Meteorological Office

forecasters' reference book

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PREFACE

The object of the *Forecasters' reference book* is to give the forecaster a set of rules, without background material, for day-to-day use. Practically all of the rules have been extracted from the *Handbook of weather forecasting*, Met.0.637, but a few have been extracted from more recent work. The book replaces the *Pocketbook for forecasters*.

Each rule is stated as concisely as possible and is followed by full instructions on any computations or use of tables and diagrams, together with notes on application. References are given.

The work is arranged in chapters according to subject, but it should be noted that the rules apply mainly to the region of the British Isles and north-west Europe and will not necessarily apply elsewhere. A list of contents by chapter and section at the beginning should facilitate rapid reference to any particular subject, cross-references within the text are provided where they are useful.

The terminology and symbols are those in common use but abbreviations are used rather more than is customary in other Office publications.

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CHAPTER 1. TEMPERATURE

1.1. TEMPERATURE RISE DURING THE DAY

1.1.1. *Temperature rise on clear days*

Alternative methods are given in 1.1.2. and 1.1.3. for estimating the rise of temperature at the surface after dawn on clear days. For both methods it is necessary to start with a tephigram on which is plotted a dawn temperature curve (actual or estimated) for the lowest few thousand feet of the air which will lie over the forecast area at the appropriate time during the morning or early afternoon. If cloud develops, an estimate has to be made of the reduction in the value of the temperatures as forecast by either method.

It should be noted that the values given in Tables 1.I. and 1.II. apply to southern England.

1.1.2. *Temperature rise* (Gold and Jefferson)

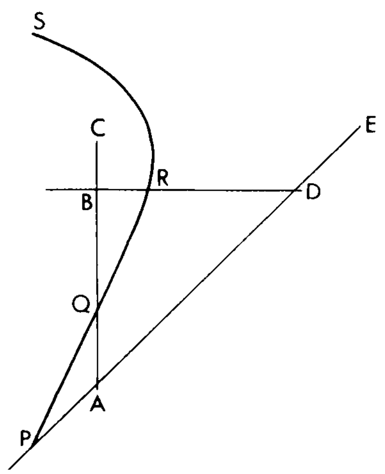


FIG. 1.1.(*a*) *The application of scales for assessing areas on the tephigram*

In Fig. 1.1.*(a)* the curve PQRS is the dawn temperature curve on a tephigram of which PE is the isobar for surface pressure. The forecast surface temperature is given by point D which is positioned so that the area between the dry adiabatic DR, the surface isobar, PD,

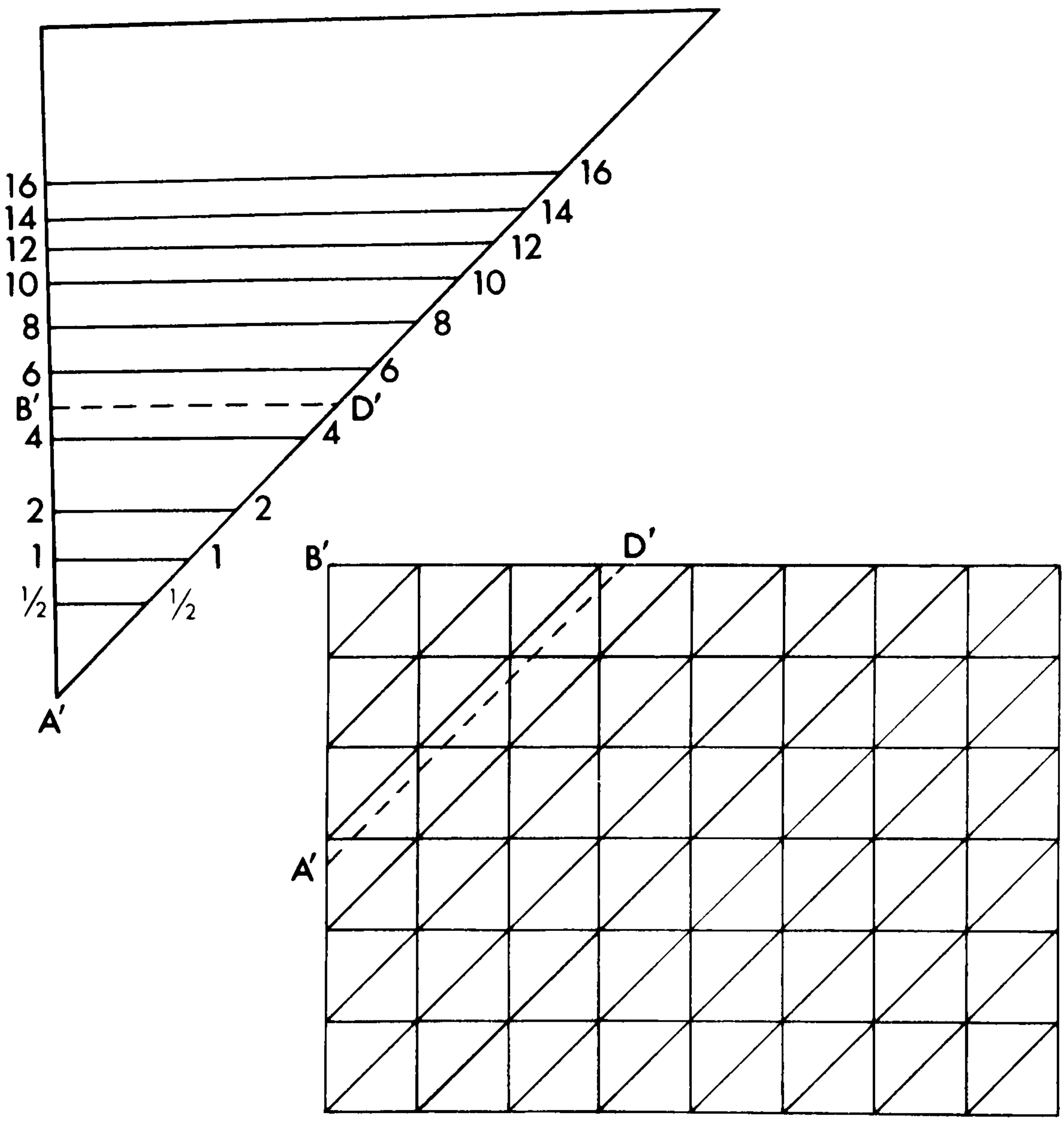


FIG. 1. 1. *(b) Two scales for assessing areas on the tephigram*

and the dawn temperature curve, PR, is that given by the appropriate entry in Table 1.I. This position D is most readily assessed by use of a scale such as that reproduced in Fig. 1. 1. *(b)*. A parallel on the scale is chosen (say B' D') to give a triangle A' B' D' of the area according to Table 1.I. The scale is placed over the tephigram and moved maintaining A' D' along the surface isobar (PE in Fig. 1. 1.(*a*)) until a position is found (in which A' , B' and D' correspond with A, B, and D) with area PQA equal to area BQR.

TABLE 1.1. *Areas on a tephigram corresponding to energy which is available for warming the lower atmosphere on a clear day between sunrise and 0800, 1000 and 1200 local time (southern England)*

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

Note

The above procedures combine the separate techniques due to Gold and Jefferson summarized in HWF 14.7.1.

Ref: Gold, *Prof. Notes*, No. 63, 1933.
Jefferson, *Met. Mag.*, Feb '50.

1.1.3. *Temperature rise* (Johnston)

TABLE 1.II. *Thickness of layer which is changed from an isothermal to an adiabatic state by insolation (southern England)*

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

Obtain from Table 1.II the layer thickness, Δp, for the month concerned and the time (hours from sunrise or time of maximum temperature). Mark the isobars p_0 and $(p_0 - \Delta p)$ on the tephigram (see Fig. 1.2.). Select a point P on the $(p_0 - \Delta p)$ isobar such that the area of the triangle PFE is equal to that enclosed by the dry adiabatic through P, the surface isobar, and the environment curve. Then F (the intersection of the dry adiabatic through P and the surface isobar) represents the temperature at the chosen time.

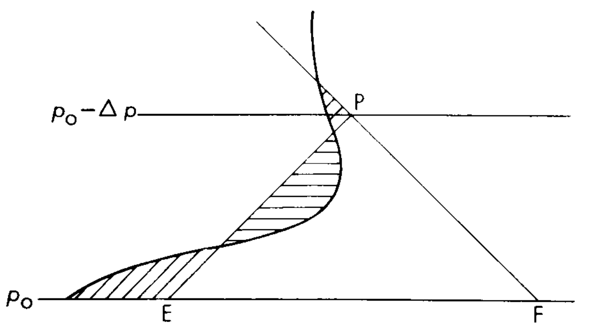


FIG. 1.2. *Estimation of maximum day temperature*

Note

Over damp soils, e.g. clay, the energy corresponding to the values given is not available for heating the lowest layers of the atmosphere during the first few hours of sunshine. In such localities, the values to the left of the lines marked A in Table 1.II. should be reduced by one-third and the values between the lines marked A and B should be reduced by one-fifth.

Ref: HWF 14.7.1.
Johnston, *Met.Mag.*, Sept.'58.
Jefferson, *Met.Mag.*, May '59.

1.1.4. *Temperature rise on foggy days*

(i) Using Table 1.II. for the month and the method of 1.1.3., draw a curve showing rise of temperature from sunrise, assuming a clear day (full line in Fig.1.3.).

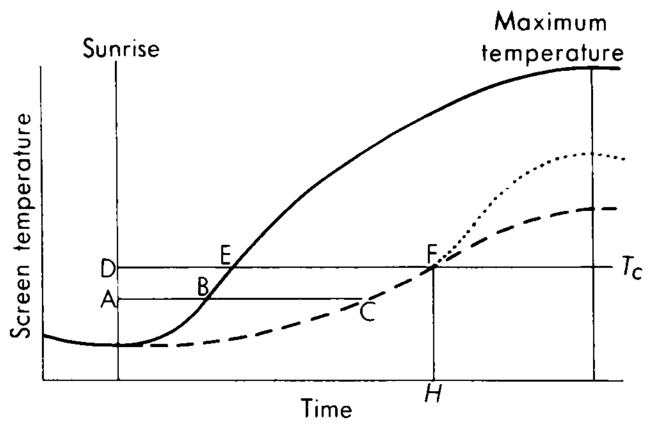


FIG. 1.3. *Construction of temperature curves*

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

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

$$AB : AC = DE : DF = \ldots = f,$$

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

(iii) If this curve reaches temperature, T_C, at which fog can be expected to disperse (see 2.6.), the curve then rises more steeply and is more in line with clear-day characteristics (e.g. dotted line in Fig. 1.3.).

Notes

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

(b) A diagram similar to Fig. 1.3. is used for estimating the time of clearance of radiation fog or low stratus (see 2.6.).

Ref: HWF 17.8.
Jefferson, *Met.Mag.*, Apr.'50.

1.2. TEMPERATURE FALL DURING THE NIGHT

1.2.1. *Minimum temperature* (Craddock and Pritchard)

If T_{12} = screen temperature at 1200 GMT (°C),

T_{d12} = dew-point temperature at 1200 GMT (°C),

then $T_{min} = 0.316\, T_{12} + 0.548\, T_{d12} - 1.24 + K$

$= X + K$.

Values of K and X can be found from Tables 1.III and 1.IV.

TABLE 1.III. *Values of K*

Mean* geostrophic wind speed	Mean* cloud amount (oktas)			
	0 to 2	2+ to 4	4+ to 6	6+ to 8
knots	*degrees Celsius*			
0-12	−2.2	−1.7	−0.6	0
13-25	−1.1	0	+0.6	+1.1
26-38	−0.6	0	+0.6	+1.1
39-51	+1.1	+1.7	+2.8	

*Mean of forecast values for 1800, 0000 and 0600 GMT.

TABLE 1.IV. *Values of X*

T_{d12} / T_{12}	−3	−2	−1	0	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16°C
°C									degrees Celsius											
3	−1.9	−1.4	−0.8	−0.3	+0.3	0.8	1.4													
4	−1.6	−1.1	−0.5	+0.0	0.6	1.1	1.7	2.2												
5	−1.3	−0.8	−0.2	0.3	0.9	1.4	2.0	2.5	3.1											
6	−1.0	−0.4	+0.1	0.7	1.2	1.8	2.3	2.8	3.4	3.9										
7	−0.7	−0.1	0.4	1.0	1.5	2.1	2.6	3.2	3.7	4.3	4.8									
8	−0.4	+0.2	0.7	1.3	1.8	2.4	2.9	3.5	4.0	4.6	5.1	5.7								
9	−0.0	0.5	1.1	1.6	2.2	2.7	3.2	3.8	4.3	4.9	5.4	6.0	6.5							
10	+0.4	0.8	1.4	1.9	2.5	3.0	3.6	4.1	4.7	5.2	5.8	6.3	6.9	7.4						
11	+0.7	1.1	1.7	2.2	2.8	3.3	3.9	4.4	5.0	5.5	6.1	6.6	7.2	7.7	8.3					
12	1.0	1.5	2.0	2.6	3.1	3.6	4.2	4.7	5.3	5.8	6.4	6.9	7.5	8.0	8.6	9.1				
13	1.3	1.8	2.3	2.9	3.4	4.0	4.5	5.1	5.6	6.2	6.7	7.3	7.8	8.3	8.9	9.4	10.0			
14	1.6	2.1	2.6	3.2	3.7	4.3	4.8	5.4	5.9	6.5	7.0	7.6	8.1	8.7	9.2	9.8	10.3	10.9		
15	1.9	2.4	3.0	3.5	4.0	4.6	5.1	5.7	6.2	6.8	7.3	7.9	8.4	9.0	9.5	10.1	10.6	11.2	11.7	
16	2.3	2.7	3.3	3.8	4.4	4.9	5.5	6.0	6.6	7.1	7.7	8.2	8.7	9.3	9.8	10.4	10.9	11.5	12.0	12.6
17	2.6	3.0	3.6	4.1	4.7	5.2	5.8	6.3	6.9	7.4	8.0	8.5	9.1	9.6	10.2	10.7	11.3	11.8	12.4	12.9

TABLE 1.IV. *(contd)*

T_{d12} / T_{12}	−3	−2	−1	0	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16°C
°C											*degrees Celsius*									
18	2.9	3.4	3.9	4.4	5.0	5.5	6.1	6.6	7.2	7.7	8.2	8.8	9.4	9.9	10.5	11.0	11.6	12.1	12.7	13.2
19	3.2	3.7	4.2	4.8	5.3	5.9	6.4	7.0	7.5	8.1	8.6	9.1	9.7	10.2	10.8	11.3	11.9	12.4	13.0	13.5
20	3.5	4.0	4.5	5.1	5.6	6.2	6.7	7.3	7.8	8.4	8.9	9.5	10.0	10.6	11.1	11.7	12.2	12.8	13.3	13.8
21	3.8	4.3	4.8	5.4	5.9	6.5	7.0	7.6	8.1	8.7	9.2	9.8	10.3	10.9	11.4	12.0	12.5	13.1	13.6	14.2
22	4.1	4.6	5.2	5.7	6.3	6.8	7.4	7.9	8.5	9.0	9.5	10.1	10.6	11.2	11.7	12.3	12.8	13.4	13.9	14.5
23	4.5	4.9	5.5	6.0	6.6	7.1	7.7	8.2	8.8	9.3	9.9	10.4	11.0	11.5	12.1	12.6	13.2	13.7	14.2	14.8
24	4.8	5.2	5.8	6.3	6.9	7.4	8.0	8.5	9.1	9.6	10.2	10.7	11.3	11.8	12.4	12.9	13.5	14.0	14.6	15.1
25	5.1	5.6	6.1	6.7	7.2	7.8	8.3	8.9	9.4	9.9	10.5	11.0	11.6	12.1	12.7	13.2	13.8	14.3	14.9	15.4
26	5.4	5.9	6.4	7.0	7.5	8.1	8.6	9.2	9.7	10.3	10.8	11.4	11.9	12.5	13.0	13.6	14.1	14.6	15.2	15.7
27	5.7	6.2	6.7	7.3	7.8	8.4	8.9	9.5	10.0	10.6	11.1	11.7	12.2	12.8	13.3	13.9	14.4	15.0	15.5	16.1

T_{12} temperature at 1200 GMT

T_{d12} dew-point at 1200 GMT

Notes

(a) The formula given by Craddock and Pritchard was derived from the combined data for 16 widely separated stations in England, all more than 10 miles inland. For forecasting night minimum temperatures at a particular place accuracy is likely to be improved by consideration of local data, for instance local values might be found for K.

(b) The formula applies to nights without fog.

(c) If the ground is snow covered the actual minimum is likely to be lower than the minimum forecast by means of this technique or other techniques described in 1.2.

Ref: HWF 14.7. 2.1.
Craddock and Pritchard, MRP 624.

1.2.2. *Minimum temperature* (Boyden)

If T_w = wet-bulb temperature at the time of T_{max} (°C)

T_d = dew-point temperature at the time of T_{max} (°C)

then $T_{min} = \frac{1}{2}(T_w + T_d) - K$.

TABLE 1.V. *Values of K at Kew*

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

Notes

(a) Table 1.V. (for Kew) applied for nights on which fog did not form.

(b) See cautionary note *(c)* in 1.2.1.

Ref: HWF 14.7.2.1.
Boyden, C. *Jnl R. Met. Soc.*, 1937, p.383.

1.2.3. *Minimum temperature* (McKenzie)

If T_{max} = maximum screen temperature (°C)

T_d = air-mass dew-point (°C), which is assumed to remain constant or almost so (see note (*a*))

then $T_{min} = \frac{1}{2}(T_{max} + T_d) - K$.

TABLE 1.VI. *Values of K at Dyce*

Average surface wind speed overnight	Average amount of low cloud overnight (oktas)				
	0	2	4	6	8
knots	*degrees Celsius*				
0	8.5	6	6	4.5	4.5
1–3	7	5.5	5	4	3
4–6	4.5	4	3.5	2.5	1.5
7–10	4	3.5	3	2	1.5
11–16	3.5	3	3	1.5	1.0
17–21	2	2	2	1.5	1.0
22–27	2	2	1.5	1.5	1.0

Notes

(*a*) McKenzie's results for Dyce were derived from data limited to occasions when the dew-point was constant to within one or two degrees Celsius from 1300 GMT on one day to 0700 GMT on the next.

(*b*) Local values for *K* are known for 8 stations in Great Britain. The average value (from all the stations) of *K* for any given condition of wind speed and cloud cover is roughly the same as the appropriate value for Dyce. Consequently Table 1.VI. might give fairly good forecasts of the most representative minimum temperature over an area, though accuracy in forecasting for a particular place should be improved by deriving local values of *K*.

(*c*) See cautionary note (*c*) in 1.2.1.

Ref: HWF 14.7.2.1.
McKenzie, SDTM No. 68.

1.2.4. *Night cooling under clear skies* (Saunders)

(i) Consider whether the observed T_{max} and T_d (the dew-point at the time of T_{max}) for the afternoon are representative of the air likely to be over the locality during the night; if they are not, make an estimate of appropriate values from observations upwind.

(ii) Estimate whether there will be an inversion base at or below 900 mb during the afternoon.

(iii) Calculate T_r, the temperature at the time of discontinuity in the rate of cooling, from the regression equation

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

At Northolt, $K = 0.3$ degC when there is no inversion with base below 900 mb and $K = 2.2$ degC with an inversion with base below 900 mb.

(iv) The approximate times (GMT) at which the discontinuity in cooling occurs at Northolt are:

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
1645	1800	1930	2045	2100	2115	2115	2045	1930	1745	1700	1630

If the topsoil is wet the discontinuity occurs about 1 h earlier than given above in late spring and early summer, and about ½ h earlier in late summer.

(v) Obtain T_{min} from the forecast T_r and the forecast mean geostrophic wind speed for the period of subsequent cooling using Table 1.VII. and Figs. 1.4.–1.7.

TABLE 1.VII. *Selection of method for obtaining T_{min}*

Season	Forecast mean geostrophic wind	Method for T_{min}
	kt	
Summer (late March–early October)	0–12	Fig. 1.4. curve A } From 5 May to 10 Aug. add to T_{min} a correction from Fig. 1.7., except after rain.
	13–18	Fig. 1.4. curve B } From 5 May to 10 Aug. add to T_{min} a correction from Fig. 1.7., except after rain.
	over 18	By $T_{min} = T_r$ (calculated) – Δ where Δ (°C) is given by Fig. 1.6.
Winter (early October–late March)	0–16	Fig. 1.5., curve C or D according to afternoon lapse rate.
	17–21	Fig. 1.5., curve E.
	over 21	As in summer for winds over 18 kt.

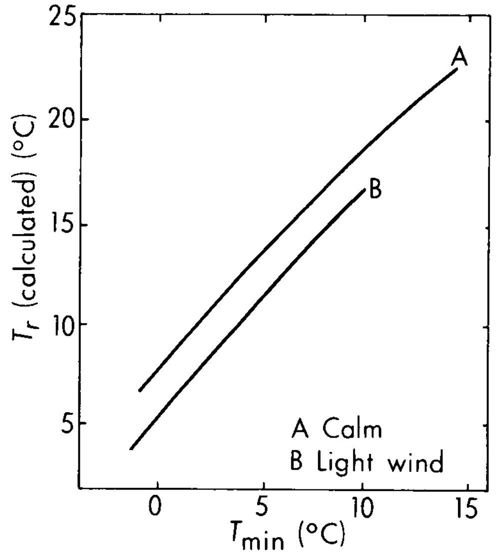


FIG. 1.4. *Relation between T_r and T_{min} in spring and late summer*

Calm = mean geostrophic wind 0–12 kt
Light wind = mean geostrophic wind 13–18 kt

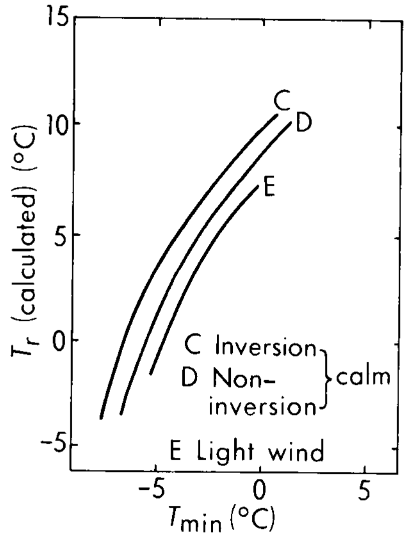


FIG. 1.5. *Relation between T_r and T_{min} in winter*

Calm = mean geostrophic wind 0–16 kt
Light wind = mean geostrophic wind 17–21 kt

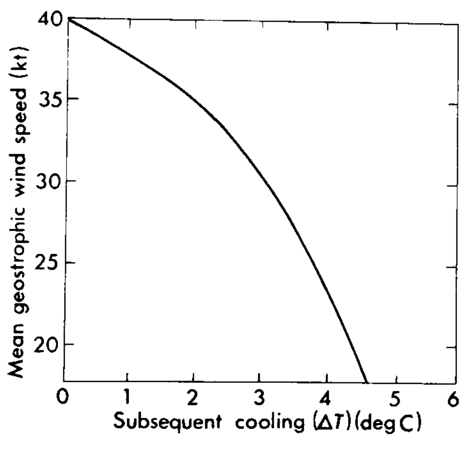


FIG. 1.6. *Relation between subsequent cooling and mean geostrophic wind speed for occasions of stronger wind*

Geostrophic wind $\geqslant$ 19 kt in summer and $\geqslant$ 22 kt in winter

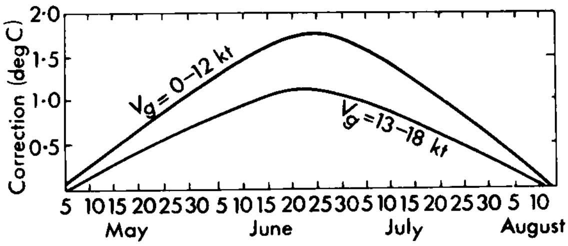


FIG. 1.7. *Correction added to T_{min} when Fig. 1.4. is used during midsummer period*

V_g = mean geostrophic wind during period

Note: if T_r is reached early because top soil is moist, this correction is partially offset.

