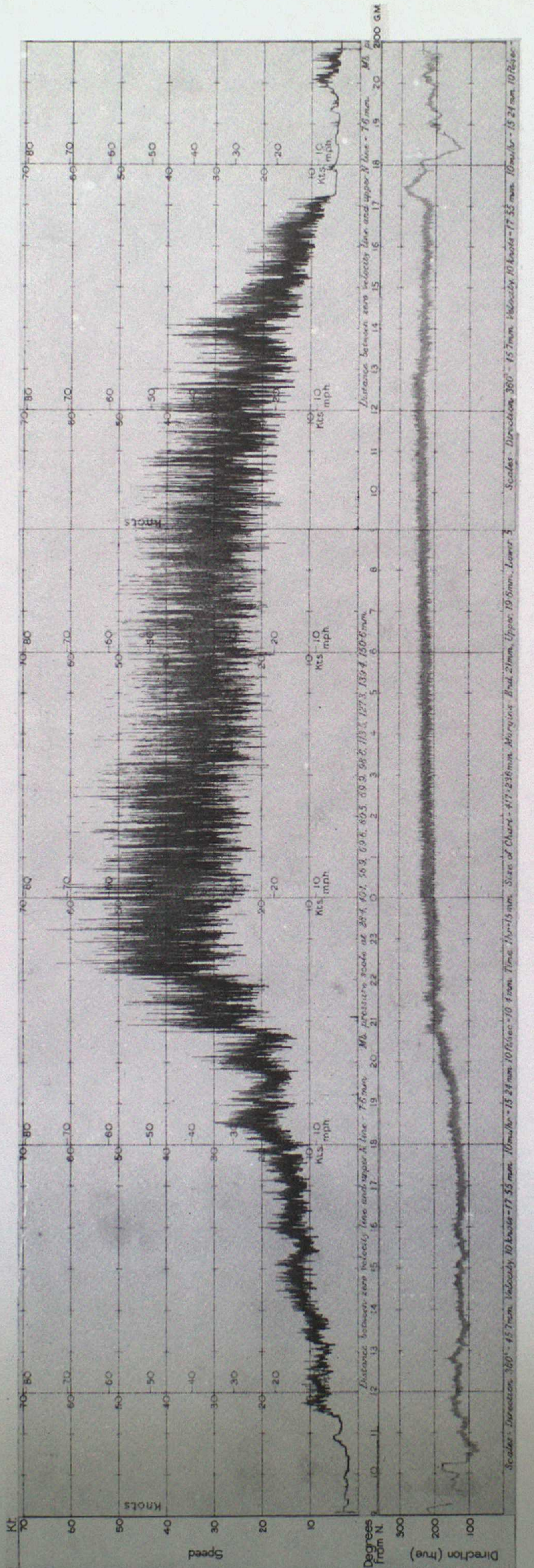


PLATE XI Anemogram for Leuchars, 0900 G.M.T, 4 February to 2100 G.M.T, 5 February 1957

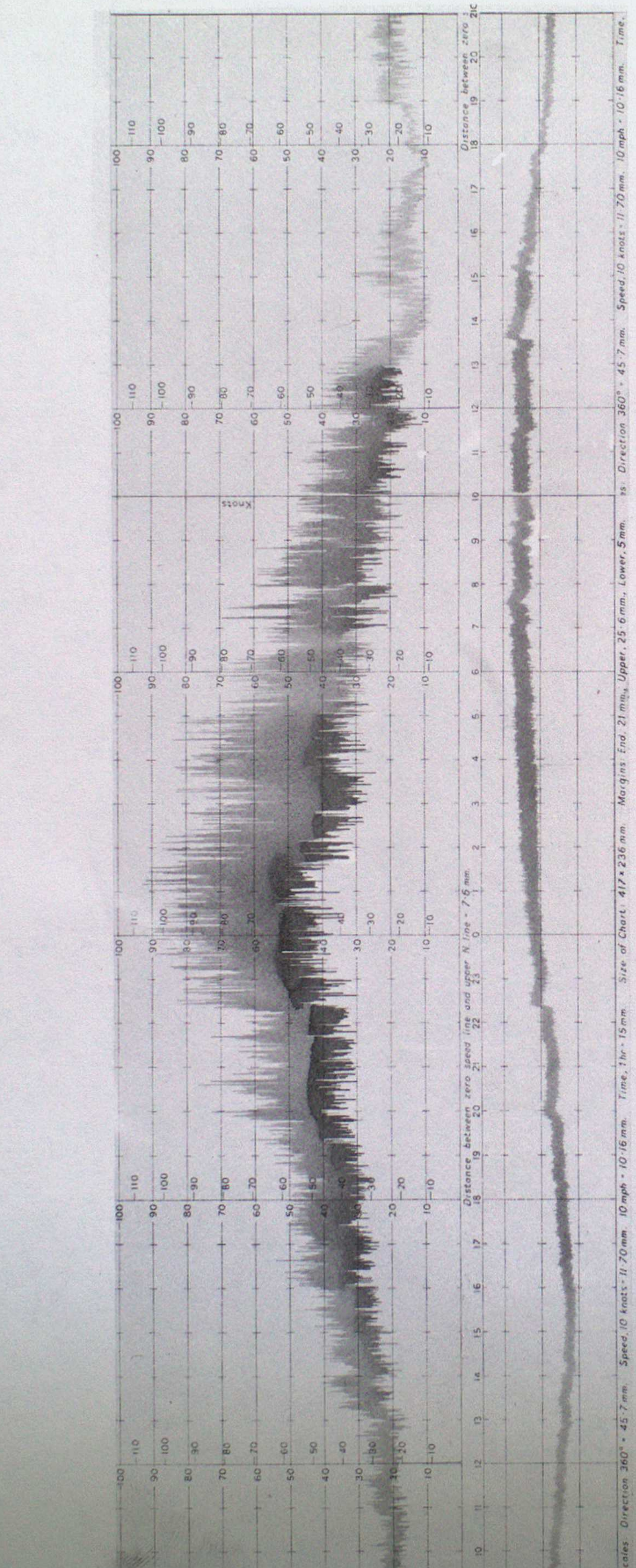


PLATE XII Anemogram for Tiree, 0900 GMT,
4 February to 2100 GMT, 5 February 1957

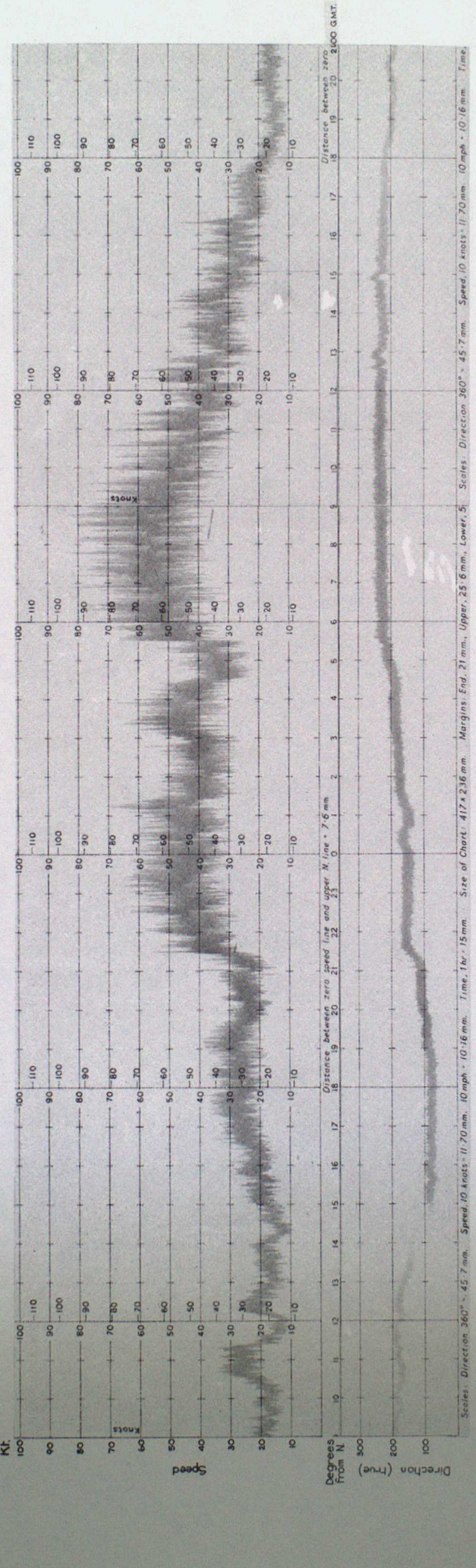