Ref: Saunders, *Q.Jnl R.Met.Soc.*, 1949, p.154.
Q.Jnl R.Met.Soc., 1952, p.603.

Notes

(a) The above applies to Northolt with a clay soil and Saunders considers that the time at which T_r is reached may be applicable to clay soils elsewhere. Regarding values for T_{min} at other localities on other types of soil, separate corrections must be deduced from a sufficient number of cases for each section of Table 1.VII.

(b) See cautionary note (c) in 1.2.1.

(c) The account of the original work is summarized in HWF 14.7.2.2. The technique has been adapted for many places with different site characteristics.

Ref: Parry, *Met.Mag.*, Dec.'53.
Saunders, *Met.Mag.*, Jan.'54.
Roberts, *Met.Mag.*, Feb.'55.
Bruce, *Met.Mag.*, Apr.'55.
Richardson, *Met.Mag.*, Oct.'55.
Met.Mag., July '56.

For a coastal station for nights with winds off the sea Saunders has suggested that the parameter T_{max} should be replaced by T_s (the temperature of the sea surface) in the formula for forecasting T_r.

Ref: Saunders, *Met.Mag.*, Mar. '55.

1.2.5. *Night cooling under clear skies* (Barthram)

This method is applicable to inland stations in the southern half of England (based on Saunders's technique).

(i) If T_{max} is day maximum temperature, T_d is the dew-point at time of T_{max}, then T_r the temperature at the time when there is a check in the rate of cooling, is given by

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

where K = 1 degC when there is no inversion in the afternoon below 850 mb,

K = 2 degC when there is an inversion in the afternoon below 850 mb.

(ii) The times (GMT) of T_r at the *beginning* of each month are:

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
1645	1745	1845	2015	2100	2130	2145	2130	2030	1900	1745	1700

(iii) From the value of T_r and the forecast geostrophic wind speed during the night, Fig. 1.8. gives an estimate of the night minimum temperature (T_{min}).

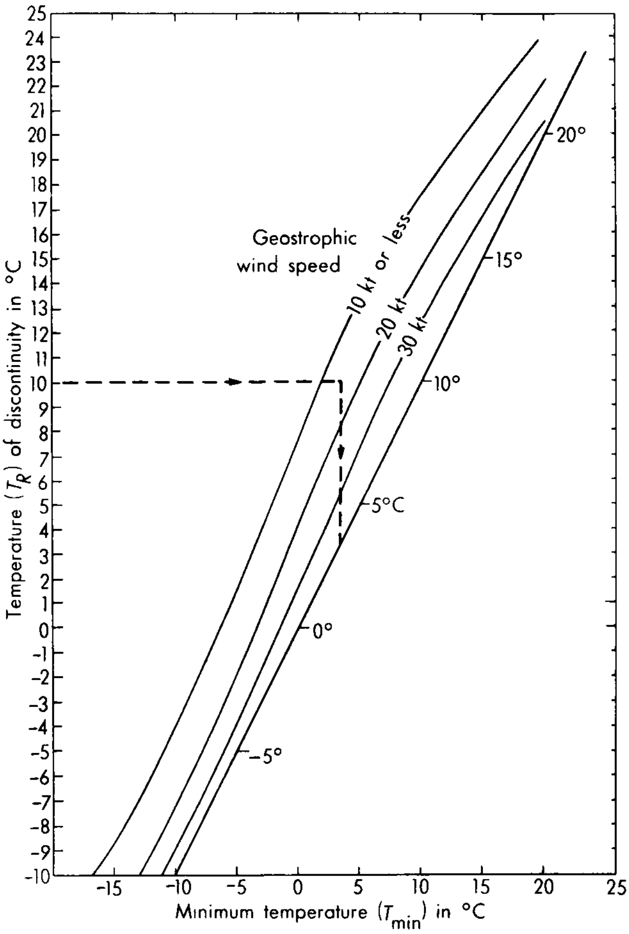


FIG. 1.8. *Graph for obtaining T_{min} from T_r*

(iv) A graph can be prepared of temperature against time by plotting T_{max}, T_r and T_{min} (assuming this value is realized at the time of local sunrise) and joining the points freehand to form a cooling curve with the point of inflexion at T_r.

(v) See note *(c)* below for case of falling temperature after fog has formed.

Notes

(a) Individual stations may have to recalculate *K* and redraw Fig. 1.8. T_r is related to time of sunset and varies with location and possibly with state of ground.

(b) The drawing of the cooling curve gives an early assessment of the time the fog-point is likely to be reached. Also hourly plots of actual temperature alongside the forecast profile draw attention to the

occasions when this forecast time may require amendment. The probable time of air frost is also readily obtained.

(c) To allow for the possibility of a continuing fall of temperature after fog has formed, use Fig. 1.9. to obtain the amended cooling curve after the expected fog-point is reached. This feature is important when, although the fog-point is significantly above 0°C, an air frost must still be considered. (Note, however, that Fig. 1.9. is based on an investigation at Wyton, Huntingdonshire, and therefore may have to be reconstructed for other stations.)

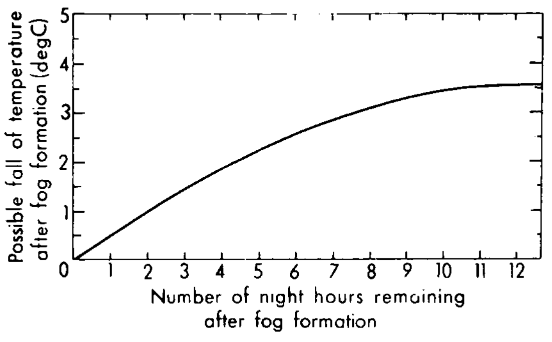


FIG. 1.9. *Possible fall of temperature after formation of radiation fog*

(d) A year's check at Laarbruch showed that on about 7/8 of occasions the method was useful, the forecast being correct to within ± 2 degC; half of these 'useful' occasions were within ± 1 degC. The most likely cause of failure is the difficulty of assessing a representative dew-point.

(e) See cautionary note *(c)* in 1.2.1.

Ref: Barthram, *Met.Mag.*, Aug. '64.

1.2.6. *Reduction of night cooling by presence of cloud* (Summersby)

If T_{min} = minimum temperature (°C) on night with cloud,

$T_{min(c)}$ = minimum temperature (°C) calculated by method of 1.2.4.

then $T_{min} = T_{min(c)} + K$.

TABLE 1.VIII. *Values of K*

N (oktas)	0	1	2	3	4	5	6	7	8
K (degC)	0	0	0.5	1	2	2.5	3.5	5	8

Notes

(*a*) It is suggested that the above formula, based on data for Northolt, would probably be satisfactory at stations with similar sub-soil and topography.

(*b*) Nights with precipitation are excluded.

Ref: HWF 14.7.2.3.
Summersby, *Met.Mag.*, July '53.

1.2.7. *Reduction of night cooling by presence of cloud* (Mizon)

When cloud is present the cooling at night is reduced to a proportion k (as given in Fig. 1.10.) of the cooling on a night which is clear of cloud but similar in other respects (e.g. initial temperature and humidity).

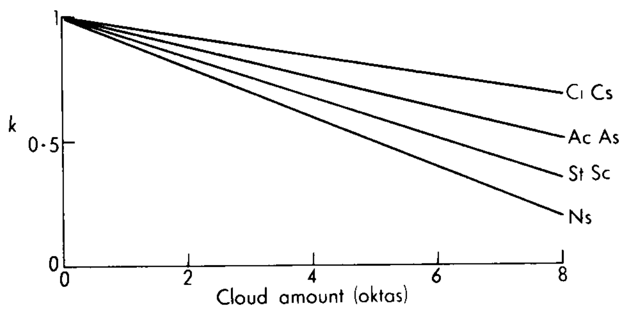


FIG. 1.10. *Values of k for various cloud types and amounts*

Ref: Mizon, *Aust. Met. Mag.*, Sept.'62.

1.2.8. *Grass minimum temperature on nights without fog* (Craddock and Pritchard)

Grass minimum temperature $T_g = T_{\min} - K$.

$T_{\min}$ is calculated by the method of 1.2.1.

TABLE 1.IX. *Values of K*

$\bar{V}_g$	$\bar{N}$ (oktas)			
	0–2	2–4	4–6	6–8
knots	*degrees Celsius*			
0–12	4	5	4	4
13–25	4	4	3	2
26–38	3.5	3	2.5	2.5
39–52	2.5	2.5	2.5	3

$\bar{V}_g$ = mean of geostrophic wind speeds for 1800, 0000 and 0600 GMT.

$\bar{N}$ = mean of total cloud amounts for 1800, 0000 and 0600 GMT.

Note

Craddock and Pritchard's formula is based on data for 16 stations in England.

Ref: HWF 14.7.3.
Craddock and Pritchard, MRP No.624.

1.2.9. *Grass minimum temperature on clear nights without fog* (Saunders)

$$T_g = T_{min} - K.$$

TABLE 1.X. *Values of K for Northolt*

Summer		Winter	
$\bar{V}_g$	K	$\bar{V}_g$	K
kt	*degC*	*kt*	*degC*
0–12	4.5	0–16	see below
13–24	6	17–24	6
over 24	4	over 24	4

Winter with $\bar{V}_g < 17$ *kt*

T_{min} (°C)	+2 to −0.5	−1 to −3.5	−4 to −6	−6.5 to −9
K (degC)	2	4	5	6

$\bar{V}_g$ = mean geostrophic wind speed overnight.

Ref: HWF 14.7.3.
Saunders, *Q.Jnl R.Met.Soc.*, 1952, p.603.

1.2.10. *Occurrence of grass minimum temperatures below freezing point* (Faust)

If the cloud amount is less than 2 oktas and the wind speed is less than 4 kt during the night, ground frost will occur if

$$(T + \tfrac{1}{2}T_d) < 17.2°\text{C},$$

where T is the screen temperature and T_d is the dew-point at 1400 local time.

Note

Occasionally ground frost has been experienced with $(T + \tfrac{1}{2}T_d)$

as high as 19°C. Therefore discretion should be used when $(T + \frac{1}{2}T_d)$ is a little above the critical value.

Ref: HWF 14.7.3.
Faust, *Ann. Met., Hamburg*, 1949, p. 105.
Jefferson, *Met.Mag.*, Oct. '51.
James, *Met.Mag.*, Mar. '53.

1.2.11. *Variation of minimum temperatures over short turf and bare soil* (Gloyne)

TABLE 1.XI. *Results at Starcross, Devon*, 1949-50

		Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
		degrees Celsius											
$T_{min} - T_b$	Mean	1.5	2	1.5	2	1.5	1	1.5	2	1.5	2	2.5	2.5
	Mean(R)	3	3.5	2.5	2.5	2.5	2	2	2.5	2.5	2.5	3.5	3.5
$T_b - T_g$	Mean	0.5	1	1	1.5	1.5	2	1.5	1.5	1.5	1.5	1	1
	Mean(R)	1.5	1.5	2	2.5	2.5	2.5	2.5	2	2	2.5	1.5	1.5

T_{min} = minimum temperature in screen

T_b = minimum temperature at ¼ to ½ inch above bare soil

T_g = grass minimum temperature

Mean(R) = mean on radiation nights, defined as nights on which $T_{min} - T_g \geqslant 4$ degC.

Ref: Gloyne, *Met.Mag.*, Sept. '53.

1.2.12. *Road minimum temperatures below freezing-point*

(i) Differences in the thermal properties of various road materials are generally quite small. Recent work has shown that the difference (minimum screen temperature–minimum road temperature) varies with the length of night and can be used to forecast the occurrence of road surface temperatures below freezing-point. The following regression equation was obtained for a concrete road at Watnall

$$M_A - M_R = 0.28t - 2.9,$$

where M_A = minimum screen temperature (°C)

M_R = minimum road surface temperature (°C)

t = time between sunset and sunrise (h).

The mean depression ($M_A - M_R$) for the middle of each month is given in the following table. If the forecast value of M_A is below the value given for the time of year, the road surface temperature is likely to fall below freezing-point.

Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Apr.
			degrees Celsius			
0.5	1.5	2	2	1.5	1	0

The standard deviation of the individual values about the mean was about 0.9 degC.

Similar results were obtained by Ritchie at Wyton.

Ref: Hay, *Met.Mag.*, Feb.'69.
Parrey, *Met.Mag.*, Sept.'69.
Ritchie, *Met.Mag.*, Oct.'69.

(ii) An alternative approach was tried by Clark. Using five years' data for a site at Newark, he found that the surface temperature of a concrete slab was likely to fall below freezing-point if the 1800 GMT concrete surface temperature was below the value given by the appropriate entry in the following table:

	Nov.	Dec.	Jan.	Feb.	Mar.
			degrees Celsius		
Clear nights (0-2 oktas)	5.5	4	3	5	8.5
Part cloudy nights (3-6 oktas)	4	3	2.5	4	6.5
Cloudy nights (7-8 oktas)	2	1	1	2	4

Ref: Clark, FTM No. 17, 1969

1.3. DAILY MEAN SURFACE TEMPERATURE (Boyden)

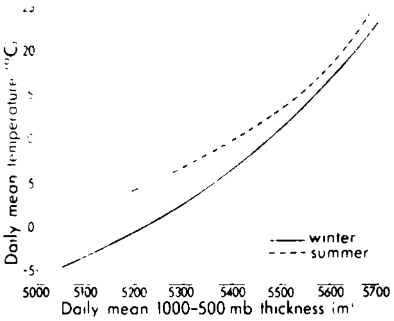


FIG. 1.11. *Relationship between daily mean surface temperature at London/Heathrow Airport and daily mean 1000–500 mb thickness at Crawley, 1956*

Winter = September to March   Summer = April to August

The daily mean temperature is the mean of the screen minimum temperature read at 0900 GMT and the maximum read at 2100 GMT

Ref: Boyden, *Met.Mag.*, Apr. '58.

1.4. MODIFICATION OF SURFACE AIR TEMPERATURE OVER THE SEA

1.4.1. *Advection of cold air over warm sea* (Frost)

If T_o (°C) and r_o (g/kg) are the temperature and humidity mixing ratio of the air *before* crossing the sea; T and r are the temperature and humidity mixing ratio of the air *after* crossing at least 60 miles of sea; and T_s and r_s are the *sea* temperature and humidity mixing ratio of *saturated* air at temperature T_s, then

$$T = T_o + 0.6\ (T_s - T_o),$$
$$r = r_o + 0.6\ (r_s - r_o).$$

Notes

(a) These formulae apply to all cold airstreams crossing a warmer sea surface, e.g. a cold N'ly outburst reaching N Scotland or a cold W'ly current reaching Norway.

(b) In applying the formulae, make the best estimate of the sea

surface temperature along the trajectory. Determine r_o and r_s from the tephigram.

Ref: HWF 14.10.1.
Frost, SDTM No. 19.

1.4.2. *Advection of warm air over cold sea* (Lamb and Frost)

For surface winds less than about 20 kt

$$T - T_s = (T_o - T_s)\, f(d)$$

$$\text{and } r - r_s = (r_o - r_s)\, f(d)$$

where d is the fetch (km) over the sea and other symbols are as in 1.4.1.

TABLE 1.XII. *Values of* f(d)

d (km)	100	200	300	400	500	600	700	800	900	1000
f(d)	0.175	0.152	0.141	0.133	0.127	0.123	0.119	0.116	0.113	0.110

Note

A cool sea surface exerts a powerful and rapid control on temperature at screen height.

Ref: HWF 14.10.2.
Lamb and Frost, M.O. 504.

1.5. COOLING OF AIR BY PRECIPITATION

1.5.1. *Cooling of air by rain*

The lowest temperature to which the air can be cooled by evaporation of precipitation is the wet-bulb temperature. A temperature close to the wet-bulb value is reached after about ½h of very heavy rain or about 1–2 h of rain of lesser intensity.

Ref: HWF 14.9.3.

1.5.2. *Downdraught temperatures in non-frontal thunderstorms* (Fawbush and Miller)

The temperature of strong downdraughts reaching the ground is very close to the surface temperature of the saturation adiabatic through the intersection of the wet-bulb curve and the 0°C isotherm (see 4.7.)

Ref: Fawbush and Miller, *Bull.Am.Met.Soc.*, 1954, p. 14.

1.5.3. *Cooling of air by snow* (Lumb)

Reduction of the surface temperature to 0°C as a result of downward penetration of snow is unlikely

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

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

Note

The relation between wet-bulb temperatures and the form of precipitation is given in 6.5.4. and 6.5.5.

Ref: Lumb, *Met.Mag.*, Jan '63.

CHAPTER 2. FOG

2.1. FOG FORMATION

2.1.1. *Fog-point* (Briggs)

Let T_f = the screen temperature at which fog is likely to become fairly general in an area which is relatively free from smoke pollution (this definition excludes the very localized fog which forms in valleys and sheltered places on almost any radiation night.)

θ_{dppp} = 'potential' dew-point at pressure level ppp

= temperature at intersection of the surface isobars with the constant mixing ratio line through the dew-point at ppp

and Γ = mean temperature lapse rate below ppp.

then (i) If $\Gamma > 2.2$ degC per 1000 ft, $T_f = \theta_{d850}$

(ii) If a more stable layer is reached at $ppp > 850$ mb, $T_f = \theta_{dppp}$.

(iii) If the air is stable from the surface upwards, T_f = surface dew-point.

(iv) If 2.2 degC per 1000 ft $> \Gamma >$ 1.7 degC per 1000 ft, $T_f = \theta_{d900}$.

Notes

(*a*) An exception to Briggs's rules is that if the water content increases with height from the ground, T_f = surface dew-point.

(*b*) Fog-points calculated by this and other methods are subject to errors in the following circumstances:

(i) If rain or showers occur in the late afternoon or evening (the calculated value is usually too low).

(ii) If the calculated T_f is near to or less than 0°C (T_f is usually too high).

Ref: HWF 17.7.2.3.
Briggs, *Met.Mag.*, Dec.'50.

2.1.2. *Fog-point* (Saunders)

The fog-point T_f in Fig.2.1. is the screen temperature at which the general visibility falls within the fog range with relative humidity 95% or more, or at which, with visibility already in the fog range, relative humidity rises to 95% or more.

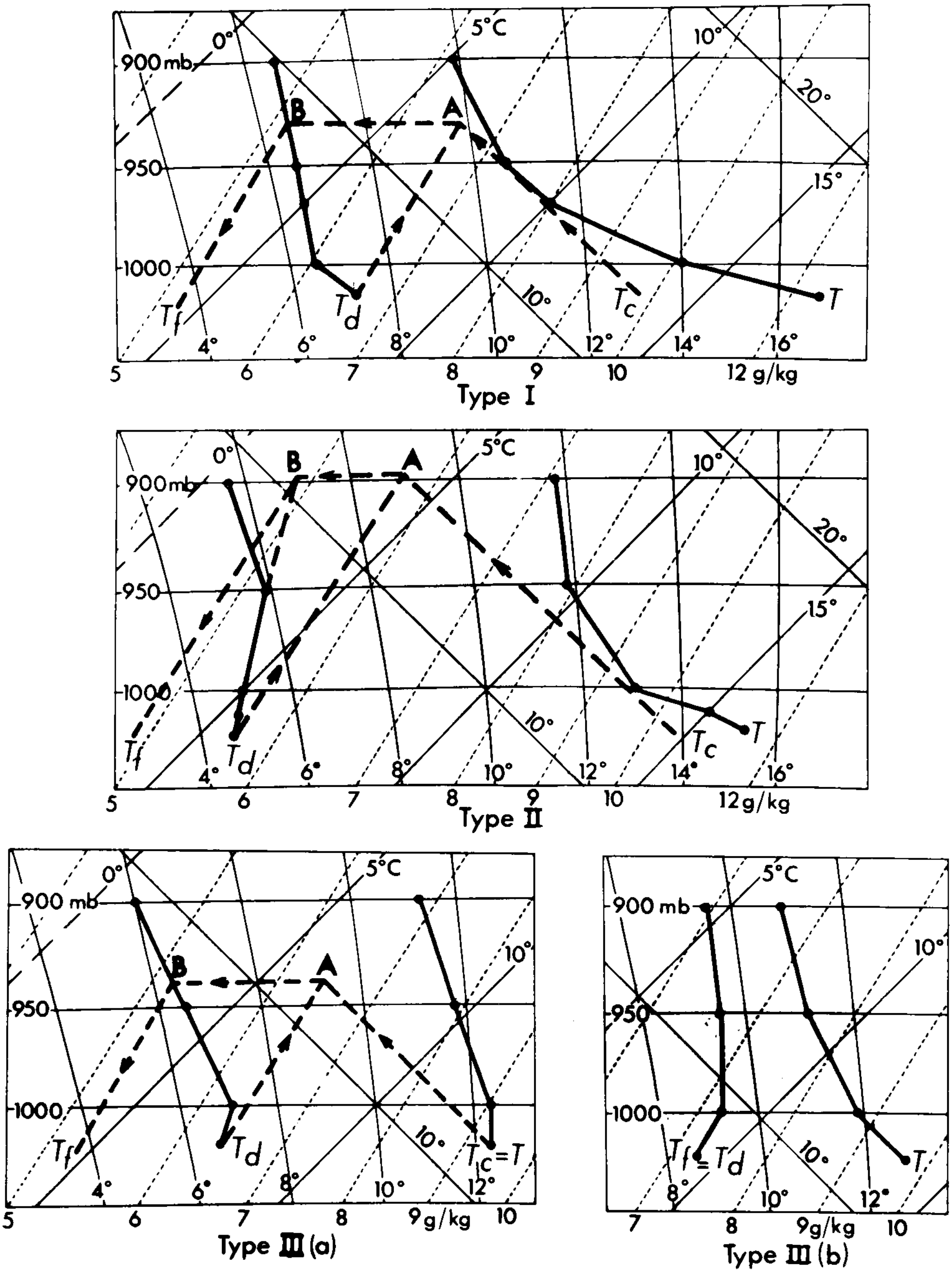


FIG.2.1. *Construction on a tephigram to obtain the fog-point for various types of dry-bulb temperature and dew-point distributions at 1200 GMT with surface values modified to 1500 GMT*

T_f = fog-point T = surface dry bulb at 1500 GMT

T_d = surface dew-point at 1500 GMT

T_c = corrected surface dry bulb (by use of this value super-adiabatic lapse rates in the lowest layers are in effect ignored)

(i) Select the 1200 GMT tephigram most representative of the air which will be over the area during the night and redraw the temperature and dew-point curves to allow for changes taking place near the surface up to 1500 GMT (see Note (*a*)).

(ii) The heavier pecked lines of Fig.2.1. indicate constructions on the tephigram to determine T_f. These depend on which of the following types of dew-point lapse is appropriate.

Type I – Constant dew-point lapse aloft; surface dew-point on, or to the right of, downward extension of upper dew-point curve.