PLATE XIII Anemogram for Stornoway, 0900 GMT, 4 February to 2100 GMT, 5 February 1957

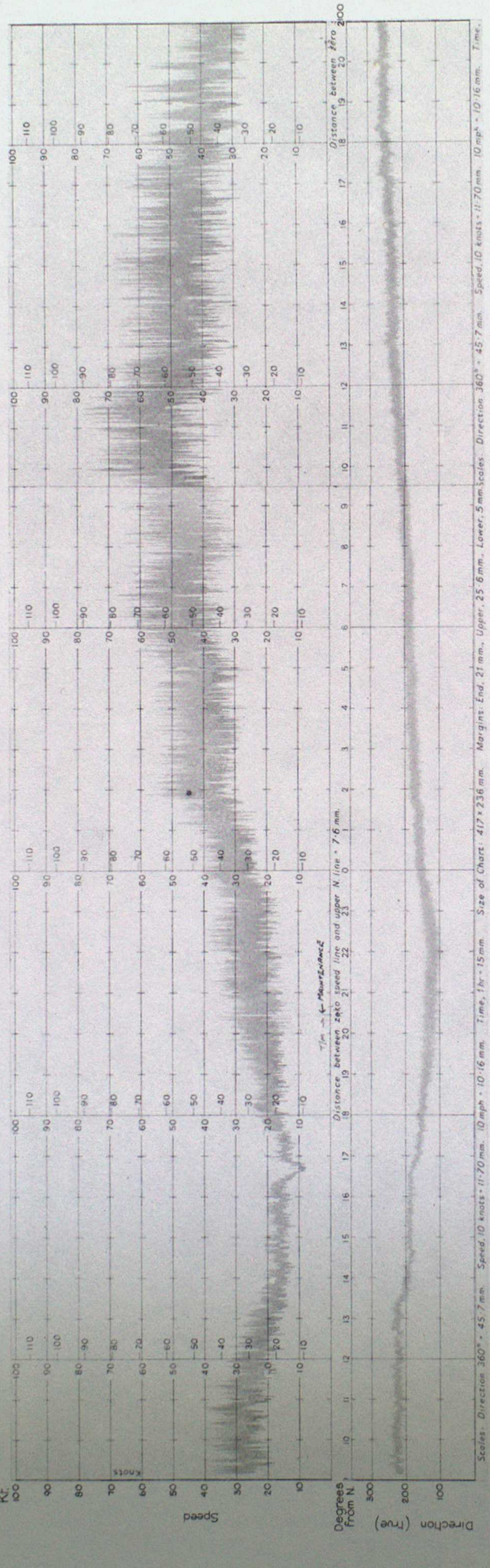


PLATE XIV Anemogram for Lerwick, 0900 G.M.T.,
4 February to 2100 G.M.T., 5 February 1957