Type II – Dew-point lapse aloft increasing at high levels within the layer (up to level AB); surface dew-point as in Type I.

Type III – Surface dew-point to the left of downward extension of upper dew-point curve;

(*a*) temperature lapse in lowest layer less than dry adiabatic,

(*b*) temperature lapse in lowest layer equal to, or greater than, dry adiabatic.

Notes

(a) Strictly the technique requires a tephigram for 1500 GMT, the time for routine radiosonde ascents when the technique was published.

(b) In extreme cases where a subsidence inversion has brought dry air down to only 20 or 30 mb above the ground, it is usually better to use the surface dew-point as the fog-point.

(c) See cautionary note *(b)* in 2.1.1.

Ref: HWF 17.7.2.2.
Saunders, *Met.Mag.*, Aug. '50.
Saunders and Summersby, *Met.Mag.*, Sept. '51.
Saunders, *Met.Mag.*, Mar. '58.

2.1.3. *Fog-point* (Craddock and Pritchard)

If T_f = the screen temperature at which fog is likely to form (see Note (*b*)).

T_{12} = screen temperature at 1200 GMT

and T_{d12} = dew-point at 1200 GMT,

then $T_f = 0.044\,T_{12} + 0.844\,T_{d12} - 0.55 + A = Y + A$.

Values of Y can be found from Fig.2.2. and A from Table 2.I.

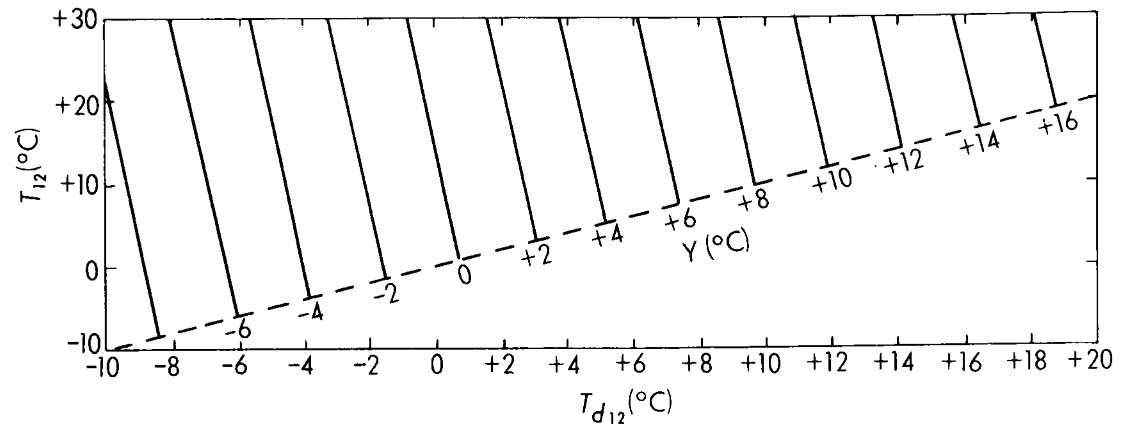


FIG.2.2. *Nomogram for determining Y from T_{12} and T_{d12}*

TABLE 2.I. *Values of A*

Mean* cloud amount	Mean* geostrophic wind speed (kt)	
	0 – 12	13 – 25
oktas	*degrees Celsius*	
0 – 2	0	–1.5
2 – 4	0	0
4 – 6	+1	+0.5
6 – 8	+1.5	+0.5

*Mean of forecast values for 1800, 0000 and 0600 GMT.

Notes

(a) The equation for T_f was derived from the combined data for 13 widely separated stations in England. There was considerable variation from station to station in the proximity to major smoke sources.

(b) Account was not taken of variations in atmospheric pollution so that, in effect, the average degree of pollution is assumed in using this technique (in contrast to the techniques given in 2.1.1. and 2.1.2. which refer mainly to fog in clean air).

(c) The authors have suggested that in forecasting for a particular station the difference between the forecast T_{min} (by the method of 1.2.1) and the forecast T_f (by the method given above) might be interpreted as follows:

$T_f - T_{min}$	Forecast
+ 1 degC or higher	Fog
+ 0.5 degC to −1.5 degC	More or less serious risk of fog
−2 degC or lower	Negligible risk of fog

Ref: HWF 17.7.2.1.
Craddock and Pritchard, MRP 624.

2.1.4. *Forecasting fog* (Swinbank)

(i) Applicable to England, SE of a line Wash-Birmingham-Southampton.

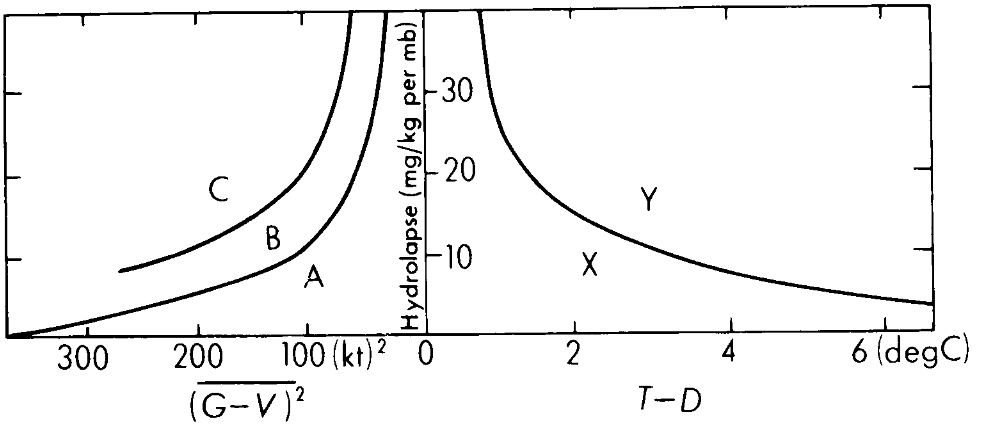


FIG.2.3. *Fog prediction diagram for October and November over England SE of a line Wash-Birmingham-Southampton*

x = humidity mixing ratio (g/kg)
p = pressure (mb)
$\Delta x/\Delta p$ = afternoon hydrolapse to inversion (or to 800 mb if no inversion lower) (milligrams/kg per mb)
$T-D$ = mean depression of dew-point below dry bulb at 1800 GMT (degC)
G = forecast gradient wind speed for 1800-0700 GMT (kt)
V = forecast wind speed at anemometer height for 1800-0700 GMT (kt)
$\overline{(G-V)^2}$ = mean values at various points in the area (kt)

If a point falls in the zones:

C and Y, fog will not develop

B and Y, fog is improbable, even locally

A and Y / C and X — fog may or may not develop; if it does it will be patchy

B and X, as above but higher probability of local fog

A and X, fog will develop; 50% chance of being widespread.

(ii) Applicable to Lincolnshire and Yorkshire.

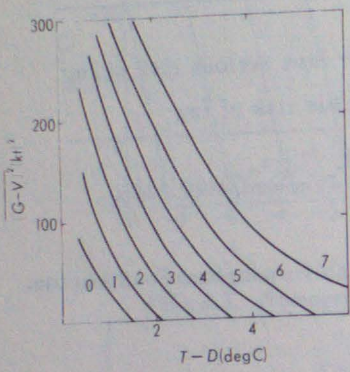


FIG.2.4.*(a) Incidence of fog in Lincolnshire and Yorkshire*

Zones 0, 1, 2 – widespread fog
3-6 – patchy fog
7 – no fog

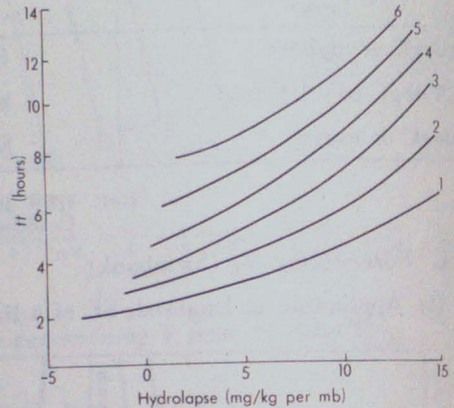


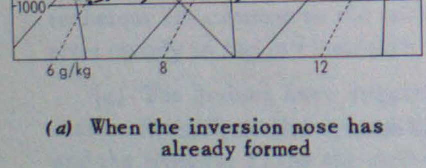
FIG.2.4.*(b) Time of formation of fog in relation to hydrolapse and zone of incidence as given by Fig.2.4.(a)*

tt = number of hours after sunset by which widespread fog will have formed at half the stations, or the average number of hours after sunset of formation of patchy fog

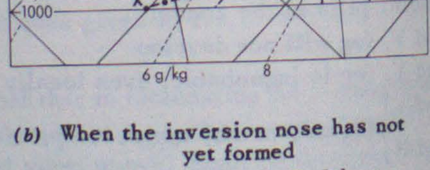
Ref: HWF 17.7.2.4.
Swinbank, *Prof. Notes* No.100.

2.2. FOG TOP

2.2.1. *Estimation of fog top at dawn from midnight Balthum* (Heffer)



(a) When the inversion nose has already formed



(b) When the inversion nose has not yet formed

FIG.2.5. *Construction for fog-top from a midnight Balthum*

——— Dry bulb — — — — — Dew point
–·–·–·–·–·–·– Construction for fog-top estimate
O Estimated fog top X Night minimum temperature

(i) If the nose has already formed on the temperature curve, the level is raised by 5 mb and the temperature is decreased by 1.5 degC and this point is joined to the night minimum temperature by a straight line on the tephigram. The point where this line and the dew-point curve intersect (O in Fig.2.5.*(a)*) represents the fog top at dawn.

(ii) If a nose has not yet formed on the temperature curve at midnight, the point 35 mb above the ground is joined to the night minimum surface temperature and the fog-point at dawn estimated as before (O in Fig.2.5.*(b)*).

Ref: Heffer, *Met.Mag.*, Sept.'65.

2.3. FOG CLEARANCE BY INSOLATION

2.3.1. *Fog clearance by insolation* (Barthram – based on Kennington)

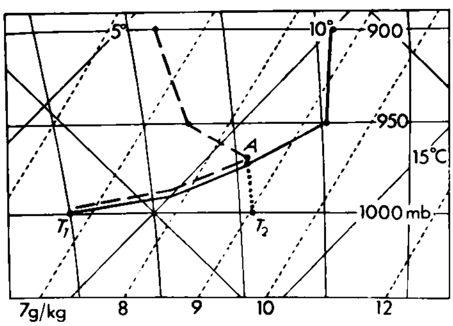


FIG.2.6. *Ascent curve on night of fog*

——— dry bulb – – – – dew-point

T_1 – dawn temperature T_2 – fog clearance temperature A – fog top

(i) The fog depth (mb) must be known or estimated. A midnight radiosonde will give a useful guide but it must be remembered that the depth of fog may have increased, say by 5 to 15 mb, by dawn (see also 2.2.1.).

Barthram suggests that the clearance temperature, T_2, may be estimated by assuming a saturated-adiabatic lapse rate from the fog top to the surface. The value of the dawn temperature, T_1, must be known.

(ii) Estimate the time of fog clearance by means of the nomogram in Fig.2.7. In part 3 of the nomogram the scale for thin fogs is used if the visibility is 600 metres or more.

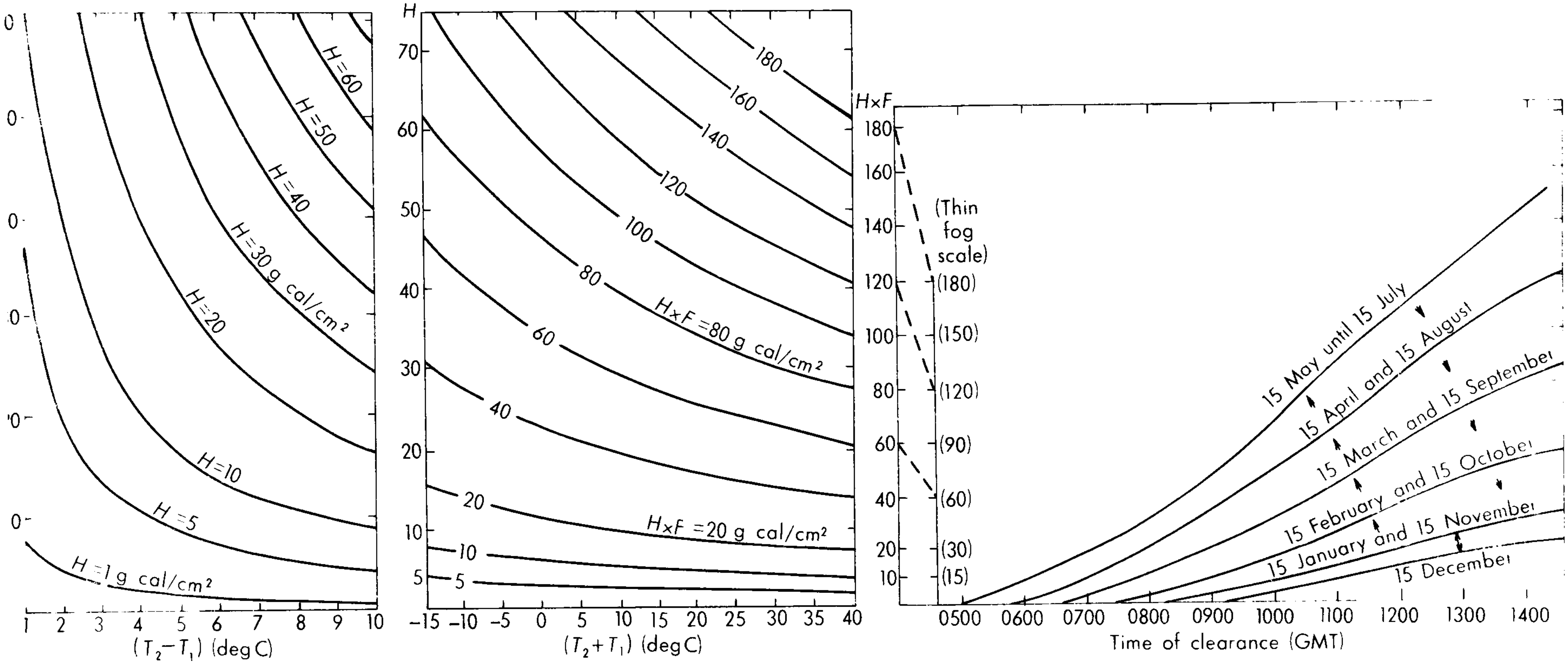


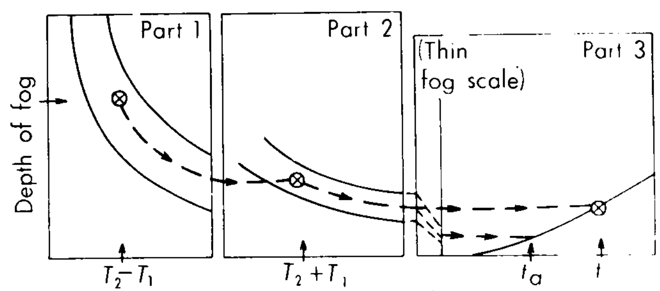
FIG. 2.7. *Nomogram for assessing time of fog clearance by insolation—southern England*

T_1 = temperature at dawn (°C) T_2 = fog clearance temperature (°C)

H = insolation required to warm layer of dry air (g cal/cm²)

$H \times F$ = total insolation required to warm layer of foggy air (g cal/cm²)

t = time of fog clearance t_a = time of clearance of thin fog



METHOD OF USE

Notes

(*a*) The assumed values of available heat are those appropriate to southern England. For northern parts of the British Isles the corresponding values are less and consequently fog is likely to clear later than would be estimated from Fig.2.7. – especially during winter.

(*b*) Heffer has tested this technique for 8 stations in East Anglia and the E Midlands.

Ref: Kennington, *Met.Mag.,* Mar. '61.
Barthram, *Met.Mag.,* Feb.'64.
Heffer, *Met.Mag.,* Sept.'65.

2.3.2. *Fog clearance by insolation* (Jefferson)

(i) Estimate the fog top either from examination of a representative tephigram or from any available reports (allowing for possible upward extension of fog due to increased turbulence around dawn).

(ii) The clearance temperature, T_c, is estimated in Jefferson's method on the assumption of a dry adiabatic from surface to fog top.

(iii) Estimate the temperature rise by the method given in 1.2.

Ref: HWF 17.8.
Jefferson, *Met.Mag.,* Apr.'50.

2.4. FOG CLEARANCE FOLLOWING ARRIVAL OF CLOUD

2.4.1. *Fog clearance following arrival of cloud* (Saunders)

The arrival of a cloud sheet over water fog often leads to clearance of the fog. Clearance is especially likely if the cloud base is low and if the soil temperature is well in excess of the screen temperature. The time taken for the fog to clear is greatest when temperatures near the surface are low (see Table 2.II.).

TABLE 2.II. *Variation with grass-level temperature of time taken for fog to clear at Exeter Airport following the arrival of cloud*

Initial grass temperature	Number of cases	Average time for fog to clear
°C		*hours*
Below 0	10	3.1
0–2	10	2.2
3–5	5	1.1
6–8	10	1.5
9–10	5	0.9
11–13	3	0.5

Ref: Saunders, *Met.Mag.,* Jan.'60.
Met.Mag., Mar. '60.

2.5. ADVECTION FOG

Conditions for formation are:

(i) Original dew-point of the air higher than the temperature of the underlying surface.

(ii) A stable lapse rate of temperature and only a slight hydrolapse.

(iii) A suitable wind in the first instance to transport the air from the warm surface to a colder surface.

2.5.1. *Sea fog*

(i) Two important cases of the formation of advection fog over the sea occur in airstreams approaching the British Isles from the SW and from the E (over the North Sea). An estimate of the cooling of air crossing the North Sea can be made by the method given in 1.4.2.

(ii) Sea fogs when surface winds are as strong as 25 kt are not uncommon.

(iii) Occasionally fog which has been lying over the sea may be brought across the coast – even at the time of maximum insolation – by a sea-breeze.

2.5.2. *Advection fog over the land*

(i) For advection fog over the land to be at all widespread or persistent the ground must be very cold – either frozen or snow covered.

(ii) Advection fog is unlikely over the land if the surface wind exceeds about 10 kt.

2.6. VISIBILITY IN FOG

2.6.1. *Visibility in sea fog*

(i) In the majority of fogs over the eastern Atlantic in the latitudes of the British Isles, visibility is likely to exceed 200 m. Table 2.III. gives some statistics for ocean weather stations J and I.

TABLE 2.III. *Visibility observations in sea fogs at stations J and I during the summers of 1957 and 1958*

Station	Visibility (m)				Total
	< 50	50-200	200-500	500-1000	
	number of occasions				
J (52° 30′N 20°W)	0	15	56	53	124
I (59°N 19°W)	9*	48	31	31	119

*All 9 observations occurred during one spell of fog at station I.

(ii) There is some evidence that visibilities are lower in sea fogs in more northern (and cooler) waters.

(iii) When sea fogs form as a moist land-breeze crosses the coast to an adjacent cold sea and there is marked rapid cooling below the dew-point (for example North Sea fogs in early summer), very dense fog may be quickly formed.

Ref: HWF 17.10.1.

2.6.2. *Visibility in radiation fog*

(i) In country districts rapid thickening of radiation fog after its formation is probably typical. For instance, at Exeter Airport and at Merryfield the deterioration of visibility from 1000 m to 200 m or less occurs within 1 h in about ½ of the fogs and within 2 h in about ¾ of the fogs.

(ii) In districts where there is heavy smoke pollution which is several hundred feet thick vertically, such sudden deteriorations may be less common. The increase in smoke concentration may reduce visibility to below 1000 m long before the temperature has reached the fog-point. There is also some evidence that the subsequent fall of visibility as condensation takes place is also more gradual than in 'clean' air.

Ref: HWF 17.10.2.
Saunders, *Met.Mag.*, Jan.'60.
Met.Mag., Dec.'57.

CHAPTER 3. CLOUD

3.1. STABILITY DEFINITIONS

Static stability (or *hydrostatic stability* or *vertical stability*) in a layer of air is normally assessed by its influence on a parcel of its air which is made to move vertically through the layer. The state of the layer of air is then defined as follows:

Stable:	the buoyancy force opposes the movement
Unstable:	the buoyancy force assists the movement
Neutral:	the buoyancy force is zero.

Latent instability. If the buoyancy force opposes upward motion initially but assists it when the parcel of air reaches a higher level, the whole layer of air is said to possess latent instability.

The buoyancy of a parcel of air depends on its temperature relative to the temperature of the air around it. The parcel of air, if it is cloudy, acquires a temperature in moving vertically which depends partly on release or absorption of latent heat. Thus the stability of a layer of air depends on its humidity. The following terms are used to describe the stability of a layer of air in relation to both temperature and humidity:

Absolute stability: stable regardless of its humidity ($\gamma < \Gamma_s$)

Absolute instability: unstable regardless of its humidity ($\gamma > \Gamma_d$, i.e. a superadiabatic lapse rate)

Conditional instability: stable if unsaturated, unstable if saturated ($\Gamma_d > \gamma > \Gamma_s$).

Potential instability (or convective instability). If a column of air is lifted bodily the temperature decrease varies from level to level, particularly because some layers become cloudy (and acquire the released latent heat) sooner than others. Thus certain layers become unstable. The layers which are thus potentially unstable are those in which the potential wet-bulb temperature decreases with height.

3.2. FORECASTING CONVECTIVE CLOUD, PARCEL AND SLICE METHODS

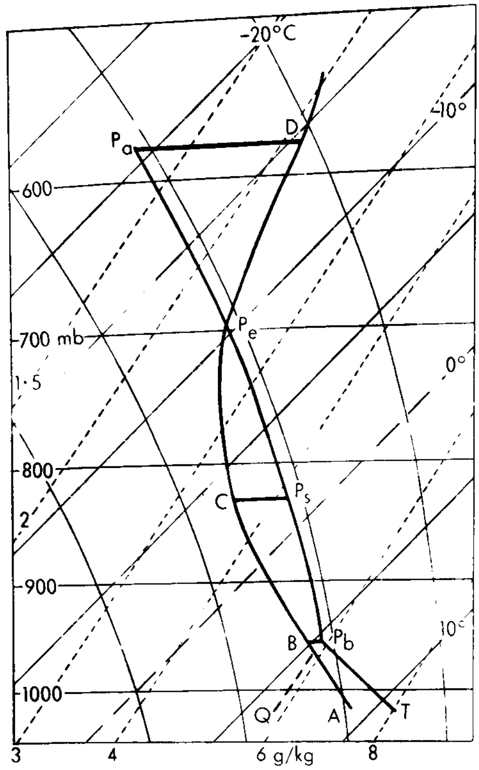


FIG. 3.1. *Constructions on tephigram for parcel and slice methods for forecasting tops of convective cloud*

ABCD dry-bulb temperature before convection begins
T = surface temperature expected as a result of day-time heating (see 1.1.1.)
Q = surface dew-point forecast for the appropriate time (see note (*a*))
$TP_bP_sP_eP_a$ = temperature changes in a parcel of air rising from T to P_a

(i) If P_b lies to the left of ABCD convective cloud does not form. Otherwise the height of the cloud base is indicated by the pressure P_b.

(ii) By the *parcel method* (see note (*b*)) the cloud tops are forecast to extend to P_e and possibly higher because the rising air overshoots the level at which it ceases to be buoyant; if the isobar P_aD is drawn to make areas P_aDP_e and $BCP_eP_sP_b$ equal then P_a indicates the absolute limit to development of tops of convective cloud.