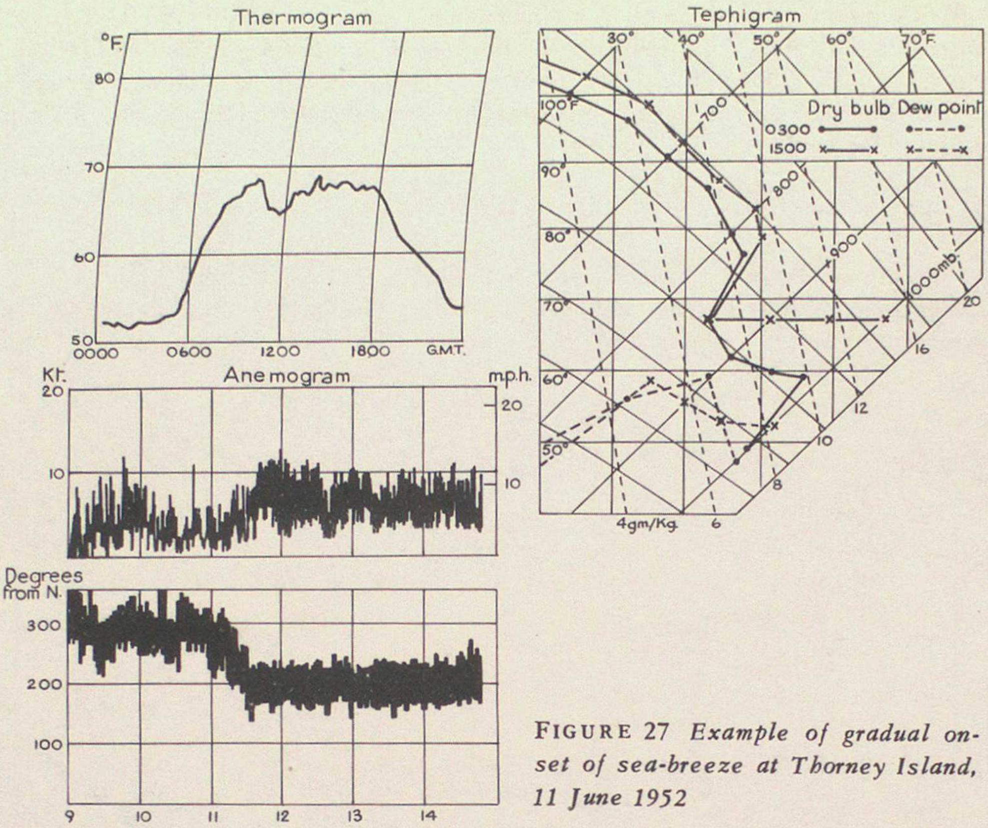


FIGURE 27 *Example of gradual onset of sea-breeze at Thorney Island, 11 June 1952*

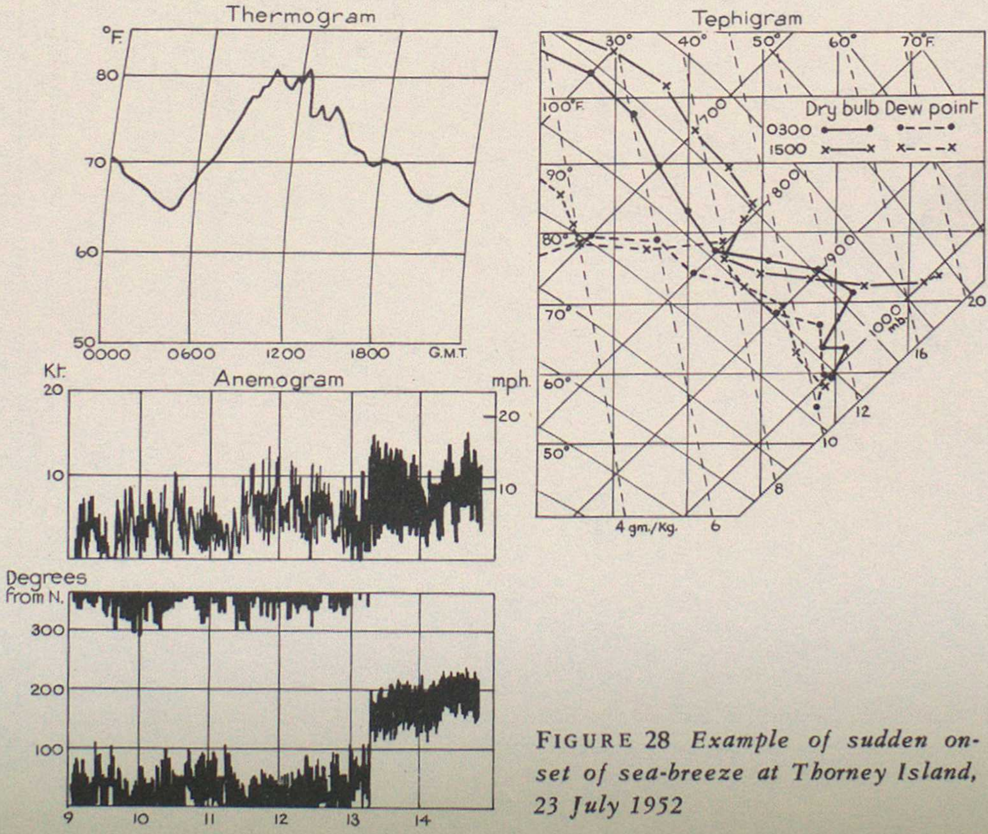


FIGURE 28 *Example of sudden onset of sea-breeze at Thorney Island, 23 July 1952*

spite of the increase in stability, a sudden onset occurred shortly after 1300 G.M.T. as can be seen from the anemogram and thermogram for Thorney Island included in Figure 28. The 1500 G.M.T. detailed chart for southern England shown in Plate IV indicates that sea-breezes were reported from a few other stations near the south coast of England.

13.13.2. *Diurnal variation of surface wind at Mildenhall, 13 April 1958*

Figure 29 shows the synoptic situation at 0600 G.M.T., 13 April 1958. The anticyclone, M, near Ireland, had been dominating the weather over England for the preceding two days. During the 13th the central pressure in the anticyclone