(iii) By the *slice method* (see note (*c*)) cloud tops are forecast to rise to, or slightly above, the level where the environment lapse rate becomes equal to the SALR (the level indicated by P_s in Fig.3.1.).

Notes

(a) The surface dew-point is subject to diurnal variation. An estimate of the value to be expected during convection can be made from the preceding midnight ascent by extrapolating to the surface level the dew-points in the layer some 50-100 mb above the surface.

(b) The parcel method depends on the simplifying assumptions that:

(1) the rising parcels of air do not cause compensating downward motion, and therefore adiabatic warming of the environmental air, and
(2) mixing of air does not occur between the parcels and the environment.

(c) The slice method is an attempt to allow for the compensatory downward motion and warming of the environment. Some relationships between cloud levels observed and those forecast by the parcel and slice methods are given in 3.3. and 3.4.

Ref: HWF 3.6.1., 3.6.2., 16.6.5.
Petterssen *et alii, Geofys. Publr, Oslo,* 16, No.10, 1945.

3.3. BASE OF CONVECTIVE CLOUD (Petterssen *et alii*)

For a convective cloud which does not give precipitation the pressure at the base can be estimated as 25 mb less than the condensation pressure (P_b in Fig.3.1.) computed from the accompanying temperature and dew-point at the surface (i.e. the cloud base is estimated to be about 700 ft higher than the computed condensation level). On 75% of occasions the actual pressure at the cloud base is within 25 mb of this estimate.

Notes

(a) This relationship was derived from aircraft ascents made at Aldergrove and Bircham Newton (Norfolk) during the months April to September.

(b) In an investigation into the diurnal variation of small shower clouds over Bedfordshire during several days in August, Browne, Day and Ludlam found a similar relationship between the cloud base and computed condensation level up to the time of maximum temperature; afterwards the cloud base remained at approximately the same height or fell only a few hundred feet although the computed condensation

level fell rapidly until it was about 3000 ft below cloud base.

Ref: HWF 16.6.5.1.
Petterssen *et alii*, *Geofys. Publr, Oslo*, 16, No.10, 1945.
Browne, Day and Ludlam, *Met.Mag.*, Mar. '55.

3.4. TOPS OF CONVECTIVE CLOUD

Relationships found by various investigators between the forecast heights of convective cloud tops by the parcel or slice methods and observed heights are not altogether consistent. On the whole reports suggest that

(i) convective cloud tops mainly reach levels between P_s and P_e (Fig.3.1.), and

(ii) with large convective storms – particularly storms which are sustained by a favourable wind field on the synoptic scale and are extensive horizontally – some cloud tops may reach P_a (Fig.3.1.). Cloud tops associated with large storms have been reported to extend as high as 20 000 ft above the tropopause (see 6.4.8.).

Ref: HWF 16.6.5.2.
Petterssen *et alii*, *Geofys. Publr, Oslo*, 16, No.10, 1945.
Roach, *Aero. Res. Coun.* 28044, 1966.
Ludlam and Scorer, *Q.Jnl R. Met. Soc.*, 1953, p.317.

3.5. FORMATION OF STRATUS BY NOCTURNAL COOLING

3.5.1. *Temperature at which stratus forms* (Saunders)

(i) Obtain the fog-point, T_f, by the method of 2.1.1.

(ii) Assess the height of the top of surface turbulence at night (e.g. by the method of 3.5.2.).

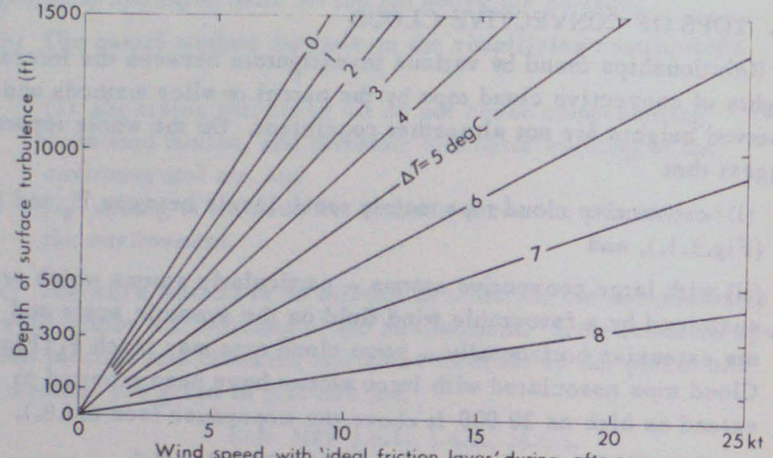
(iii) On a tephigram draw the constant mixing ratio line through T_f (plotted on the surface isobar).

(iv) From the intersection of this line with the isobar corresponding to the top of the surface turbulent layer draw a dry adiabatic. The intersection of the adiabatic with the surface isobar indicates the temperature at which stratus will form.

Ref: HWF 16.6.3.

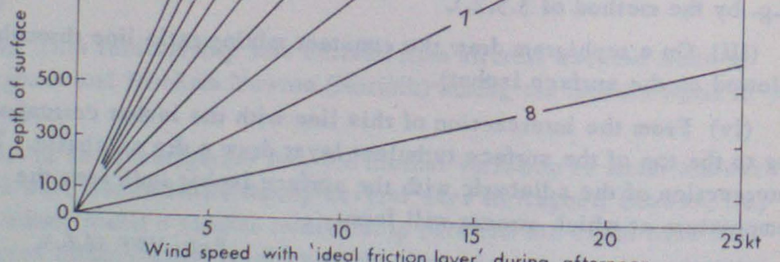
3.5.2. *Depth of surface turbulence at night* (Gifford)

This technique gives an estimate of the height of the top of surface turbulence following an afternoon when there was an 'ideal friction layer' – in effect a friction layer developed without any substantial check to surface heating. Adjustments to the technique are suggested in note *(b)* for occasions when an 'ideal friction layer' was not realized during the afternoon.



(a) For rather smooth surface

$z_0 = 0{\cdot}65$ cm



(b) For rather rough surface

$z_0 = 7$ cm

FIG. 3.2. *Nomogram for estimating the depth of surface turbulence at night*

ΔT = fall of surface temperature below the afternoon value on occasions with an 'ideal friction layer'

The nomograms given in Fig.3.2. are for a latitude of 53° (the variations of latitude within the U.K. make only a slight difference to the results) and for an anemometer height of 10 metres (the height above open level terrain for standard exposure).

An estimate has to be made of z_0, a parameter of surface roughness, which can vary from place to place and from sector to sector at a given place according to the average height of upwind obstructions. The values of z_0 for which the nomograms are given can be considered fairly extreme: 0.65 cm is the sort of value appropriate to flow off a rather smooth surface (e.g. the sea) and 7 cm to flow off a rather rough surface (e.g. a city).

Notes

(a) This technique has been shown to be suitable for several aerodromes in the U.S.A. The technique has not been tested in the U.K.

(b) After a day on which an 'ideal friction layer' is not realized (because of low cloud cover or precipitation preventing normal heating) the depth of turbulence at night is restricted and stratus can be expected to develop on average at only half the height shown by the appropriate nomogram.

(c) The technique implies that with a fall of surface temperature of about 9 degC or more from the afternoon value at the time of the 'ideal friction layer', stratus will form (humidity being favourable) at the surface regardless of the afternoon wind speed.

Ref: Gifford, *Bull. Am.Met. Soc.,* 1950, p.31.

3.6. SPREAD OF STRATUS FROM THE NORTH SEA ACROSS EAST ANGLIA (Sparks)

When winds are off the North Sea and stratus covers the sea but has dispersed over land on account of day-time heating, then the inland spread of stratus during the evening and night can be forecast as follows:

(i) In the absence of a suitable temperature sounding through the stratus, the temperature at which the stratus cleared in the morning provides the best estimate of the temperature at the coast when stratus will start to move inland.

(ii) The movement of stratus inland can be forecast from the directions and speeds of surface winds at a time when convection and turbulence are still operative in the lowest layers (say 1800 GMT in summer). This is likely to give a better forecast than one based on the pattern of isobars, particularly if there is sharp anticyclonic curvature.

(iii) Stratus may form over high ground before the arrival of the main cloud sheet forecast as in 3.5.1.

Ref: Sparks, *Met.Mag.*, Dec. '62.

3.7. DISPERSAL OF STRATUS BY INSOLATION

It is assumed that the cloud will clear when the surface temperature reaches such a value as to establish a dry-adiabatic lapse rate up to the level of the stratus top. The time taken to attain this temperature can be assessed by the method of 1.1.2.

Ref: HWF 16.6.3.2.

3.8. DISSIPATION OF STRATOCUMULUS BY CONVECTION (Kraus)

A cloud layer will not disperse by convective mixing with the air above if the pressure at the cloud top is less than P_c, as given below. (If the pressure at the top is greater than P_c the cloud may or may not disperse.)

$$P_c = P + a\,(P_0 - 1000)$$

where P_0 is the surface pressure (mb) and values of P and a are given in Table 3.I.

TABLE 3.I. *Values of P and a*

	Temperature at cloud top	Magnitude of inversion containing the cloud layer (degC) 10		8		6		4		2	
		P	*a*	*P*	*a*	*P*	*a*	*P*	*a*	*P*	*a*
	°C	*millibars*									
Water cloud	20	833	0.80	861	0.83	891	0.87	924	0.90	960	0.95
	10	803	0.75	834	0.79	869	0.82	906	0.87	951	0.93
	0	755	0.67	789	0.71	830	0.76	877	0.82	932	0.90
	−10	680	0.56	719	0.60	765	0.66	823	0.73	898	0.84
Ice cloud	0	779	0.71	812	0.75	850	0.79	891	0.85	941	0.91
	−10	702	0.59	739	0.63	786	0.69	839	0.76	908	0.85
	−20	586	0.45	628	0.49	679	0.54	747	0.62	841	0.74
	−30	451	0.30	489	0.34	540	0.38	613	0.45	728	0.58

Ref: Kraus, SDTM No. 67, 1944.

3.9. NOCTURNAL DISPERSAL OF STRATOCUMULUS OVER LAND (James)

The cloud will break if $D_m > D_c$ where

D_m = maximum depression (degC) of the dew-point below the dry-bulb temperature in the 50-mb layer *above* the cloud.

D_c is given in Table 3.II. below, where

h = difference (g/kg) between the humidity mixing ratios at the top and bottom of the 50-mb layer *below* the cloud

z = cloud thickness (mb).

TABLE 3.II. *Values of* D_c

	h (g/kg)					
z	0.25	0.5	0.75	1.0	1.25	1.5
mb	*degrees Celsius*					
10	–	–	1	3	6	8.5
20	0	2.5	5	8	10	13
30	4	7	9	12	14.5	17
40	9	11	14	16	19	21
50	13	15	18	20.5	23	26
60	17	20	22	25	27	30
70	21	24	26.5	29	32	34

Note: a linear hydrolapse in the layer is assumed.

The technique applies under the following conditions:

(i) The stratocumulus sheet is bounded at its top by a dry-type inversion, that is, a rapid decrease of humidity with height through the region of temperature increase.

(ii) There is no surface front within 400 miles of the locality of the cloud sheet.

(iii) The cloud base is above the condensation level of any convection from the sea (the rule applies only to stratocumulus over land).

(iv) The cloud sheet is extensive, covering several hundred square miles, and gives almost complete cloud cover, more than 6/8 for at least 2 consecutive hours. (The cloud was regarded as having dissipated if it broke to 2/8 or less for at least 2 consecutive hours.)

Failures of the technique in day-to-day forecasting can often be attributed to

(i) inaccurate assessment of the cloud thickness (in the absence of reports from aircraft), and

(ii) uncertainties as to the magnitude and steepness of the temperature inversion and hydrolapse, because of the lag of radiosonde elements.

Ref: HWF 16.6.3.4.
James, *Met.Mag.*, July '56.
Q.Jnl R. Met. Soc., 1959, p.120.

3.10. FORECASTING CIRRUS OVER THE BRITISH ISLES

3.10.1. *Forecasting cirrus* (James)

(i) *Occurrence 6-9 h ahead*

Four or more oktas of cirrus cloud are likely if the reply to *5 or more* of the following questions is affirmative (D_{ppp} = depression of dew-point below air temperature at pressure level *ppp*):

(1) Is $D_{500} \leqslant 10$ degC? (2) Is $D_{450} \leqslant 10$ degC?

(3) Is $D_{400} \leqslant 10$ degC?

(4) Is the lapse rate in the 500-300 mb layer greater than the saturated adiabatic?

(5) Is the 400-mb wind between SW and NW?

(6) Is there a veer of wind of 20° or more between 500 mb and 300 mb?

(7) Is the 1000-500 mb thermal wind > 20 kt?

(8) Is the forecast area in a ridge in the 1000-500 mb thickness pattern?

(9) Is there anticyclonic curvature in the 1000-500 mb thickness lines?

(10) Is there a deep cold pool, or intense thickness trough, in the 1000-500 mb thickness pattern?

(11) Is the forecast area in, or just to the rear of, a ridge in the 300-mb contour pattern?

(12) Is the forecast area on the anticyclonic side of a 300-mb jet stream and within 300 miles of the jet axis?

(13) Is the forecast area up to 300 miles ahead of a surface warm front or occlusion?

(ii) *Occurrence 24–36 h ahead*

Four or more oktas of cirrus cloud are likely to occur if the reply to *2 or more* of the following questions is affirmative:

(1) In the air which will be over the area during the forecast period, is the present dew-point depression at 500 mb or higher levels $\leqslant$ 10 degC (trace back the air using the 500-mb forecast chart)?

(2) Will the area be up to 300 miles ahead of a surface warm front or occlusion (use the surface forecast chart)?

(3) Will the area be on the anticyclonic side of a 300-mb jet stream and within 300 miles of the jet axis?

(4) Will the area be in, or just to the rear of, a 300-mb ridge?

(5) Will the area be in a thermal ridge (use 1000-500 mb pre-thickness chart)?

(6) Will the veer of wind between 500 mb and 300 mb be 20° or more?

Ref: HWF 16.6.4.1.
James, *Prof. Notes* No.123, 1957.

3.10.2. *Forecasting cirrus 6–12 h ahead* (Singleton and Wales-Smith)

The method given in 3.10.1. has been simplified and adapted to give some indication of the high-cloud cover to be expected. A score of 0, ½ or 1 is given for each of the following questions according to whether the answer is 'no', 'uncertain' or 'yes' respectively:

(1) Is the forecast area in, or just to the rear of, a ridge in the 200-mb contour pattern?

(2) Is the forecast area on the anticyclonic side of a 200-mb jet stream and within 300 miles of the jet axis?

(3) Is the forecast area up to 300 miles ahead of a surface warm front or occlusion?

The scores are added. Cirrus amounts are roughly associated as follows:

Total score	0	½	1	1½	2*	2½	3
Amount (oktas)	0	1	2	3	4	5	6-7

*For a score of two the scatter of cloud amount values is particularly large so use of the fuller method of 3.10.1. is recommended to decide whether or not the amount will be at least 4 oktas.

Note

This technique is not likely to be adequate for forecasting cirrus produced by convection.

Ref: Singleton and Wales-Smith, *Met.Mag.*, Apr. '60.

3.11. FORECASTING CIRRUS FORMED BY CONVECTION (Gayikian)

(i) If straight or anticyclonic flow exists at 300-200 mb over the area downstream from a thunderstorm area, cirrus may appear the next day and advance ahead of the ridge-line.

(ii) If the contours over the area downstream from the thunderstorm area are cyclonically curved, cirrus may or may not appear. It is most likely to appear if the flow is weak.

Ref: Gayikian, *Forecasting manual for SAC operations*, Offut, 1955, p.194.

3.12. CIRRUS ASSOCIATED WITH THE JET STREAM

3.12.1. *Cirrus associated with the jet stream* (Sawyer and Ilett)

(i) Cirrus is considerably more frequent and extensive to the right of the jet stream than to the left.

(ii) The frontal and layer types of cirrus (and medium clouds) are more common to the right of the jet stream; anvil cirrus (and altocumulus from the spreading out of cumulus) is more common to the left.

(iii) When a well-defined boundary to frontal cirrus occurs the axis of the jet stream is often close to the boundary.

Ref: Sawyer and Ilett, *Met.Mag.*, Oct. '51.

3.12.2. *Cirrus associated with the jet stream* (Frost)

From a pilot's point of view the main features consist of cirrus streamers with long tufted streaks and complex shear lines; the bands of cloud lie along the jet, sometimes for great distances together with cross striation and at times tufted cirrus with long streamers ('fallstreifen') beneath.

Ref: Frost, *Shell Aviat. News*, No. 186, 1953, p.4.
No. 195, 1954, p.14.

3.13. HEIGHT OF BASE AND TOPS OF CIRRUS

3.13.1. *Height of base and top of cirrus* (Murgatroyd and Goldsmith)

Observations of the Meteorological Research Flight are summarized in Table 3.III. below.

TABLE 3.III. *Base and top of cirrus related to the height of the tropopause*

Height of tropopause	Cirrus Base	Cirrus Top
feet	*feet*	
25 000	21 000	25 000
30 000	25 000	30 000
35 000	28 000	34 000
40 000	29 000	36 000
45 000	30 000	38 000

Note: With tropopause above 35 000 ft the departure from these mean heights may often be up to 5000 ft and occasionally greater.

Cloud thicknesses up to 12 000 ft have been observed.

Ref: Murgatroyd and Goldsmith, *Prof. Notes*, No. 119, 1956.

3.13.2. *Height of base and top of cirrus* (Gayikian)

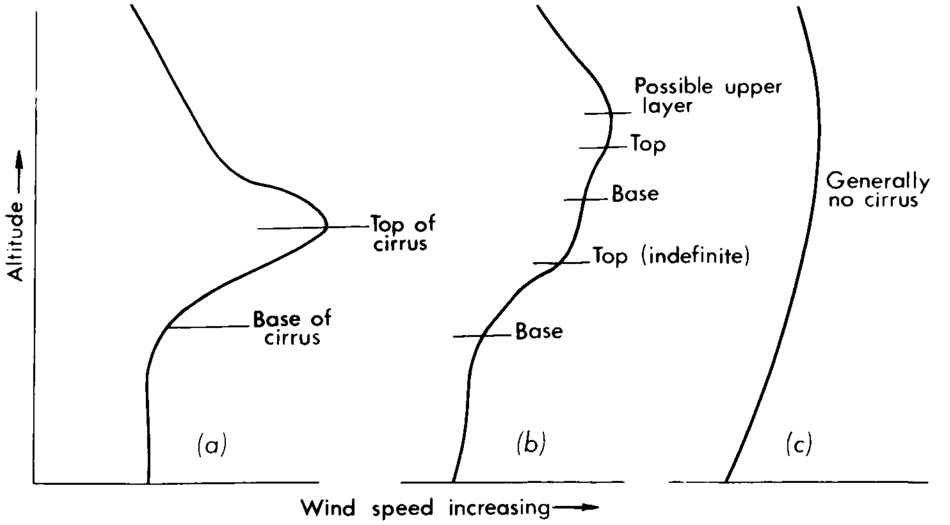


FIG. 3.3. *Association between cirrus levels and profiles of wind speed*

The presence and density of cirrus is related to variations of wind speed with height at the appropriate levels, as shown in Fig.3.3. Occurrence of cirrus is favoured and the cirrus is likely to be dense in a layer through which the wind increases rapidly with height, as in Fig.3.3.*(a)*. With a wind-speed profile such as that of Fig. 3.3.*(b)* cirrus is likely to be of low density and in layers through a great depth.

Ref: Gayikian, *Forecasting manual for SAC operations*, Offut, 1955, p.194.

3.14. THE LOWERING OF CLOUD BASE DURING CONTINUOUS RAIN (Goldman)

(i) The base of a cloud layer will be at a height where the temperature lapse changes from positive to less positive or to negative.

(ii) The base of a cloud layer will be at a height where the wet-bulb or dew-point depression is a minimum.

(iii) If the rain is of sufficient duration a ceiling (see note *(a)*) will occur below 2000 ft. Most frequently it is a ceiling below 800 ft.

(iv) During continuous rain a ceiling generally does not occur at the height of temperature discontinuity and/or maximum humidity until after the occurrence of a ceiling corresponding to the next higher level of temperature discontinuity and/or maximum humidity.

(v) The ceiling remains practically constant until the next lower cloud layer appears and increases in amount sufficiently for its base to become the ceiling.

Notes

(a) The ceiling is the height above the ground of the base of the lowest cloud layer covering more than half the sky.

(b) This technique was devised for places in the U.S.A. and has not been tested in the U.K.

Ref: HWF 16.6.2.2.
Goldman, *Mon. Weath. Rev., Washington*, 1951, p.133.

CHAPTER 4. WINDS

4.1. GEOSTROPHIC WINDS

Though the value of the geostrophic wind is often used as an approximation to the value of the actual wind, steady geostrophic flow occurs only when:

(i) the isobars (or contours) are straight, parallel and not changing with time,
(ii) there is no vertical motion, and
(iii) there are no external forces, e.g. friction.

Geostrophic scales for use with isobars on a constant-level chart (unlike those for contours on an isobaric chart) apply strictly for only one value of air density. With the Meteorological Office scales for surface charts, the correction required over the British Isles for departure from the assumed density exceeds 5% of the indicated speed only in exceptional conditions (i.e. with very low temperature and high pressure or with very high temperature and low pressure).

4.2. AGEOSTROPHIC EFFECTS

The ageostrophic wind (i.e. the vector difference between the actual wind and the geostrophic wind) can be expressed mathematically as the sum of five terms (see Appendix I.7.). One of these terms is due to ageostrophic acceleration and it is not possible to deal with this; the effect is, however, probably negligible in most cases. The remaining four terms are due to:

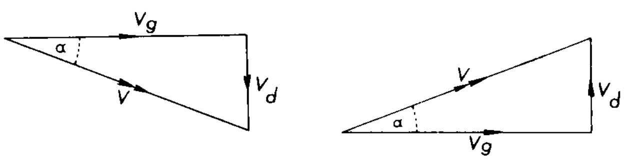
(i) *Friction*. This term is important for surface winds but negligible for winds at upper levels. The effects of friction can be allowed for in forecasting surface wind by using statistical relationships such as those given in 4.4.

(ii) *Vertical motion.* This term is negligible both near the surface and near the maximum-wind level, but may reach about 5 kt in areas of strong vertical wind shear above and below the level of the jet core. The vector is directed towards the axis above and below jet entrances and away from the axis above and below jet exits. This term therefore acts in opposition to the cross-isobar component of the advective term (see (iv) below) on the cold side of a jet but in the same sense on the warm side.

(iii) *Time changes in the pressure field (isallobaric).* This term is important near the surface when there are strong gradients of pressure tendency – not only for associated convergence or divergence but also for forecasting surface winds and frontal movement. The term frequently exceeds 10 kt and may amount to 40 kt or even more on occasions. To calculate the magnitude, draw isallobars at intervals of 1 mb per 3 h. Regard these as 1-mb isobars and measure a fictitious geostrophic wind. The isallobaric wind component will then be approximately 0.8 times the speed so measured at about 50°–55°N and fractionally less further north. The vector is directed across the isallobars towards the algebraically smaller tendencies. The isallobaric wind component may also be important in the upper air when there are strong winds and fast-moving patterns of wind. It is, however, difficult to calculate this component from upper air charts.

(iv) *Downwind changes in the pressure field (advective).* This is usually considered in two components. The component directed along the isobars or contours is the gradient wind correction (see Section 4.3.); this may be important at all levels and can be as large as 50 kt in the upper air when there are strong gradients and the contours are curved anticyclonically.

The component across the isobars (contours) is usually small near the surface but may be large in the upper air because it is proportional to the wind speed. This component is directed as shown in Fig. 4.1.



(a) V_g decreasing downwind (e.g. jet exit), V_d directed towards higher contour values

(b) V_g increasing downwind (e.g. jet entrance), V_d directed towards lower contour values

FIG. 4.1. *Geostrophic departure, V_d, due to change of geostrophic speed along the contour*

V_g = geostrophic wind

V = resultant, i.e. actual wind when V_d is the only geostrophic departure

$V_d = V_g \tan\alpha$.

Table 4.I gives some examples of α and $\tan \alpha$ along the axis of a jet stream.

TABLE 4.I. *Values of α and $\tan \alpha$ for various distances along the jet axis between isotachs at 40-kt intervals*

Latitude	Distances along the jet axis between isotachs at 40-kt intervals (n.miles)									
	120		180		240		300		360	
	α	$\tan \alpha$	α	$\tan \alpha$	α	$\tan \alpha$	α	$\tan \alpha$	α	$\tan \alpha$
50° N	40°	0.85	29°	0.55	23°	0.40	18°	0.30	15°	0.25
60° N	37°	0.75	26°	0.50	20°	0.35	17°	0.30	14°	0.25

Values of $\tan \alpha$ are to the nearest 0.05.

Note

The angle α depends on the rate at which V_g changes downwind but is independent of the value of V_g itself.

Ref: Saucier, *Principles of meteorological analysis*, University of Chicago Press, 1955, p.242.
Simla, *Circ. Met. Div. Dep.Transp., Toronto*, No.3018, 1958.

4.3. GRADIENT WIND CORRECTION

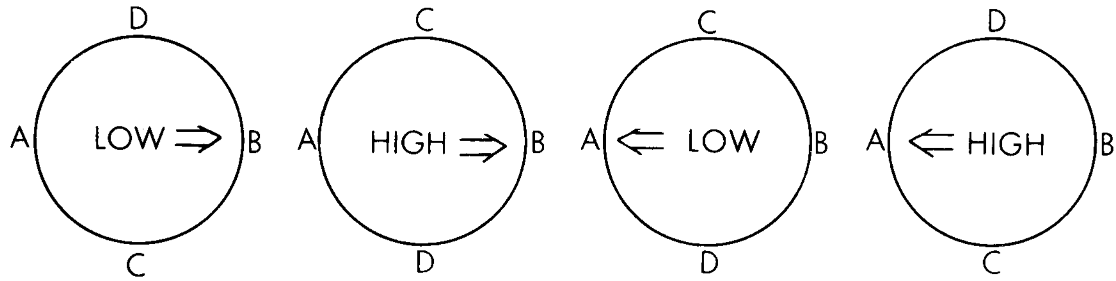
TABLE 4.II. *Correction to be applied to geostrophic wind speed to obtain gradient wind speed when the system of isobars or contours is stationary*

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

Notes

(a) The corrections are given to the nearest 5 kt. The theoretical maximum gradient wind for anticyclonic curvature is equal to twice the geostrophic wind.

(b) If the system of contours or isobars is moving, the gradient wind correction is related to the appropriate value in Table 4.II. as follows:



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

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

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

(c) For 200-mb charts Zobel has found that the gradient wind computed from contour curvature, even without adjustment for the movement of the system, gives a better estimate of the actual wind than does the geostrophic wind.

(d) For surface charts Boyden has found that the geostrophic wind gives a better estimate of the actual wind just above the friction layer (i.e. at 900 m) than does the gradient wind computed from isobar curvature, except when the radius of curvature is small and cyclonic. When the radius is 500 n.miles one-third of the gradient wind correction should be applied and when 250 n.miles, a half.

Ref: London, Met. Off., *Gradient wind tables.* Unpublished.
Zobel, *Met.Mag.,* Feb.'58.
Boyden, *Met.Mag.,* Apr. '63.

4.4. SURFACE WIND (Findlater *et alii*)

A study of the ratio (V_0/V_{900}) between the surface wind, V_0, and the 900-mb wind, V_{900}, and the angle, α, by which V_0 is backed from V_{900}, yielded the following overall relationships:

		V_0/V_{900}	$\alpha°$
Open sea	OWS I	0.82	7
	OWS J	0.77	9
Land	Heathrow	0.41	20
	Leuchars	0.38	19

These relationships may be used in forecasting. They may be further refined by taking into account the lapse rate and 900-mb wind speed.

The lapse rate classification is illustrated in Fig. 4.2.

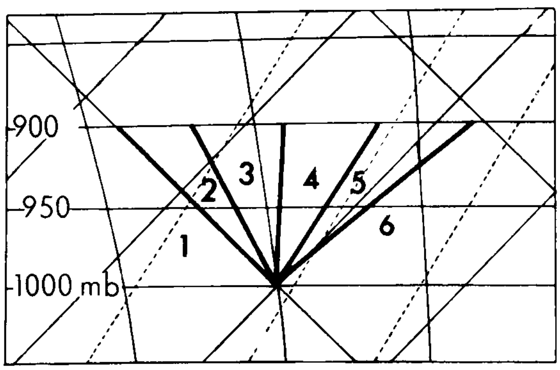


FIG. 4.2. *Schematic representation of lapse rate classes*

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

(i) *Forecasting the surface wind over the open sea*

(a) Forecast the wind at 900 mb on the assumption that this will be the same as the wind which would blow under the influence of the forecast sea-level isobaric pattern if the sea surface were frictionless. i.e. forecast the 900-mb wind on the basis of the geostrophic relationship but allowing, if necessary, for curvature of the air trajectory and isallobaric effects.

(b) Use Fig.4.2. to decide which lapse rate class will be appropriate.

(c) Using Table 4.III., forecast the surface wind to be backed from

the 900-mb wind (V_{900}) by α° and to have a speed V_0 according to the appropriate ratio V_0/V_{900}.

TABLE 4.III. V_0/V_{900} *and* α *from data for ocean weather stations I and J*

Lapse class	900-mb wind speed class (kt)									
	10-19		20-29		30-39		40-49		$\geqslant 50$	
	V_0/V_{900}	α°	V_0/V_{900}	α°	V_0/V_{900}	α°	V_0/V_{900}	α°	V_0/V_{900}	α°
1	0.95	0	0.90	0	0.85	0	0.80	0	0.90	0
2	0.90	5	0.85	5	0.80	5	0.75	5	0.75	5
3	0.85	10	0.75	10	0.70	15	0.65	10	0.65	10
4	0.80	15	0.70	20	0.65	20	0.60	20	0.60	20
5	0.75	15	0.70	20	0.65	20	0.60	20	0.55	25
6	see below									

Notes

Data for lapse class 6 are too few for values to be given for each wind speed class. The means for all observations in lapse class 6 are $V_0/V_{900} = 0.75$ and $\alpha = 25^{\circ}$.

Values for V_0/V_{900} are given to the nearest 0.05 and for α to the nearest 5°.

(ii) *Forecasting the surface wind over a fairly flat land surface*

It is not possible to give a general technique for land stations because effects of lapse rate and 900-mb wind speed are at times overshadowed by effects of orography which differ from place to place and from one wind direction to another. Table 4.IV. may give some guidance in forecasting surface winds over fairly flat terrain. It should be appreciated, though, that even over fairly flat terrain, such as surrounds Heathrow, orographic effects are at times significant.

TABLE 4.IV. V_0/V_{900} *and* α *for London/Heathrow Airport*

Lapse class	900-mb wind speed class (kt) 10–19		20–29		30–39		40–49		≥50	
	V_0/V_{900}	α°	V_0/V_{900}	α°	V_0/V_{900}	α°	V_0/V_{900}	α°	V_0/V_{900}	α°
Day-time										
1, 2	0.65	5	0.55	5	0.50	10	0.50	10	0.35*	15*
3	0.50	20	0.45	20	0.45	20	0.45	20	0.45	15
4	0.45	35	0.45	30	0.40	25	0.30	20	0.40*	25*
5, 6	0.35	45	0.40	35	0.35	35	0.40*	30*	0.40*	30*
Night-time										
1, 2	0.25*	20*	0.35	25	0.30*	35*	0.40*	15*	0.40*	25*
3	0.35	25	0.35	30	0.35	25	0.35	20	0.35	15
4	0.30	35	0.30	35	0.30	30	0.35	30	0.35*	15*
5, 6	0.30	45	0.25	40	0.25	35	0.30*	30*	No obs	

*Results based on less than 10 observations.

Values for V_0/V_{900} are given to the nearest 0.05 and for α to the nearest 5°.

Ref: Findlater *et alii, Scient. Pap.* No. 23.

4.5. VARIATION OF WIND SPEED IN THE LOWEST FEW HUNDRED FEET OVER FAIRLY LEVEL GROUND

4.5.1. *Variation with height when winds are strong* (Shellard)

Any law expressing wind speed as a function of height will only apply strictly to one particular place and to one temperature lapse rate. However, in strong winds the temperature lapse rate will be adiabatic, or neutral, because of the thorough mixing in the lower layers. Most investigations have found that in neutral conditions the variation of wind speed (V) with height (h) in the boundary layer can most simply be expressed by the formula $V = k\,h^p$ where k is a constant and p depends on the roughness of the terrain. Fig. 4.3. is based on this relationship.

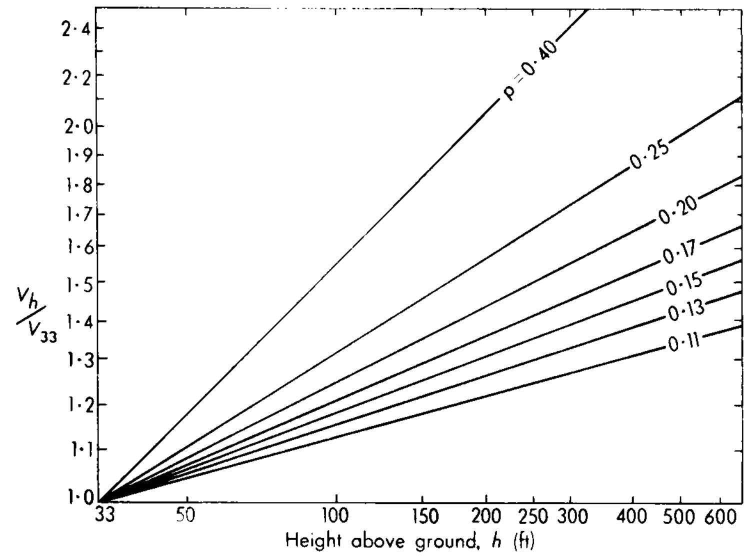


FIG. 4.3. *Ratio of wind speed at height h (V_h) to that at 33 feet (V_{33}) for adiabatic conditions*

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

Note

The relationships given in Fig. 4.3. apply only with neutral stability in the lower layers (as might be expected with strong winds) and with flow over fairly level ground. Some results for various lapse rates are summarized in 4.5.2. Significant departures from the relationships may occur in the vicinity of hills.

Ref: Shellard, *Met.Mag.*, Aug. '67.

4.5.2. *Variation according to lapse rate* (Frost)

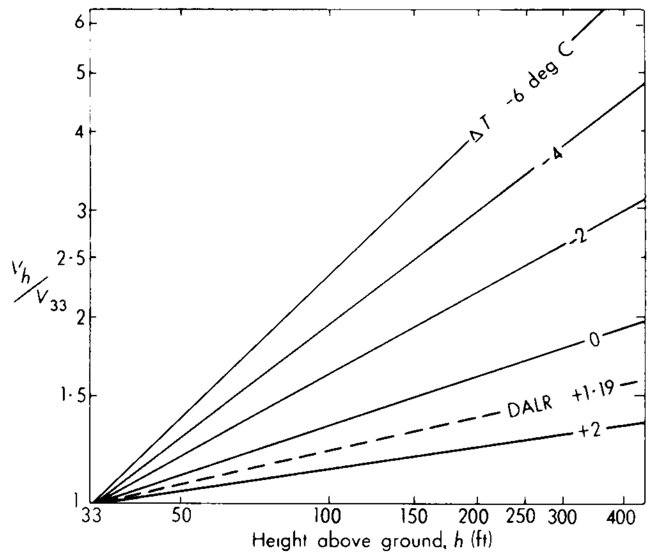


FIG. 4.4. *Ratio at Cardington of wind speed at height h (V_h) to that at 33 ft (V_{33}) for different values of ΔT*

ΔT = temperature at 4 ft minus temperature at 400 ft (degC) (heights above ground)

Note

At Cardington the country is fairly level and open with few trees. Figure 4.4. is unlikely to be applicable for winds over other types of surface.

Ref: Frost, *Met.Mag.*, Jan. '47.

4.6. GUSTS NEAR THE SURFACE (Shellard)

Shellard has shown that the ratio of the maximum gust to mean hourly speed in strong winds is mainly dependent on surface roughness. For anemograph sites in the U.K. variations in the ratio from 1.4 to 2.3 were found. Using additional data, more recent unpublished work has shown considerable agreement with the original calculations. From anemograph site characteristics it is possible to classify the values of the ratio according to terrain roughness as in Table 4.V.

TABLE 4.V. *Ratio of maximum gust to mean hourly speed for open sea and various terrains*

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

Notes

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

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

Ref: Shellard, *Climat.Memor.* No.50.

4.7. FORECASTING PEAK WIND GUSTS IN NON-FRONTAL THUNDERSTORMS (Fawbush and Miller)

(i) Plot on a tephigram the wet-bulb curve for an upper air ascent which is representative of the air mass; only the portion of the curve near 0°C is required.

(ii) From the intersection of this curve with the 0°C isotherm trace downwards the appropriate saturated adiabatic.

(iii) The intersection of this adiabatic with the surface isobar gives a forecast of the temperature at the surface in the downrush of air beneath the thunderstorm.

(iv) According to the expected temperature fall in the downrush (i.e. surface temperature just before the storm minus temperature expected in the downrush) the peak wind speed can be forecast from the regression curve of Fig. 4.5.

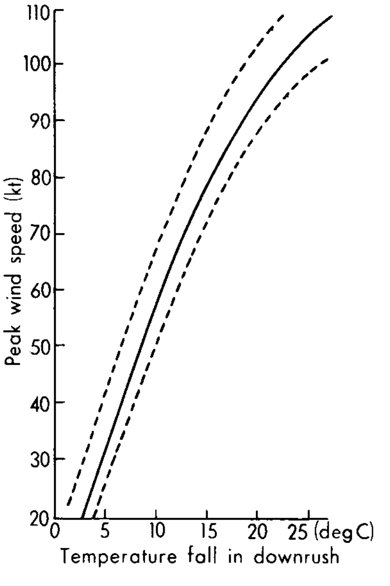


FIG. 4.5. *Peak wind speed in thunderstorms in the U.S.A.*

——— Regression curve
– – – Standard error of estimate

Note

The technique was derived from a study of thunderstorm squalls in the U.S.A. McNair and Barthram, investigating a squall-line over England, found the technique gave a good estimate of the maximum gusts.

Ref: HWF 13.9.6.
Fawbush and Miller, *Bull. Am. Met. Soc.*, 1954, p.14
McNair and Barthram, *Met. Mag.*, Oct. '66.

4.8. GUST VARIATION WITH HEIGHT UP TO 500 FEET (Deacon)

At Sale (Victoria, Australia) the gust speed was found to vary with height as shown in Fig. 4.6. on occasions when the mean wind at 40 ft was more than 17 kt.

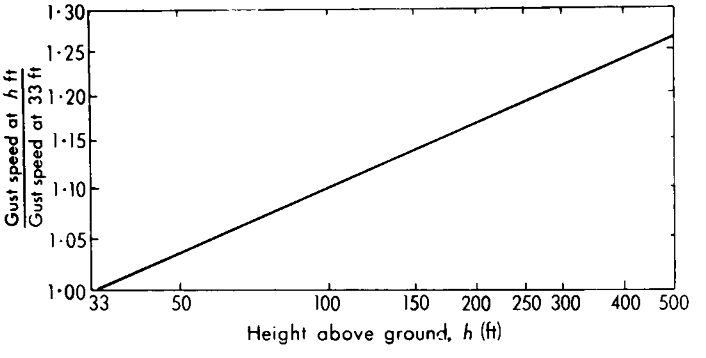


FIG. 4.6. *Gust variation with height at Sale, Australia*

Note

The relationship was derived from observations made over rolling grassland with few trees.

Ref: Deacon, *Q. Jnl R. Met. Soc.*, 1955, p.562.

CHAPTER 5. UPPER AIR AND SURFACE CHARTS

5.1. PHRASEOLOGY

5.1.1. *Zonal and meridional flow*

A predominantly W-to-E airflow is termed *zonal* and an E-to-W airflow is generally regarded as *negative zonal.* Both N-to-S and S-to-N airflows are termed *meridional.* The strength of the zonal flow in any sector may be expressed in terms of a *zonal index* given by the difference in average contour height along two latitude circles through the sector; the zonal index is often (but not always) measured between 35° and 55° latitude. Zonal flow will give a high zonal index and meridional flow a low zonal index.

5.1.2. *Jet cores and jet axes*

A *jet core* is a line, usually varying in height, along which the wind speed is a maximum. The line of maximum wind speed on an isobaric surface is known as the *jet axis.*

5.1.3. *Long and short waves*

Long waves are smooth, persistent and slow-moving features of a wave pattern, having an amplitude of about 20° of latitude, and a wave-length such that, typically, four or five long waves are to be found in the hemisphere. They extend through the middle and high troposphere and are still apparent in the lower stratosphere up to, say, 100 mb. A long-wave trough may be termed an *anchor trough* if it is expected to continue quasi-stationary for some time (e.g. the trough often found to the lee of the Rockies). *Short waves* are shorter-lived, faster-moving features, with a smaller amplitude and shorter wavelength (say less than 60° longitude) than the long-wave pattern on which they are normally superimposed and through which they move. They do not extend through the high troposphere and are normally not apparent on the 100-mb chart.

5.1.4. *Progression and retrogression*

Movement of a wave train or of closed centres in a W-to-E direction is termed *progression* and movement in the opposite direction is termed *retrogression.*

5.1.5. *Trough extension and relaxation*

If an upper trough or thermal trough develops a markedly increased amplitude it is said to have undergone *meridional extension* whilst if the amplitude decreases with time it is said to have undergone *relaxation.* These changes are often measured in terms of the latitude change of a chosen contour or thickness line.

5.1.6. *Trough disruption*

If the northern portion of a trough moves forward leaving a quasi-stationary cut-off low in the base of the trough, the process is described as ***anticyclonic disruption.*** The opposite case where the southern portion of the trough advances, perhaps developing a cut-off circulation, whilst the northern portion of the trough becomes quasi-stationary is termed *cyclonic disruption.*

5.1.7. *Blocking*

Blocking is the obstruction on a large scale of the normal E-to-W progression of surface cyclones in the middle latitudes; the upper-flow pattern changes from predominantly zonal to markedly meridional. This change is accompanied by anticyclonic development in high latitudes and cyclonic development in low latitudes. The upper current divides upwind of the established block, flows round the vortices, and recombines to form a single current downstream (Fig.5.1.).

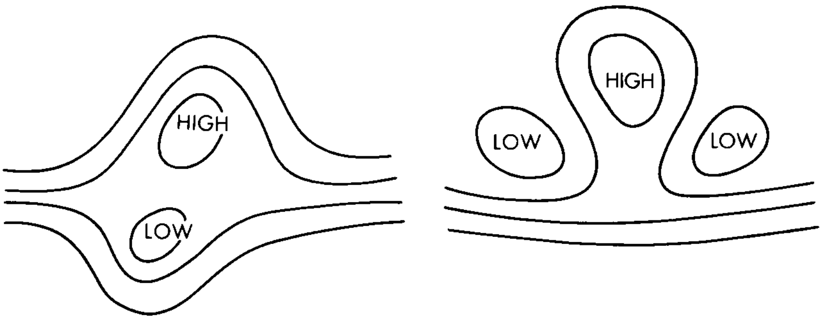


FIG.5.1. *Contour patterns associated with blocks*

If the southern branch of the stream suffers relatively little distortion, as in the right-hand pattern of Fig. 5.1. it is referred to as an *omega block.* The anticyclone is termed a *blocking high* and the cyclonic eddy a *cut-off low.*

5.1.8. *Confluence and diffluence*

Confluence is the approach of adjacent streamlines in the direction of flow; *diffluence* is the increasing separation of the streamlines (Fig. 5.2.). The terms can be used to describe features on both contour and thickness charts.

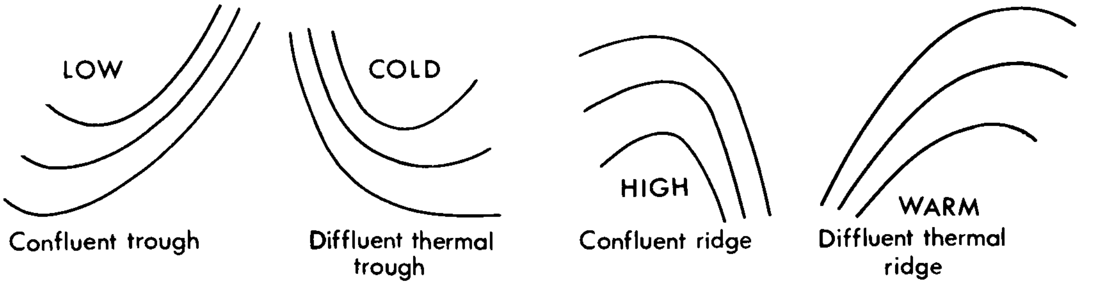


FIG.5.2. *Examples of confluence and diffluence*

5.1.9. *Divergence and convergence*

The *divergence* (of velocity) is the time rate of expansion of air per unit of volume; convergence is negative divergence.

The divergence, $\mathrm{div}V \equiv \nabla . V = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z}$.

The horizontal divergence, $\mathrm{div}_H V \equiv \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}$.

As $\mathrm{div}V$ in the atmosphere is small, $\mathrm{div}_H V \approx -\frac{\partial w}{\partial z}$.

Horizontal divergence is therefore linked with vertical contraction of an air column and horizontal convergence with vertical expansion. Thus, at the surface, horizontal divergence will lead to downward motion of the air above and horizontal convergence will lead to upward motion.

5.1.10. *Vorticity*

The (vector) spin of a small element of fluid is specified by its *vorticity;* this by convention equals twice the angular velocity of the element. The component of *relative vorticity* about the local vertical,

$\zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} = \frac{V}{r} + \frac{\partial V}{\partial n}$ and is counted positive when the spin is cyclonic and negative when the spin is anticyclonic.

The *absolute vorticity* equals the relative vorticity plus the vorticity resulting from the earth's spin, i.e. $\zeta_a = \zeta + f$. The simplified vorticity equation indicates that $\frac{\partial \zeta_a}{\partial t} \approx -\zeta_a \,\mathrm{div}_H V$. Horizontal convergence is therefore associated with the generation of cyclonic vorticity (as well as with vertical motion).