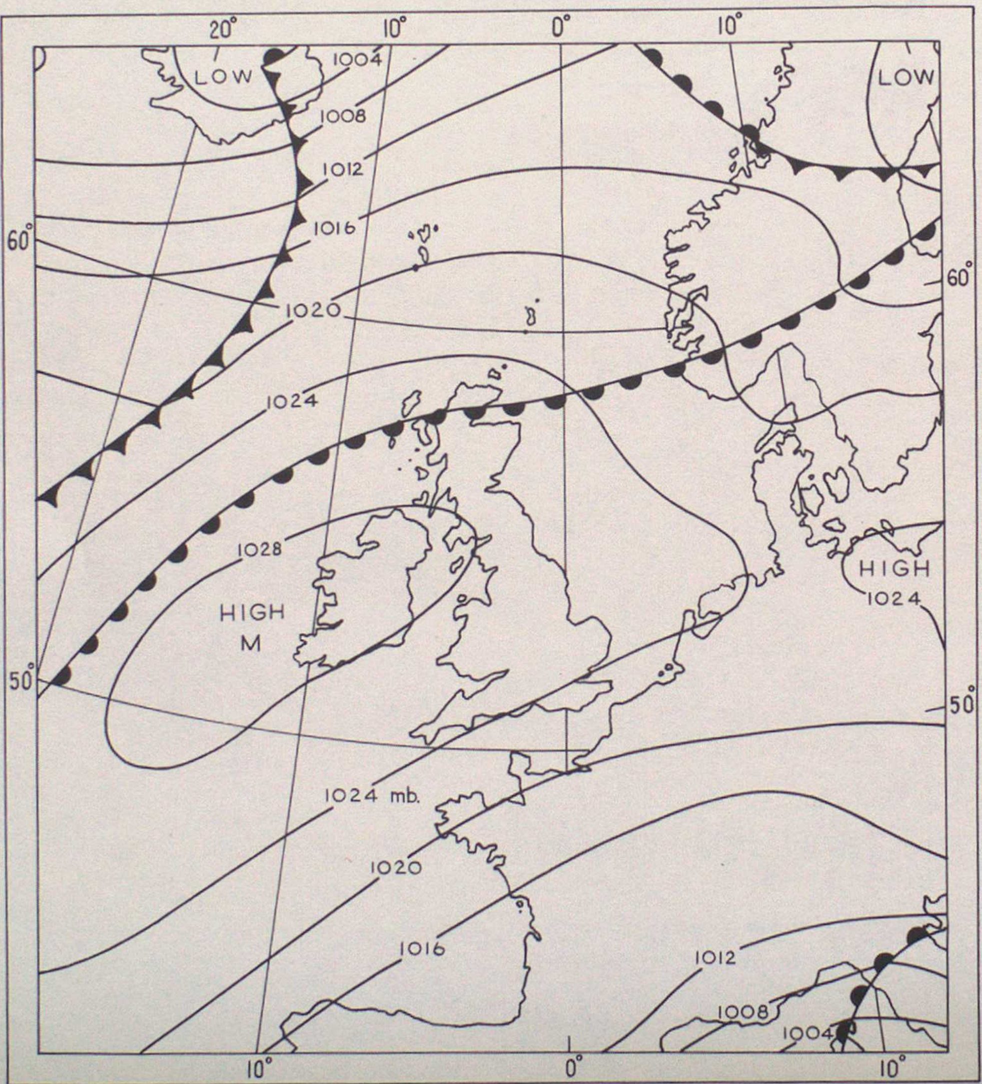


FIGURE 29 *Synoptic situation, 0600 G.M.T., 13 April 1958*

declined slightly and the system drifted west-south-westwards as the fronts on its northern flanks moved across Scandinavia and the North Sea. The pressure gradient in the region of Mildenhall was almost constant during the day on the 13th (the corresponding geostrophic wind being 070° 16-17 knots). During the evening of the 13th the direction of the isobars backed slightly (about 10°) and there was a tendency for the gradient to increase. At Mildenhall on the 13th there was a rather more than half-cover of stratocumulus with base 2,500 feet or above, some 5·4 hours of sunshine and the minimum and maximum temperatures in the screen were 29° and 50°F. respectively. Plate V shows the anemogram for Mildenhall for the whole of the 13th (values of the hourly winds are shown above the trace). Until about 0900 G.M.T. on the 13th mean surface winds were mainly between 030 and 050°, speeds generally 5 knots or below. After 0900 G.M.T. the trace shows a noticeable veer and increase in speed. Between 1400 and 1500 G.M.T. the strongest hourly wind was recorded but between 1300 and 1700 G.M.T. speeds averaged 13 to 15 knots from mean directions between 060° and 070°. Subsequently wind speeds fell away and the direction backed and from 2100 to 2359 G.M.T., the mean hourly wind was from 030°, speeds between 4 and 6 knots - values very close to those recorded in the early hours of the 13th.

13.13.3. *A squall, 27 June 1947*

Plate VI shows the synoptic situation at 1200 G.M.T., 27 June 1947. (The series of thunderstorms which occurred in this situation have been described by Douglas.[59]) The main front was off Scilly at 1200 G.M.T. and moving very slowly. The thunderstorms affecting eastern districts of England had developed over northern France early on the 27th near the eastern boundary of a region of widespread rain which, through cooling by evaporation, accentuated an already steep horizontal temperature gradient. The storms moved quickly north. There were a number of scattered early storms but the main storm reached the London area shortly before noon. Heavy rain fell and some squalls occurred in places in association with the main storm. Lewis[60] has described the events at Croydon as follows: "At Croydon, a violent squall accompanied by a thunderstorm and heavy rain occurred between 1100 and 1115 G.M.T. At 1110 the barometer rose 7mb. and then fell 5mb. Temperature dropped 12°F., approximately 1 in. of rain fell in about 15 minutes. The wind which had been light suddenly increased to gale force and gusted up to 73 m.p.h. (63 kt.) and, a few minutes before the squall, backed from roughly NW. to S. and then veered rather more gradually to NW. again." Tracings of the autographic records are shown in Figure 30.

At Kingsway, in central London, the squall gave a maximum gust of 56 knots from the south-south-west at 1120 G.M.T. with a sharp rise of pressure of 2 millibars. At Kew Observatory a peak gust of 46 knots from 230° occurred at 1115 G.M.T. However at South Farnborough the peak gust did not quite reach 17 knots from the south-south-west at 1050 G.M.T. and at Dunstable the extreme was just over 17 knots also from the south-south-west.

The upper winds at about 700 millibars were strong from a direction around south-south-west which is close to that of the peak gusts. If the descending air in the squall originated at these upper levels it would have considerable initial momentum which would add to the severity of the squall. Experience shows that the upper winds around the 700 - 600-millibar levels are usually strong when severe squalls occur.