5.1.11. *Barotropic and baroclinic atmospheres*

A *barotropic* atmosphere is a (hypothetical) one in which the surfaces of constant pressure and constant temperature coincide at all levels, i.e. the surfaces are also surfaces of constant density. Such an atmosphere would be non-developmental and the thickness gradients would be zero. If thickness lines are widely spaced the atmosphere is sometimes termed *quasi-barotropic*. In a *baroclinic* atmosphere the isobaric and isopycnic (constant density) surfaces intersect at some levels; the temperature on a given pressure surface then normally varies from place to place and thickness gradients exist. The degree of baroclinicity in a vertical section of the atmosphere is given by the product of the thermal wind and the Coriolis parameter and is therefore indicated by the gradient of thickness.

5.1.12. *Ana-front and kata-front*

Fronts may be regarded as circulation systems. When the warm air ascends relative to the cold air the front is termed an *ana-front*; when the warm air descends relative to the cold air it is termed a *kata-front* (Fig.5.3.).

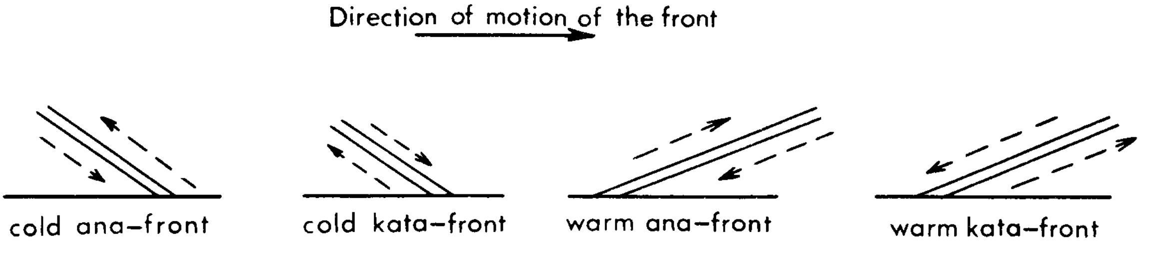


FIG.5.3. *Ana-fronts and kata-fronts*

5.2. JET-STREAM ANALYSIS

5.2.1. *Average position of jet relative to surface fronts*

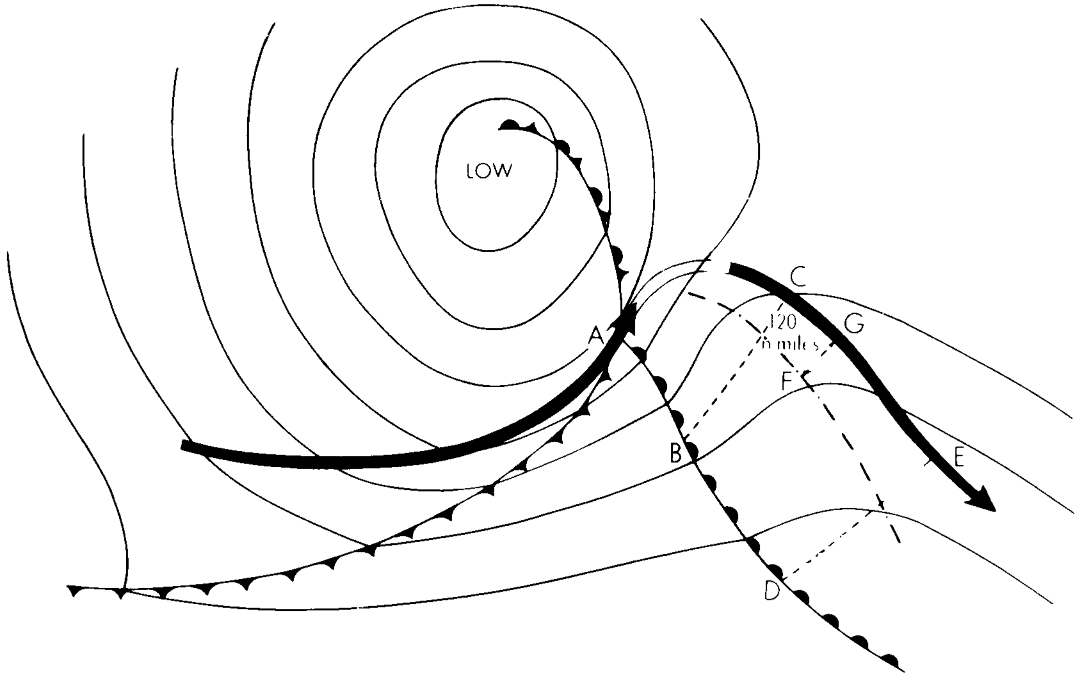


FIG. 5.4. *Jet stream in relation to surface fronts*
·—·—· Axis of surface ridge associated with warm front.
Broad arrow represents the 300-mb jet, the unshaded portion being weak or broken.

(i) The jet normally passes over tip of warm sector A in a direction roughly parallel to the warm-sector isobars near A.

(ii) The warm-front jet normally passes through a point G about 120 n.miles ahead of the surface ridge where this latter is farthest from the front (i.e. at F); G will normally be 300 to 450 n.miles from the front itself, the larger distances usually occurring when the front is slow moving.

(iii) The distance of the jet from the warm front remains roughly constant towards lower pressures until the point C, opposite B, where AB equals 7° latitude.

(iv) The distance of the jet from the warm front does not decrease towards higher pressures, i.e. DE $\geqslant$ BC.

(v) On average, cold-front jets diverge from the front by 140 n.miles for each 600 n.miles along the front.

Ref: Boyden, *Met.Mag.*, Nov.'63.

5.2.2. *Position of jet stream relative to thickness lines*

Jet axis at 300 mb is on average 60 n.miles to the cold side of the peak 1000-500 mb thickness gradient.

Ref: Boyden, *Met.Mag.*, Sept.'63.

5.2.3. *Position of jet axis relative to contour lines*

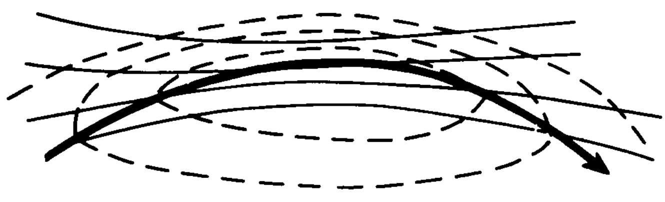


FIG.5.5. *Position of jet axis relative to contours*

Jet axes are curved anticyclonically relative to the contour lines, i.e. across the contours towards low pressure at jet entrances and towards high pressure at jet exits. For angle of cross-contour flow, see 4.2.(iv).

5.2.4. *Vertical shear*

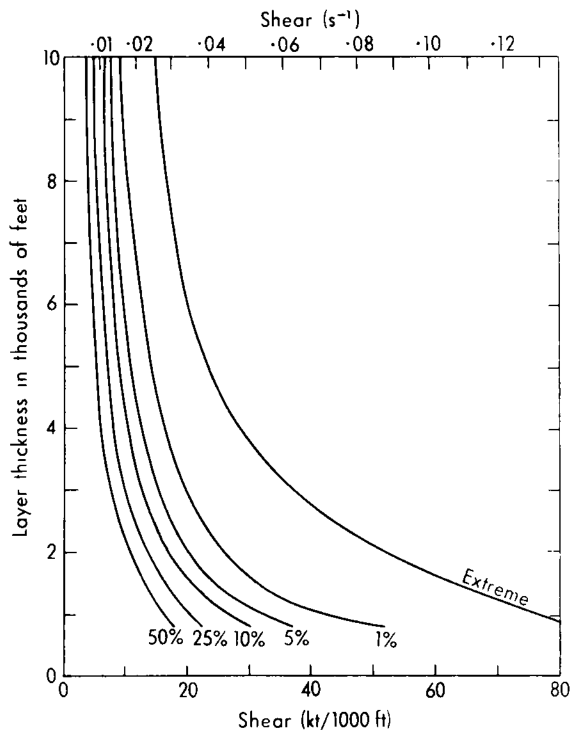


FIG.5.6. *Estimated frequencies and extremes of vertical shear*

Ref: Crossley, *Scient. Pap.* No.17.

5.2.5. *Mean horizontal shear*

Mean percentages of peak speed at various distances on either side of the jet axis or core (jet speeds ⩾ 100 kt) are shown in the table below.

Distance from axis (n.miles)						
on warm side				on cold side		
150	100	50	0	50	100	150
70	77	89	100%	89	72	57

Ref: Boyden, *Met.Mag.*, Sept.'63.

5.2.6. *Extreme cyclonic shear*

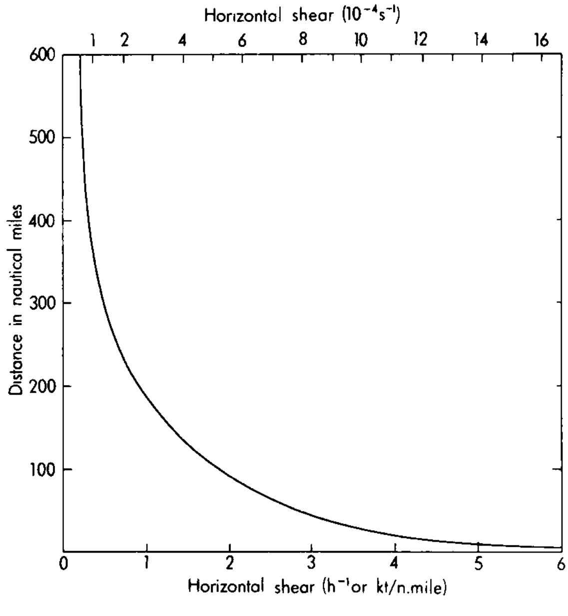


FIG.5.7. *Curve of estimated extreme horizontal cyclonic shear*

Ref: Crossley, *Scient.Pap. No.17.*

5.2.7. *Extreme anticyclonic shear*

Theoretically the anticyclonic shear cannot exceed the Coriolis parameter, the values of which are given in the following table.

Latitude (degrees)	30	35	40	45	50	55	60	65
Coriolis parameter (kt/100 n.miles)	26	30	34	37	40	43	45	47

These values are only rarely, if ever, exceeded, but may be attained within 3–5° latitude of the jet axis on the warm side.

Ref: Crossley, *Scient. Pap.* No. 17
(with arithmetical corrections)

5.2.8. *Indirect circulations associated with jet streams*

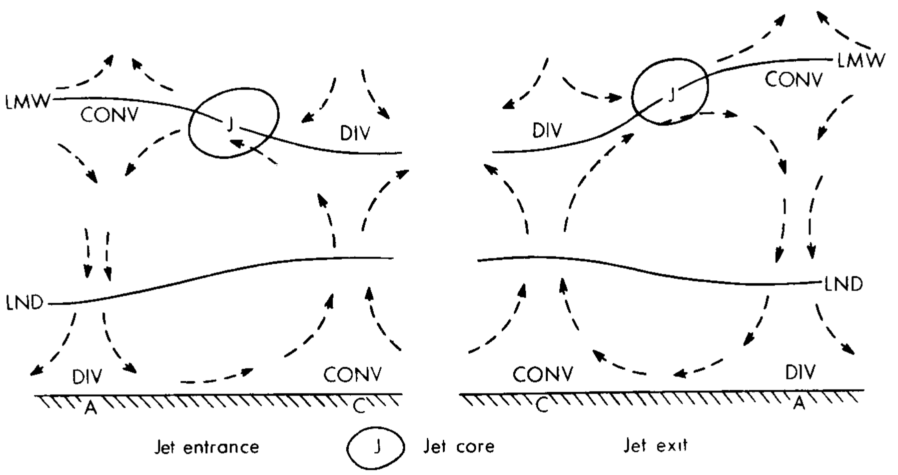


FIG. 5.8. *Section, looking downstream, showing indirect circulations associated with jet*

LMW – level of maximum wind LND – level of non-divergence

(i) The level of maximum wind is a level of maximum convergence/divergence but of minimum (zero) vertical motion.

(ii) The level of non-divergence is a level of maximum vertical motion.

(iii) Cross-contour flow at jet level is towards low pressure at entrances (sub-geostrophic) and towards high pressure at exits (super-geostrophic).

(iv) C and A indicate surface development areas (see 5.6.1.).

Ref: HWF Fig. 5.16.

5.3. CONTOUR AND THICKNESS PATTERNS

5.3.1. *Instability of zonal flow*

A W'ly airstream which keeps roughly the same direction over a considerable distance is usually found to have thickness lines nearly parallel to the isobars. Such a flow seldom persists for more than two or three days without developing a mobile wave of moderate amplitude. The advection accompanying this buckling of the flow produces a concentration of thickness lines which initially is downwind of each ridge and trough. The majority of jet streams develop as an outcome of this process.

Ref: *Mét. Rep.* No.21.
Boyden, *Met.Mag.*, Oct.'63.

5.3.2. *Trough and ridge movement* (Rossby)

(i) Estimate to the nearest 5°, the central latitude of the 500-mb flow pattern covering the trough whose movement is required and the next *major* (long-wave) trough upwind; note the wavelength L between the two trough axes. Using a sector from the 10° meridian just W of the upwind trough to the 10° meridian just E of the trough in question, and a band width 10° either side of the central latitude, note the contour-height differences across the band at 10° meridian intervals. The mean contour-height difference gives a zonal index.

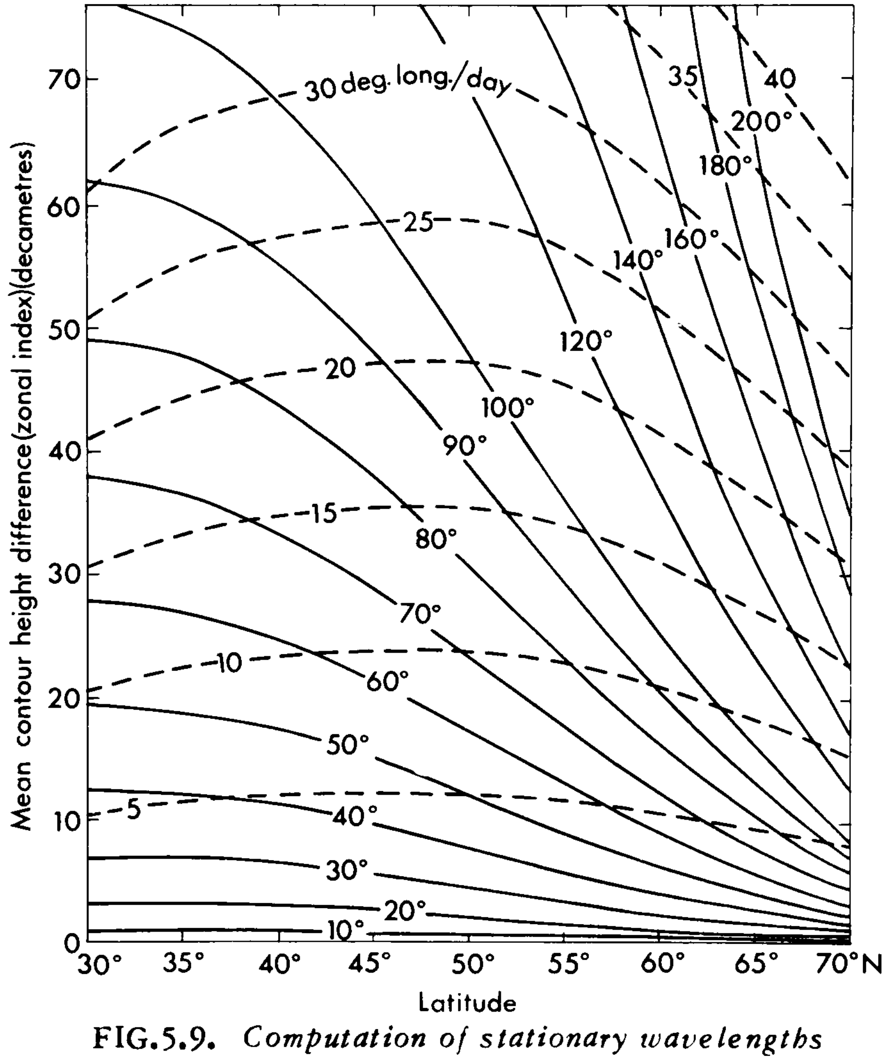


FIG.5.9. *Computation of stationary wavelengths*

——— Lines of equal stationary wavelength expressed in degrees of longitude measured along the appropriate latitude circle

— — — Zonal displacement of trough in degrees of longitude per day

(ii) With the chosen central latitude as abcissa and the computed zonal index as ordinate, read from Fig.5.9. the stationary wavelength L_s (full lines) and the zonal displacement (pecked lines) in degrees per day. Read off also at the same central latitude the zonal displacement corresponding to the actual wavelength L. The difference between these two displacements gives the probable 24-h displacement of the downwind trough, progressive if $L < L_s$ and retrogressive if $L > L_s$.

Ref: *Met.Rep.* No. 21.

5.3.3. *Cut-off highs*

Development of a closed high at 300 mb normally requires the existence of a jet from a S'ly point, not veering with time and preferably backing ahead of an upper trough or low. Once formed, the high moves on average towards 080°, but the track shows a progressive veer with time. The speed of movement is usually about 200-300 n.miles a day, but about one in three of the highs associated with SE'ly jets remains almost stationary. Of all closed highs 50% last for no more than 3 days, while 25% persist for 5-10 days.

Ref: Boyden, *Met.Mag.*, Oct.'63.

5.3.4. *Cut-off lows*

Cut-off low development at 300 mb normally requires the existence of a jet from a N'ly point associated with a 300-mb trough with an amplitude of at least 10° of latitude; such development is very unlikely if the jet weakens by 20-30 kt. The probability of the formation of a cut-off low depends upon the direction of the jet, as shown in Fig.5.10.

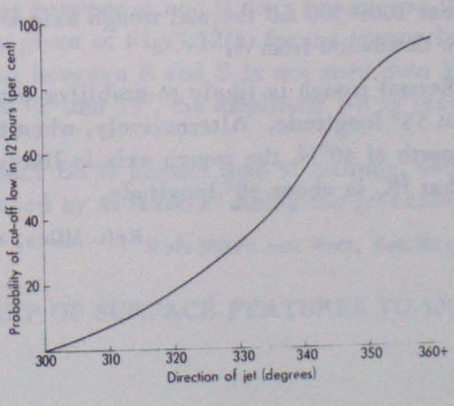


FIG.5.10. *Probability of a cut-off low in relation to jet direction*

About half the total number of 300-mb lows move SE during the first 24 h although if there is already an eastward-moving surface system the upper low usually moves E or N of E. 85% of the lows move at a speed of 10-20 kt during the first 24 h.

Ref: Boyden, *Met.Mag.*, Oct.'63.

5.3.5. *Meridional extension of thermal troughs*

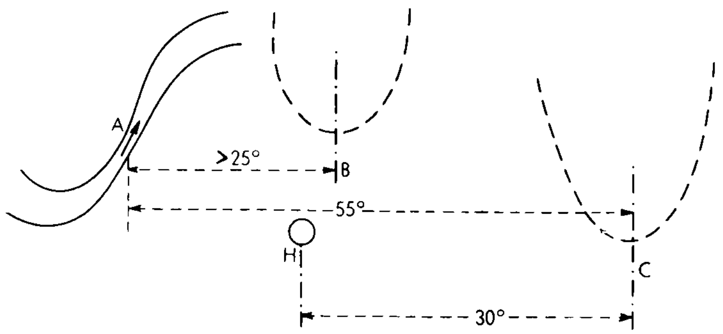


FIG.5.11. *Conditions for meridional extension of a thermal trough*

On 77% of occasions meridional extension of a thermal trough by at least 5° of latitude will begin within 12-24 h when a strong S'ly flow develops upstream, provided that

(i) the direction of the 500-mb flow at the inflexion point A (Fig. 5.11.) lies between 150° and 225°,

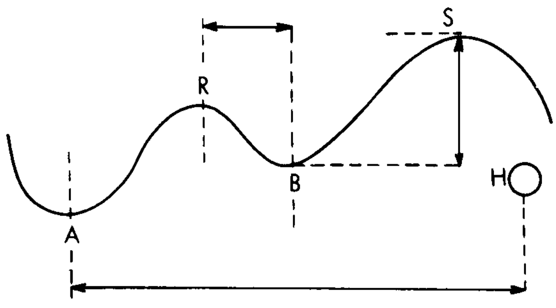
(ii) the speed of the 500-mb flow at A is not less than 40 kt, and

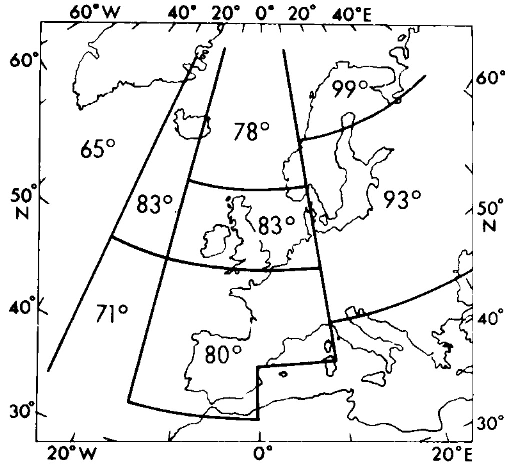
(iii) the nearest 1000-500 mb thermal trough axis B is at least 25° longitude downwind from A.

The axis of the thermal trough is likely to stabilize within 48-72 h at C, where AC is about 55° longitude. Alternatively, when there is a strong surface high, H, north of 40°N, the trough axis is likely to stabilize in such a position that HC is about 30° longitude.

Ref: Miles, *Met.Mag.*, July '59.

5.3.6. *Relaxation of thermal troughs*





(a) Identification of thermal troughs (A,B) and ridges (R,S) and downwind anticyclone (H)

(b) Maximum separation (degrees longitude) between downwind anticyclone H and upwind trough A, for expectation of relaxation of trough B

FIG.5.12. *Relative positions of thermal troughs and ridges and downwind anticyclone for relaxation of trough* B

Relaxation of thermal trough B by at least 5° latitude can be expected within 24 h, provided that

(i) there is a thermal trough, A, upwind of B, and a surface anticyclone, H, downwind of B (Fig.5.12.(*a*)),

(ii) the central pressure of H is not less than 1020 mb (February to September) or 1024 mb (October to January),

(iii) the spacing between A and H does not exceed the maximum separation given in Fig.5.12*(b)* for the appropriate location of H,

(iv) the spacing between R and B is not more than 35° longitude and if between 30° and 35°, the amplitude BR is less than 15° latitude, and

(v) the amplitude BS is greater than 5° latitude, and if less than 15° has increased by at least 2° during the preceding 24 h.

Ref: Miles and Watt, *Met.Mag.*, May '62.

5.4. RELATIONSHIP OF SURFACE FEATURES TO 500-mb CONTOUR PATTERNS

5.4.1. *Upper low*

Surface low: at first usually to E or NE of the upper low, but mature centres are concentric. In typical cold-cored depressions the circulation intensifies upwards, but it decreases with height in warm-cored depressions such as tropical hurricanes and some old occluding lows.

Exceptions to this rule are continental cold-pool vortices in winter, which may lie over a surface high and some low-latitude cut-off lows which may lie over a surface trough only.

5.4.2. *Upper trough*

Surface trough or low: beneath it or on its eastern flank.

5.4.3. *Upper high*

Surface high: beneath it or to the **E** of it.

5.4.4. *Upper ridge*

Surface high or ridge: directly below or on its eastern flank.

5.4.5. *Broad zonal flow*

Surface zonal flow: but incipient wave depressions or breakaways may be embedded in the stream and yet not be reflected in the upper flow.

The first indication at 500 mb of a developing surface wave is normally a weak trough just behind the surface feature and a weak ridge just ahead of it.