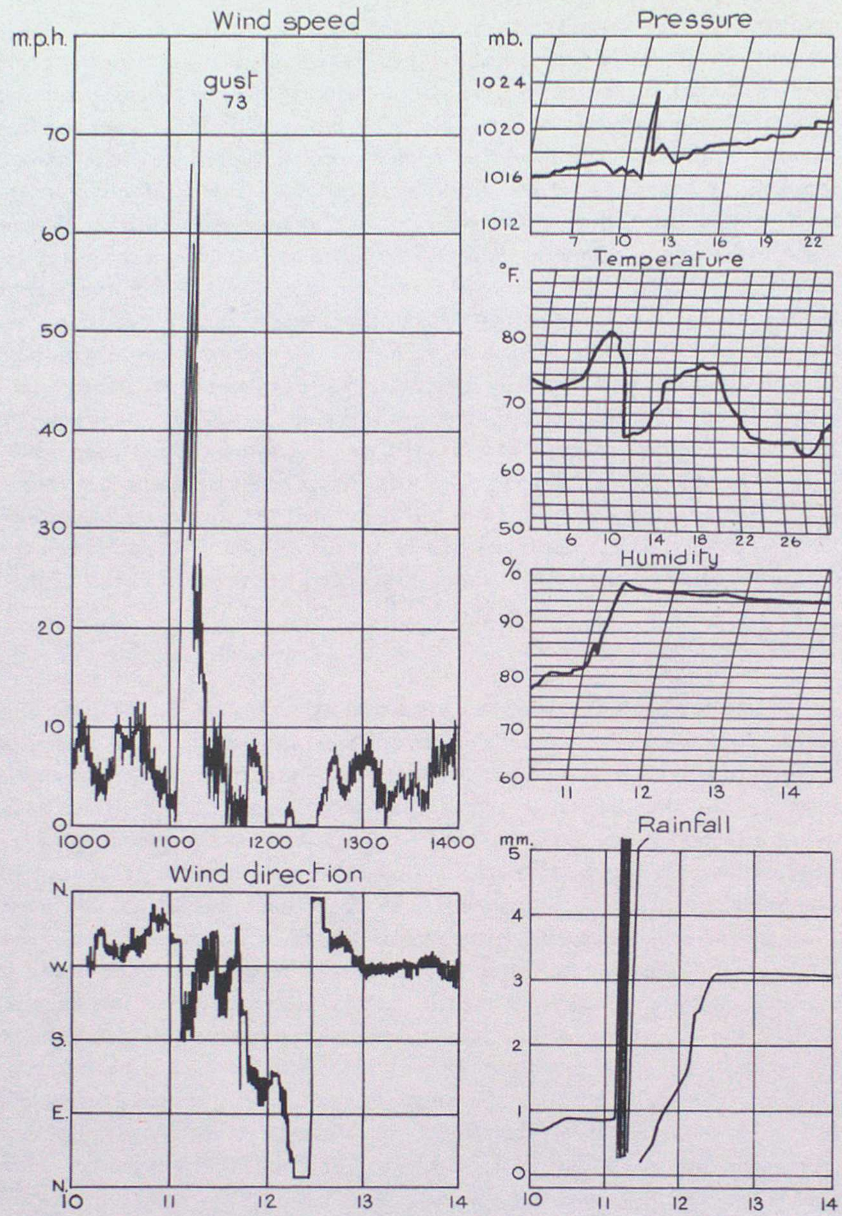


FIGURE 30 *Autographic records at Croydon, 27 June 1947*

Note:- The time scales of the barograph, thermograph and wind direction graphs have been adjusted to allow for the slowness of the clocks.