5.4.6. *Jet axis*

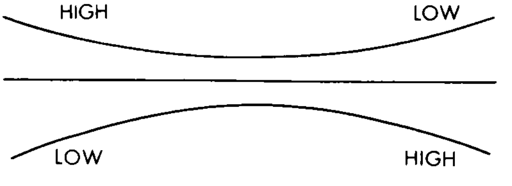


FIG.5.13. *Preferred locations of surface lows and highs in relation to a jet stream*

5.5. RELATIONSHIP OF SURFACE FEATURES TO 1000-500 mb THICKNESS PATTERNS

5.5.1. *Frontal depressions*

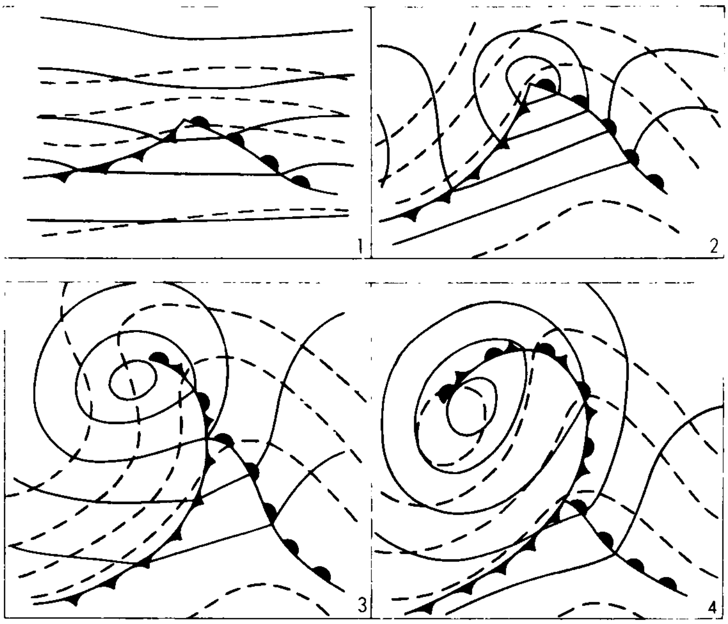


FIG.5.14. 1000-500 *mb thickness patterns over a frontal depression at various stages of development*

—— Isobars ---- 1000-500 mb thickness lines

5.5.2. *Anticyclones*

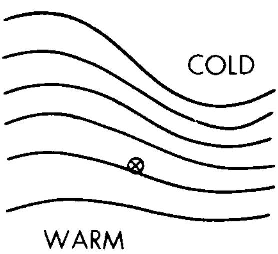


FIG.5.15. *(a) Mobile cold high*

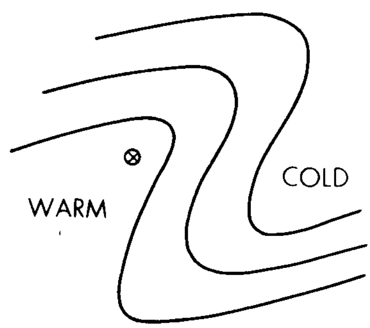


FIG.5.15. *(b) Warm high*

5.6. USE OF UPPER AIR CHARTS IN FORECASTING DEVELOPMENT AND MOVEMENT OF SURFACE PRESSURE SYSTEMS

5.6.1. *Sutcliffe development patterns on thickness charts*

Developments at the surface are indicated by the sign of the relative divergence, $\mathrm{div}V_{500} - \mathrm{div}V_{1000} = \mathrm{div}V'$, which is given by the equation

$$-f\,\mathrm{div}V' = \underset{\substack{\text{latitude}\\ \text{term}}}{V'\frac{\partial f}{\partial s}} + \underset{\substack{\text{thermal}\\ \text{steering}\\ \text{term}}}{2V'\frac{\partial \zeta_0}{\partial s}} + \underset{\substack{\text{thermal}\\ \text{development}\\ \text{term}}}{V'\frac{\partial \zeta'}{\partial s}}$$

The right-hand side of the equation is negative for cyclonic development (C) and positive for anticyclonic development (A). Major development areas can be located on 1000-500 mb thickness charts by recognition of vorticity changes in the direction of flow, as shown by examples in Fig.5.16.

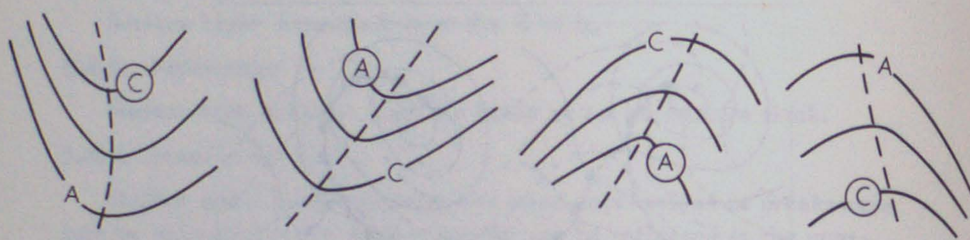


FIG.5.16. *Development areas in relation to 1000-500 mb thickness patterns*

A Anticyclonic development C Cyclonic development

Notes

(a) The thermal wind speed, V', appears in each term, so the stronger the thermal field, the stronger the development.

(b) The latitude term contributes to A development in northward motion and to C development in southward motion (cf. 5.3.2. and 5.3.3.).

(c) The steering term causes advection of surface pressure systems along the thickness lines. The steering will be most effective when the lines are straight and close together; if the thickness lines are curved, or the thermal gradient weak, the steering term may be swamped by the development term.

(d) The development term will be most effective where indicated in the diagrams above, and particularly so where the C and A areas are ringed.

(e) Petterssen has shown that these development areas may be identified similarly in relation to the patterns on 300-mb contour charts.

Ref: HWF, Chap. 5.

5.6.2. *Deepening of depressions over the Atlantic north of 35° N* (George)

(i) The position of the surface centre is marked on the 500-mb chart for the appropriate time. Deepening of the depression over the next 24 h is forecast if the contour through this position is open and if the position lies between a trough (upstream) and a ridge.

(ii) The amount by which the depression will deepen over the next 24 h can be assessed from Fig.5.17.(*c*) after evaluating ΔH and ΔT from Figs 5.17.*(a)* and *(b)*.

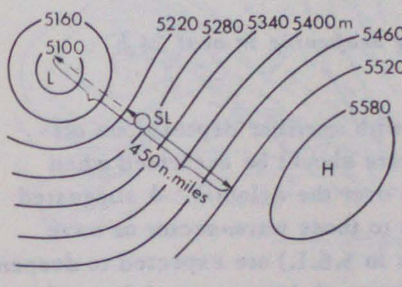


FIG.5.17. *(a) Determination of* ΔH *from 500-mb contours*
$\Delta H = 300$ m

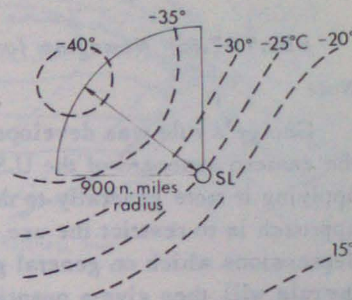


FIG.5.17.*(b) Determination of* ΔT *from 500-mb isotherms*
$\Delta T = 17$ degC

ΔH = the difference (metres) in 500-mb height across a distance of 900 n.miles centred at the position of the surface low (SL) and measured normal to the contours. Height changes within closed contours should be ignored. Thus, in the example of Fig.5.17*(a)* $\Delta H = 300$ m.

ΔT = the difference (degC) at 500 mb between the temperature above the surface low and the lowest temperature within a distance of 900 n. miles in the NW quadrant. Thus, in the example of Fig.5.17.*(b)* $\Delta T = 17$ degC.

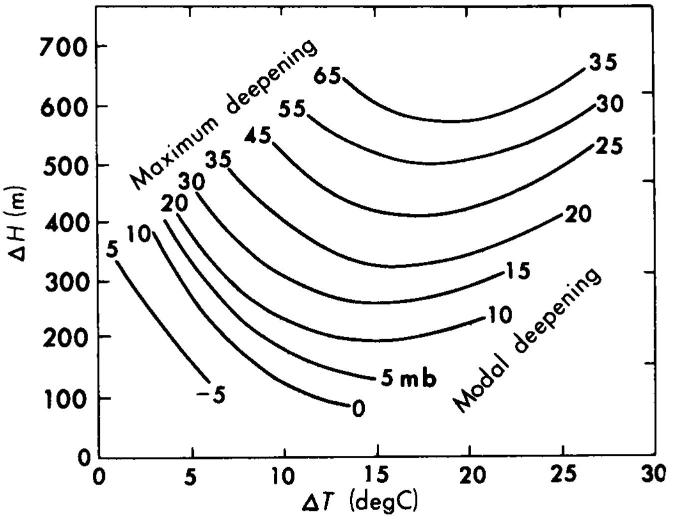


FIG.5.17.*(c) Nomogram for predicting deepening in next 24 h*

Note

George's rule was developed to deal with maritime depressions off the eastern seaboard of the U.S.A., and care should be exercised when applying it more generally to depressions over the Atlantic. A suggested approach is to restrict the use of the rule to those warm-sector or wave depressions which on general grounds (as in 5.6.1.) are expected to deepen; the rule will then give a quantitative estimate of the amount of deepening.

Ref: George, *Weather forecasting for aeronautics*, New York, 1960.

5.6.3. *Movement of warm-sector depressions*

To a first approximation, movement is given by the speed and direction of the geostrophic wind in the warm sector. A better estimate may be made from the 1000-500 mb thickness field by determining the mean thermal wind within 300 n.miles of the centre and moving the low in the direction of the thickness line over the centre at 4/5ths of the mean thermal speed; this method has a probable error of less than 5 kt in speed and 10° in direction over 24 h. As the thermal field becomes more distorted, the rule becomes less reliable. Rapid deepening is often followed by turning to the left and slowing down, but this is *not* a reliable forecasting rule.

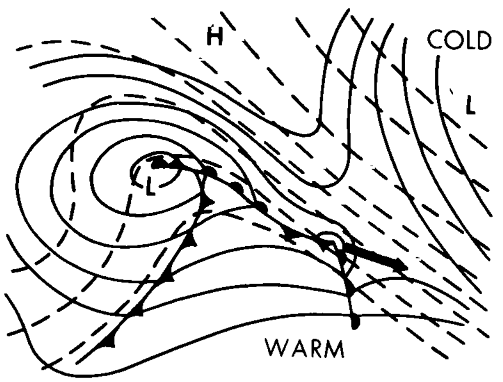
Ref: HWF 6.5.1.1.
Hoyle, *Met.Mag.*, July '55.

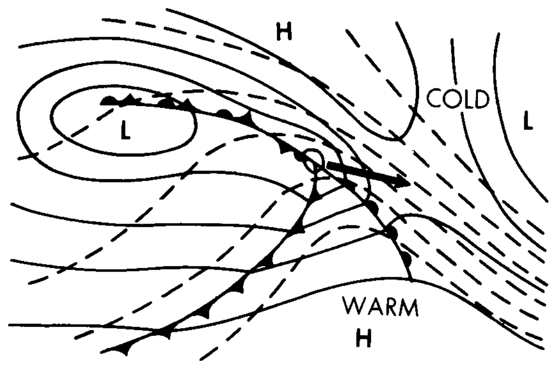
5.6.4. *Movement of occluded and filling depressions*

As occlusion proceeds, the thermal gradient over the centre weakens, and the speed is reduced until the low becomes slow moving or stationary. Although the track normally turns to the left, upstream developments causing a veer of the cold-front jet can turn the centre to the right (see page 80). Once there is little thermal gradient near the centre, movement is normally slow and in the direction of the strongest current round the low.

Ref: HWF 6.5.1.1.

5.6.5. *Warm-front waves and triple-point secondaries on warm occlusions*





—— Isobars   – – – Thickness lines

FIG.5.18. *Characteristic isobaric and thickness patterns for formation of a warm-front wave*

FIG.5.19. *Characteristic isobaric and thickness patterns for formation of a warm-occlusion secondary*

(i) For both types the important feature is a strongly confluent thermal ridge. For triple-point formation, the primary depression must be slow moving, with the strong thermal wind (usually 40 to 80 kt) ahead of the primary.

(ii) Formation usually follows about 24 h after the first appearance of the characteristic thickness patterns.

(iii) Once formed, both types move quickly away from the primary, usually at 30 to 50 kt, and hence are called *breakaway lows*.

(iv) The direction of movement is at about 30° (triple-point low) or 10° (warm-front wave) towards the cold side of the direction of the strong thermal gradient before distortion.

(v) Warm-front wave development may also be assessed from the 300-mb chart; it is likely if the warm front reaches a position near the right entrance of a jet to the E of a 300-mb ridge. Jones found that the direction of movement averages 15° across the 300-mb flow towards the cold side, at 1/3 of the jet speed at 300 mb or 1/2 the 300-mb wind speed along the wave path.

Ref: HWF 5.14.2.1., 5.14.2.3.
Jones, *Met.Mag.*, Oct '62.

5.6.6. *Cold-front waves*

(i) Waves are likely to occur on cold fronts which are at least 1200 n.miles long and are associated with a 1000-500 mb thermal wind of 25 kt or more along this length.

(ii) Waves are likely to form on ana-fronts but not on kata-fronts.

(iii) The wave is likely to develop where the front is subject to orographic distortion, where a bulge on the front can be produced by the flow round adjacent pressure systems, or where the Sutcliffe development term is a maximum (see 5.6.1.). The most probable position for development is 200 to 600 n.miles E of a confluent thermal trough line.

(iv) Both the thermal wind and the surface isobars in the warm air are good guides to the direction of motion; the speed is roughly 4/5 of the geostrophic speed in the warm air (*not* the thermal wind speed).

Ref: HWF 5.14.2.2., 7.2.

5.6.7. *Triple-point secondaries on cold occlusions*

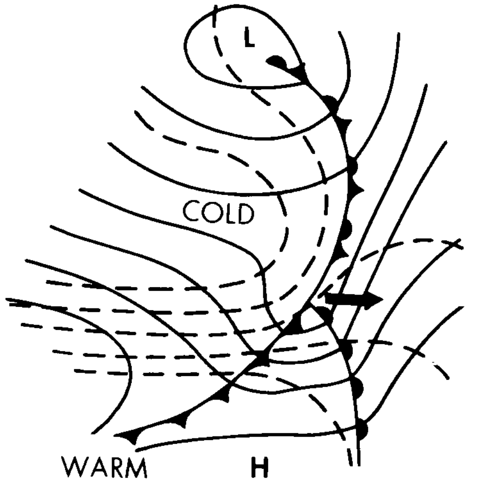


FIG.5.20. *Characteristic isobaric and thickness patterns for formation of a secondary on a cold occlusion*

(i) There must be a marked thermal diffluence ahead of the point of occlusion.

(ii) There is usually a lag of up to 24 h between the first appearance of the pattern and the formation of the secondary.

(iii) Once formed, movement depends a good deal on the details of the thickness pattern around the secondary. If the diffluence is broadly symmetrical, as in Fig.5.20., movement is usually in the direction of the strongest thermal wind and with a speed of 10-20 kt. If, however, the thermal gradient is stronger in the left-hand or the right-hand branch, the direction of movement is often correspondingly towards the left or right and the speed somewhat greater than in the symmetrical phase.

Ref: HWF 5.14.2.4.

5.6.8. *Surface developments linked to jet streams*

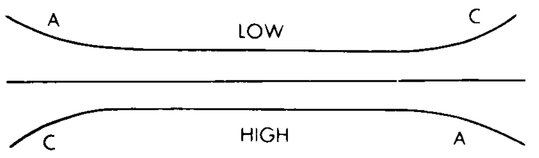


FIG.5.21. *Regions of cyclonic* (C) *and anticyclonic* (A) *development associated with a jet stream*

(i) For cross-sections through jet entrance and exit, see 5.2.8.

(ii) Development areas associated with jets, as in Fig.5.21., are important because of the high wind speed (cf. 5.6.1.).

(iii) Veering of W'ly jets leads to increasing tendency for cut-off low formation in low latitudes (see 5.3.4.). If a depression is associated with the left exit, a veering cold-front jet will cause south-eastward turning of the surface low; this can initiate a major change of type in winter.

(iv) Backing of W'ly jets leads to increasing tendency for cut-off high development in high latitudes (see 5.3.3.).

(v) Left and right exit regions are frequently associated respectively with major low- and high-pressure systems. Eastward propagation of the jet stream gives a good indication of the movement of these surface features.

Ref: HWF Chap.5.

5.6.9. *Formation of new anticyclones*

There are three important anticyclonic development situations (see (i) to (iii)) associated respectively with diffluent thermal ridges, confluent thermal troughs, and a combination of these patterns (see 5.6.1.).

(i) *Amplification of a wave system.*

(a) Anticyclogenesis is initiated by the development of a jet from a S'ly point (see 5.3.3.); building occurs in the right exit region of the jet, ahead of the axis of the diffluent ridge produced to the E of the jet, and often leads to a blocking high. This development normally implies that a deep low has become stationary to the W of the jet.

(b) Winter anticyclones over Scandinavia are normally formed by this process, developing about 600 n.miles to the E of a thermal ridge axis over Iceland. A deep surface low at about 60°N between 30° and 45°W (i.e. some 20°W of the thermal ridge axis) is usually required to produce a well-marked S'ly flow N of 50°N and lying between 20° and 35°W. North-eastward displacement of the Azores high towards the English Channel is also normally required.

Ref: HWF 6.6.2.
Miles, *Scient.Pap.* No.8.

(ii) *Anticyclonic disruption of a trough.*

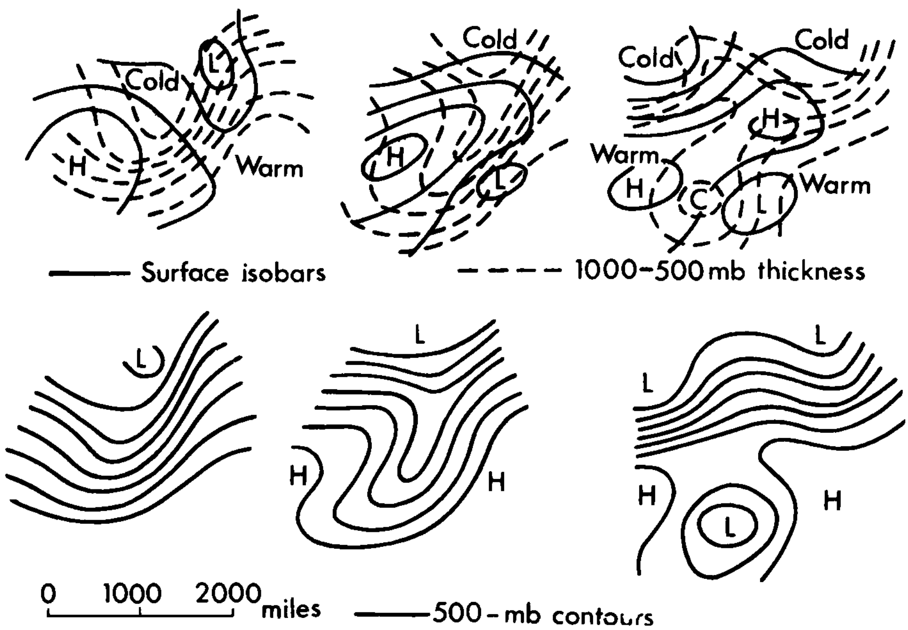


FIG.5.22. *Anticyclonic disruption of a trough*

(a) This development is normally foreshadowed by the forward tilting of the upper trough, the axis becoming aligned NE–SW.

(b) Increasing confluence occurs to the rear of the trough to initiate the anticyclogenesis.

(c) Strong surface pressure rises of about 1 mb per hour extend from the W across the neck of the upper trough, accompanying the development of a marked surface ridge or a new mobile high cell.

(d) A surface low is normally cut off to the S of this ridge on the eastern flank of the upper trough.

Ref: Sutcliffe, MRP 755.
HWF 6.6.2.
Smith, *Met.Rep.* No.21.

(iii) *Combined diffluent ridge and confluent trough pattern.*

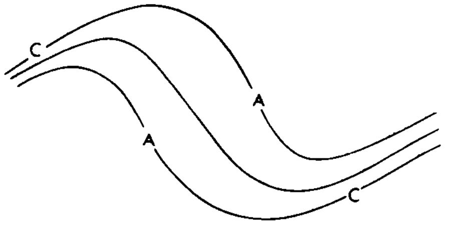


FIG.5.23. *Combined diffluent ridge and confluent trough in thickness pattern*

(a) Anticyclogenesis may be initiated, in the areas marked A, by the development N of 50°N of the type of thermal pattern shown in Fig.5.23., provided that the pattern is self-maintaining (i.e. the advection is such that it keeps the pattern in existence for some time).

(b) Thermal ridge and trough lines should be sharp with marked diffluence and confluence.

(c) The development is most marked when the direction of the thermal jet upwind is S'ly and the thermal winds at AA are W'ly, and least marked when the jets are W'ly and the thickness lines at AA are aligned N–S. There is consequently a decreasing anticyclogenetic effect with time when the jet veers, the surface high then declining slightly or at best only holding its intensity.

Ref: HWF 6.6.2.
Haworth and Houseman, *Met.Mag.*, Nov.'57.

5.6.10. *Movement of anticyclones*

(i) The mobile cold high (see 5.5.2.) is characterized by a strong thickness gradient over the centre (i.e. it is baroclinic) and moves roughly perpendicular to the axis of the thermal trough ahead (*not* parallel to the thickness lines over the centre), with about half the speed of the thermal wind over the centre. As the thickness pattern distorts, the high slows down and normally turns to the right.

(ii) The warm high (see 5.5.2.) is characterized by a weak thickness gradient over the centre (i.e. it is quasi-barotropic) and is slow moving, drifting in a direction between that of the thermal ridge line and a line from the high centre to the point on the axis of the downstream trough at which the thermal wind is strongest.

Ref: HWF 5.15.1., 5.15.2., 5.15.3.

5.6.11. *Decay of anticyclones*

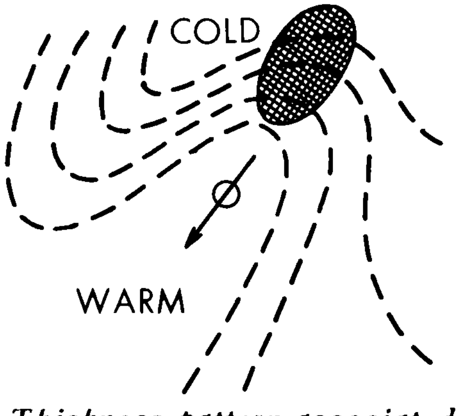


FIG.5.24. *Thickness pattern associated with the decay of an anticyclone*

Decay is usually brought about by an advancing upper trough. In Fig.5.24., the upwind thermal trough is moving towards the surface high centre. The hatched, isallobaric-low area is cyclogenetic, bringing about retreat and/or decline of the high. The high centre usually moves in the direction shown at 15 to 20 kt and the central pressure decreases, sometimes rapidly.

Ref: HWF 5.15.4.