13.13.4. *Gales on 4 and 5 February 1957*

Plate VII shows the synoptic situation at 1800 G.M.T., 4 February 1957. The deep vigorous low, LR, just off north-west Ireland, had been located to the south of Newfoundland at midday on the 2nd when it was a wave depression. It moved at about 45 knots on an east-north-easterly track until late on the 3rd when, at 1800 G.M.T., it was centred at 47°N.27°W. Subsequently the track became progressively more north-easterly and later north-north-easterly and the speed of movement remained high at about 40 knots. The depression had been deepening since the 2nd and continued to do so until about 1800 G.M.T., 4 February, at about which time it attained its greatest depth. Shortly after 1800 G.M.T. the depression occluded rapidly. Movement toward the north-north-east was maintained at a fairly high speed of 35 knots until 0600 G.M.T., 5 February, after which the depression slowed down considerably. Plate VIII shows the synoptic situation at 0600 G.M.T. on the 5th, by which time the depression was centred to the north of Scotland and the associated frontal system had cleared all districts of the British Isles except southern England.

Gales were widespread in Scotland and Northern Ireland. A selection of traces of the anemograms for Valley, Aldergrove, Leuchars, Tiree, Stornoway and Lerwick are shown in Plates IX to XIV. These illustrate the severity of the gales. Changes of wind with frontal passage can be recognized on some of the traces. At Tiree a mean hourly wind of 60 knots was recorded and gusts reached 93 knots, the highest gust in February since records began at Tiree in 1927. Late on the 4th a gust of over 70 knots was recorded at Aldergrove. On the 5th gusts reached 87 knots at Stornoway. Behind the cold front where the gradient was west-south-west Leuchars, near the lee coast of Scotland, recorded a gust of 65 knots. At Lerwick, closer to the track of the centre, hourly winds exceeded gale force for 20 consecutive hours. Gusts of 70 knots or more were recorded for 11 consecutive hours, the highest gust being 83 knots. At Valley, well away from the track of the centre but exposed to winds from the Irish Sea, mean hourly winds reached or exceeded 40 knots for 8 consecutive hours and the highest gust was 68 knots.

Table 24 shows the mean hourly wind and the highest gust recorded at six locations when the mean hourly wind reached or exceeded 34 knots.

TABLE 24 *Mean hourly wind and highest gust recorded at six locations when the mean hourly wind reached or exceeded 34 knots during the gale of 4-5 February 1957*

Mid-time of hourly period (knots)

4 February 1957

	1330	1430	1530	1630	1730	1830	1930	2030	2130	2230	2330
Valley	34	37	38	40	40	41	46	45	44	43	41
	48	52	53	54	57	61	68	64	66	61	61
Aldergrove	..	..	..	..	37	35	45	45	44	42	37
	..	..	..	..	58	67	69	68	71	68	62
Leuchars	..	..	..	..	..	..	..	..	..	40	39
	..	..	..	..	..	..	..	..	..	63	65
Tiree	..	..	..	35	36	36	43	46	47	55	59
	..	..	..	54	51	55	73	77	73	83	91
Stornoway	..	..	..	..	..	..	..	..	34	44	47
	..	..	..	..	..	..	..	..	67	65	67
Lerwick	..	..	..	..	..	..	..	..	..	..	..
	..	..	..	..	..	..	..	..	..	..	..

5 February 1957

	0030	0130	0230	0330	0430	0530	0630	0730	0830	0930	1030	1130	1230	1330	1430	1530	1630	1730	1830	1930	2030
Valley	40	36	34	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..
	57	52	47	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..
Aldergrove	35	34	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..
	62	52	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..
Leuchars	39	36	36	35	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..
	64	61	59	56	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..
Tiree	60	59	53	50	50	45	43	37	36	34	..	..	..	..	..	..	..	..	..	..	..
	92	93	88	84	84	75	71	69	59	51	..	..	..	..	..	..	..	..	..	..	..
Stornoway	47	44	40	46	39	47	55	57	55	55	49	44	42	35	35	..	..	..	..	..	..
	66	71	58	68	58	69	87	84	84	78	71	68	62	48	52	..	..	..	..	..	..
Lerwick	..	37	41	40	41	45	48	46	46	46	56	55	50	48	49	47	47	44	42	42	35
	..	59	59	57	58	66	75	70	77	84	79	81	74	71	71	71	71	66	62	66	54

For each entry the upper figure is the mean hourly wind and the lower figure is the highest gust

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