5.7. SURFACE FRONTS

5.7.1. *Analysis – use of hodographs*

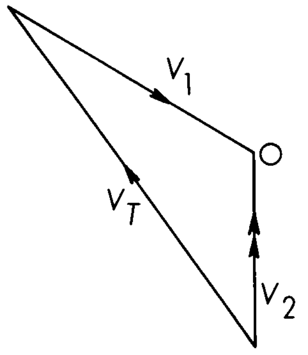


FIG.5.25. *Determination of the thermal wind between two levels*

Plot the actual wind vectors towards the origin, O. The thermal wind for the layer between any two levels is given, in direction and magnitude, by the line joining the tail of the vector for the higher level to the tail of the vector for the lower level. In Fig.5.25., V_1 represents the (NW'ly) wind at the lower level, V_2 the (S'ly) wind at the upper level, and V_T the (SE'ly) thermal wind for the layer.

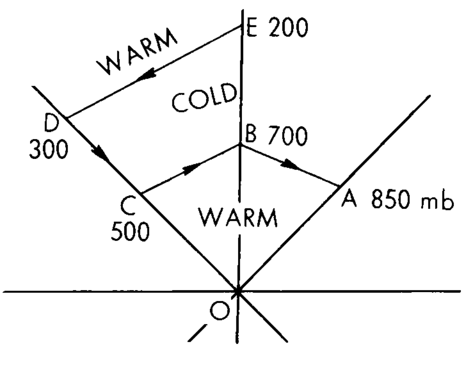


FIG.5.26. *Hodograph to illustrate warm and cold advection*

(i) *Use of hodograph to identify warm and cold advection.* In Fig.5.26., the vector AO represents the 850-mb wind, BO the 700-mb wind, and so on. The thermal wind in the layer 850-700 mb is then represented by the vector BA, the actual wind backing with height. The direction of the thermal wind implies that, in this layer, colder air lies to the NNE of the station and warmer air to the SSW. The mean wind in the layer lies between N and NE and is therefore bringing colder air over the station. Hence, if the wind backs with height cold advection occurs.

Similarly, if the wind veers with height, warm advection occurs (300-200 mb layer in Fig.5.26.); and if the wind direction does not change with height, neither warm nor cold advection occurs (500-300 mb layer in Fig.5.26.).

If cold advection overlies warm advection the air mass becomes more unstable. If warm advection overlies cold advection the air mass is becoming more stable.

(ii) *Use of hodograph to identify type of front and assess frontal speed.*

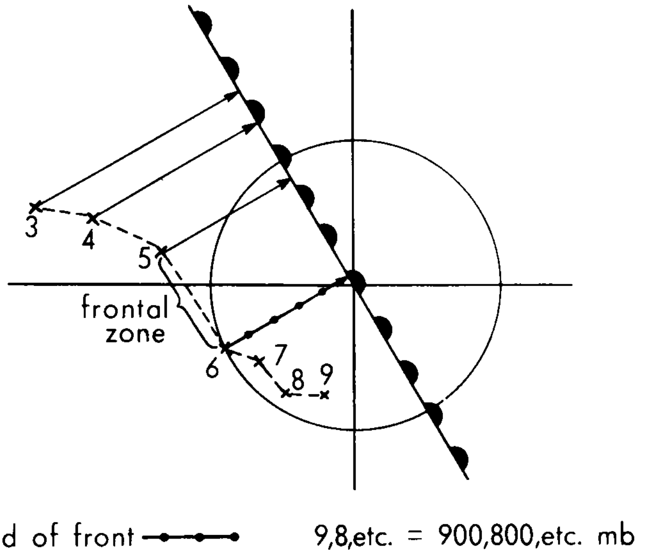


FIG.5.27. *Use of hodograph to identify ana- and kata-fronts and to assess frontal speed*

(a) Mark orientation of surface front on hodograph.

(b) Determine height of frontal surface from tephigram; thermal wind direction in mixing zone is roughly parallel to front.

(c) Measure wind components normal to front at each level.

(d) Frontal speed at surface approximates to speed of cold air normal to front just below mixing zone.

(e) If in the warm air the wind component normal to the front increases with height, then

1. a warm front is an ana-front (as in Fig.5.27.) and
2. a cold front is a kata-front.

If the wind component normal to the front decreases with height, then

1. a warm front is a kata-front and
2. a cold front is an ana-front.

Ref: HWF 3.4.3.

5.7.2. *Cold fronts* (Sansom)

Ana	Kata
Sharp surface frontal trough, with marked drop of wind speed behind front.	Surface wind veer gradual at front and speed changes slight.
Sharp surface temperature fall at front, average over 3 degC.	Only slight and gradual surface temperature fall at front, average less than 1 degC.
Wind backs with height by about 65°.	Wind backs with height only about 20°.
Component of wind normal to front in warm air decreases with height.	Component of wind normal to front in warm air increases with height.
Thermal winds nearly parallel to front.	Thermal winds make angle of 30° or more (veered) across the front.
Slow clearance of upper cloud behind surface front.	Rapid clearance of upper cloud at surface front.

Ref: HWF 7.2., 10.3.
Sansom, *Q.Jnl R.Met.Soc.,* 1951, p.96.

5.7.3. *Speed of movement of warm fronts (and warm occlusions)*

(i) Over land, 2/3 of the component of the geostrophic wind normal to the front.

(ii) Over sea, 5/6 of the component of geostrophic wind normal to the front, or the full value measured 75-100 n.miles ahead of the front.

Ref: Hinkel and Saunders, *Met.Mag.,* Aug.'55.

(iii) Component of wind in the cold air normal to the front measured beneath the frontal surface and above the friction layer, say at 900 mb.

Ref: Matthewman, *Met.Mag.,* Sept.'52.

Notes

(a) Method (iii) implicitly takes into account ageostrophic effects.

(b) Methods (i) and (ii) do not take into account ageostrophic effects in the cold air. Approximate quantitative corrections can be made for curvature and for pressure changes (see 4.3. and 4.2.(iii)).

5.7.4. *Movement of cold fronts (and cold occlusions)*

Full component of geostrophic wind normal to the front.

Notes

(a) Corrections for curvature and pressure changes should be applied (see 4.3. and 4.2.(iii)). This is particularly important when there is a strong isallobaric gradient across the front.

(b) Supergeostrophic motion may occur, notably in summer at the end of hot spells, where the fronts are upper fronts in character, the pseudo-surface boundary extending down to the surface in descending squalls.

Ref: HWF 6.7.3.

CHAPTER 6. PRECIPITATION

6.1. GENERAL PRINCIPLES

In order to produce precipitation, there must be both sufficient ascent and sufficient moisture.

6.1.1. *Dynamical ascent: use of contour/thickness charts*

Areas in which large-scale dynamical ascent (frontal or non-frontal) will occur, may broadly be identified with C (cyclonic) development areas on contour or thickness charts (see 5.6.1. and 5.6.8.).

Dynamical ascent and descent also exert an influence on convective precipitation; if the flow at both lower and upper tropospheric levels shows cyclonic vorticity the chance of showers is enhanced, whereas anticyclonic vorticity is associated with less shower activity. If the air mass is one in which shower development is expected, this will be stimulated in the region immediately ahead of an upper trough, but the passage of the upper trough axis will usually lead to a marked reduction of both the frequency and the intensity of the showers, even when the surface flow is unchanged.

Ref: HWF 16.7.3.

6.1.2. *Moisture: use of tephigrams and 700-mb dew-point depression charts*

Detailed information on the vertical distribution of moisture can only be determined from tephigrams; from these, vertical cross-sections along particular lines can be constructed, and are sometimes useful.

Although the moisture pattern at 700 mb is not necessarily representative of that at nearby levels, 700-mb dew-point depression charts are often useful as an indication of available moisture at medium-cloud levels. Areas of moist air indicated by isopleths on these charts may be moved with the 700-mb winds, but patterns will be modified by vertical motion; areas of moist air are maintained by ascent, but descending air will lead to drying out. It can sometimes be foreseen that 700-mb advection will bring an area of moist air into a region of marked dynamical ascent; this will lead to development or intensification of precipitation.

Precipitation is initiated more readily in air of recent maritime origin than in air of continental origin.

Ref: Morris, *Met.Mag.*, Dec. '66.
Fletcher, N.H., *Physics of rainclouds*, 1962, p. 120.
McCaffery, *Met.Mag.*, Jan. '51.

6.1.3. *Existing precipitation: use of satellite nephanalyses*

Existing areas of precipitation may be located on surface charts and from radar reports, but nephanalyses may also be used to infer the distribution of precipitation (Fig. 6.1.).

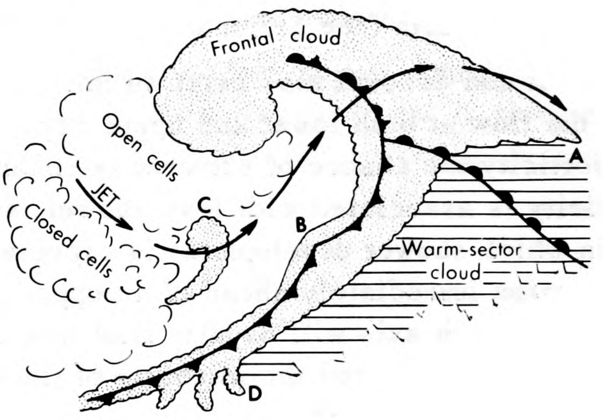


FIG. 6.1. *Schematic representation of significant cloud patterns*

(i) The appearance of a bulge (A) in the frontal cloud towards the warm side usually indicates the propagation of a jet axis across the front. The actual position of the warm front can seldom be located on a nephanalysis, but following this development, precipitation can be expected to be carried well ahead of the surface front. Spiral cloud formations usually indicate positions of depressions.

(ii) A bulge of frontal cloud (B) along the cold front into the cold air indicates the formation of a cold-front wave.

(iii) The formation of a comma-shaped cloud mass (C) in the cold air behind a cold front, usually near the bottom of an upper trough, indicates the development of a secondary maximum of cyclonic vorticity. In this region showers will be merged into an area of continuous precipitation.

(iv) Bands of cumulus (D) in the warm air which approach and merge with a quasi-stationary front indicate that the front is active.

(v) Reports of 'open cells' imply the presence of deep convection and therefore showers. 'Closed cells' imply convection of limited depth and far less shower activity.

Ref: *WMO Tech. Note.* No. 75

6.2. *FRONTAL PRECIPITATION*

6.2.1. *General approach*

(i) Use *thickness pattern* to check surface analysis (amending the latter if necessary), and to identify areas of frontogenesis and frontolysis. Frontogenesis usually leads to active ascent of the warm air and hence precipitation before typical frontal characteristics become apparent on the surface chart.

(ii) Examine *500-or 300-mb contours* in conjunction with surface chart, and compare speeds of movement of corresponding surface and upper air features during the previous 24 h. Fronts likely to become associated with areas of dynamical ascent will intensify (e.g. upper trough catches up surface front, frontal wave reaches left exit or right entrance region of a jet); the converse also holds (e.g. anticyclonic disruption). In assessing the implications of an upper flow pattern, development of the pattern itself must be taken into account as well as advection (e.g. meridional trough extension; buckling of zonal flow leading to jet-stream development).

(iii) *Use tephigrams and 700-mb dew-point depression charts* to check on the maintenance of a supply of moisture. *Wet-bulb potential temperatures* may sometimes reveal useful information; in particular high values indicate large amounts of moisture (e.g. ahead of ex-tropical hurricanes).

(iv) *Use surface charts, nephanalyses and radar reports* to assess latest pattern of precipitation. Developmental changes indicated by (i)-(iii) should be applied.

(v) *Use hodographs* (see 5.7.1.) to infer the relative vertical motion at a front, and so identify those fronts which possess marked ana or kata characteristics.

(vi) Consider *orographic effects*, including the possibility of the release of potential instability. The average rate of frontal rainfall over level country in the British Isles is about 1 to 2 mm/h, but the rate is probably greater by a factor of two to four over high ground. Orographic effects are most marked, leading to large orographic rainfall (e.g. 50 mm in 24 h over Welsh Mountains) when

(a) strong winds are perpendicular to a mountain range,

(b) air is moist in depth,

(c) air is near neutral stability, and

(d) $\frac{\partial^2 V}{\partial z^2} > 0$ up to 500 mb (i.e. wind shear increases with height).

Ref: HWF 10.6., 10.7.
16.5., 16.8.2.

6.2.2. *Warm fronts*

(i) *Amount of precipitation*

Few reliable inferences about frontal activity can be made from the temperature and humidity characteristics of the air masses involved, as correlations of amounts of rain with simple synoptic and stability parameters are low. Current activity is the best guide from which to start, applying modifications to this in accordance with 6.2.1. The following generalizations are however usually dependable:

Assuming continuing ascent	Water content of warm air high to 800 mb or above	Appreciable rain
	Warm air dry above 950 mb	Little but drizzle, and that often near windward coasts only

If the general direction of the lower tropospheric flow in the warm sector is between 180 and 220°, larger amounts of rain are likely than if the 700-mb flow is from 230° or further to the west

If pressure tendencies (especially those in the warm sector) are consistently negative and becoming increasingly so	An increase in the extent, amount and intensity of the rain is likely
If a warm-front wave develops	Marked local intensification of rain occurs, moving along the front away from the centre (see 5.6.5.)

(ii) *Mesoscale variations of rain along a warm front*

Rainfall intensity normally varies along a warm front both in time and space, and commonly reaches a rate of 10 mm/h for short periods. Cells of more intense rainfall may be identified by radar reports, by hourly rainfall charts, or sometimes by the mesoscale tendency field. Such cells appear to have a typical life of 6 to 12h, so that if identified early in their life they can be followed and forecast. The cells usually behave like warm-front waves, moving along the front away from the low centre; alternatively the 700-mb flow often gives a rough indication of movement.

Ref: HWF 10.2., 16.5.
Wallington, *Weather*, June '63.
Jones, *Met. Mag.*, Nov. '66.

6.2.3. *Cold fronts*

(i) Cold ana- and kata-fronts (see 5.7.2.) have markedly different characteristics.

	Cold ana-fronts	Cold kata-fronts
Frequency in British Isles		
Dec. to May (29% unclassified)	54%	17%
June to Nov. (19% unclassified)	8%	73%
Wave development	Likely	Unlikely
Activity	Strong	Weak
Precipitation on front	Fairly heavy rain	Little or none
Precipitation behind front	Light (occasionally moderate) rain extends about 100 n. miles behind surface front, with rain continuing almost to upper cloud edge	Very little rain behind surface front

(ii) *Mesoscale variations of rain along a cold front*

The amount of rainfall along a particular cold front is characteristically extremely variable both in time and in space; rainfall rates of as much as 20 mm/h probably occur quite frequently for periods of a few minutes. Areas of more intense rainfall are sometimes associated with embedded convective cells, which may be detected by radar reports. On cold-ana fronts, local intensification of rainfall is also caused by minor waves which may be identified on nephanalyses or by the mesoscale tendency field. These areas of intensification behave like normal cold-front waves, and move along the front towards the low centre, their movement often being well indicated by the 700-mb flow. Although these flat 'waves' may not develop a circulation they can cause prolonged rainfall.

Ref: HWF 10.3., 16.5.
Sansom, *Q. Jnl R. Met. Soc.*, 1951, p.96.

6.3. *PRECIPITATION AT THE GROUND FROM NON-FRONTAL LAYER CLOUD*

6.3.1. *Northern Ireland investigation* (Mason and Howarth)

(i) *Minimum cloud thickness*

Drizzle	2000 ft
Rain/snow	3300 ft
Continuous rain/snow	7500 ft

(ii) *Critical temperatures at top and base of cloud layer*
Rain (or snow) is unlikely if cloud top is warmer than -12°C. Precipitation of any kind is unlikely if cloud base is colder than -5°C.

(iii) *Relation of precipitation to thickness of cloud and height of base*

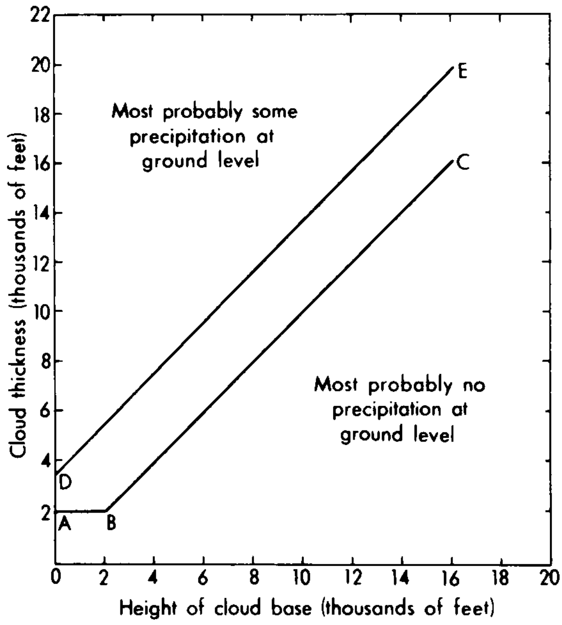


FIG. 6.2. *Effect of cloud thickness and height of cloud base in distinguishing between precipitating and non-precipitating cloud*

Notes

(a) When a point falls between the lines ABC and DE in Fig. 6.2. judgement must be exercised.

(b) The investigation excluded all cases when cumulus cloud or showers were reported; the results must therefore be applied with discretion when a cloud layer is convectively maintained (e.g. over the North Sea in winter).

Ref: Mason and Howarth, *Q. Jnl R. Met. Soc.*, 1952, p.226.

6.3.2. *Met.R.F. investigation over central England* (Stewart)

(i) *Clouds with bases above the 0°C isotherm level*

Height of cloud base above 0°C level	*Cloud thickness*	*Probability of ppn reaching ground*
Less than 3300 ft	More than 6600 ft	90%
More than 3300 ft	All values	5%

(ii) *Clouds with bases below 0°C isotherm level*

Minimum thickness for:	any precipitation	1320 ft
	slight rain	3300 ft
	moderate rain	9900 ft

Note

Based on 8 years of observations.

Ref: Stewart, *Q. Jnl R. Met. Soc.*, 1964, p.287.

6.4. CONVECTIVE PRECIPITATION

6.4.1. *General approach*

(i) *Occurrence*

The likelihood, frequency and intensity of convective precipitation are increased by:

(*a*) Dynamical ascent (see 6.1.1.)
(*b*) Adequate moisture (see 6.1.2., 6.4.7. and 6.4.8.)
(*c*) Appearance of 'open cell' cloud structure (see 6.1.3.)
(*d*) High instability indices (see 6.4.7.)
(*e*) Low cloud-top temperatures (see 6.4.2.)
(*f*) Large depth of convection (see 6.4.3., 6.4.6., 6.4.7. and 6.4.8.)

(g) Small vertical wind shear (see 6.4.4. but also 6.4.8.)
(h) High cloud-base temperatures (see 6.4.8.)

(ii) *Life and direction of movement*

The typical life cycle of a convection cell (i.e. growth, shower production and decay) is completed within about one hour. Individual showers can best be followed by radar, but *extrapolations* should not be made for periods in excess of 1 h; the best guide for the movement of air-mass showers during this period is the 700-mb wind.

Note

These remarks do not apply to severe storms (see 6.4.8.).

6.4.2. *Cloud-top temperatures*

For showers by the Bergeron /Findeisen process:

Cloud-top temperature *degrees Celsius*	Shower probability
0 to −12	Slight possibility
−12 to −40	Likely
Below −40	Almost certain

Ref: HWF 16.7.3.

6.4.3. *Cloud depth*

(i) Showers rarely occur unless the cloud depth as calculated by the parcel method exceeds 150 mb.

(ii) Showers from quite shallow convective cloud (say 5000 ft) are not frequent, but those that do occur are most common near windward coasts in the early or mid-morning; in these cases, larger clouds formed later in the day sometimes fail to produce showers.

Note

Investigation was carried out using data for southern England.

Ref: HWF 16.7.3.1.

6.4.4. *Effect of vertical wind shear*

Only scattered light showers occur if the mean vertical wind shear from the ground to the expected cloud top exceeds 3 kt/1000 ft.

Notes

(a) Applies to air-mass showers only and *not* to those initiated by dynamical uplift.

(b) Does *not* apply to severe local storms for which vertical shear is necessary (6.4.8.)

Ref: HWF 16.7.3.2.

6.4.5. *Effects of local topography*

(i) Areas over, and immediately downwind of southward facing slopes are preferred regions for shower development.

(ii) The ascent of moist air at a sea-breeze front may initiate showers. If two sea-breeze fronts approach one another, the shower likelihood is increased in the region of convergence.

(iii) When a moderate or strong airflow crosses hilly country, standing-wave conditions may develop (see 8.4.3.). Areas of persistent downcurrents may then remain shower-free all day, whilst areas of persistent upcurrents have frequent showers.

Ref: HWF 16.8.2.
Findlater, *Met. Mag.*, Mar. '64.
Hosler, Davis and Booker, *Met. Abstract* No. 212.

6.4.6. *Hail* (excluding soft hail)

(i) None of the known objective techniques for local forecasting of hail has yet been found suitable for use in the U.K.

(ii) An empirical rule suggest that hail is likely when the parcel method indicates cloud tops of at least 15 000 ft, with the path and environment curves at least 5 degC apart throughout most of the cloud depth.

(iii) As hail is often associated with thunderstorms, it is customary to forecast hail whenever thunder is forecast. There is, however, evidence to suggest that damaging large hail is unlikely if there is frequent lightning.

Ref: FTBM No. 1.
Sansom, *Weather*, Sept. '66.

6.4.7. *Thunderstorms*

(i) *Instability indices*

Boyden Index

$$I = Z - T - 200,$$

where Z = 1000–700 mb thickness in decametres
T = 700-mb temperature in °C.

Notes

(*a*) Index isopleths can be advected with 700-mb winds,

(*b*) Threshold value is around 94/95.

Rackliff Index

$$T = \theta_{w900} - T_{500},$$

where θ_{w900} = 900-mb wet-bulb potential temperature in °C,
T_{500} = 500-mb temperature in °C.

Note

Threshold value is around 29/30.

Modified Jefferson Index

$$T_{mj} = 1.6\theta_{w900} - T_{500} - \tfrac{1}{2}T_{d700} - 8,$$

where T_{d700} = 700-mb dew-point in °C.

Notes

(*a*) θ_{w850} may be used instead of θ_{w900} and has little effect on the result provided there is no inversion between 850 and 900 mb.

(*b*) Threshold value is around 26/28.

Ref: Boyden, *Met. Mag.*, July '63
Rackliff, *Met. Mag.*, May '62
Jefferson, *Met. Mag.*, Oct. '63
FTBM No. 14.

The most successful indices for thunderstorm forecasting in the U.K. appear to be

(*a*) Boyden Index – especially for frontal and trough situations, and

(*b*) Rackliff and modified Jefferson – especially for air-mass convection.

Note

Trials suggest that a subjective assessment using the indices gives a better result than the use of indices alone.

(ii) *General considerations*

Convection may be released by surface heating, convergence, or forced ascent. The possibility of release of potential (convective) instability must also be considered, particularly in frontal and trough situations.

(a) Depth of convection:

Cloud tops less than 13 000 ft	Thunder most unlikely
Depth of convection 400 to 500 mb	Thunder probable
Depth of convection over 500 mb	At least 50% of showers accompanied by thunder

When assessing likely depth of convection, subjective allowance must be made for subsidence and for advective temperature changes.

(b) For general thunderstorm development, the air must be moist in depth; dry air aloft implies only isolated storms, but with adequate moisture thunderstorms can readily occur with only slight instability through sufficient depth.

(c) SFLOC reports are most valuable.

Ref: HWF 16.7.5.
16.7.3.1.
FTBM Nos 8 and 14.

6.4.8. *Severe Local Storms*

The term Severe Local Storm has a special connotation; it refers only to an intense convection cloud system which has become self-maintaining. Severe Local Storms normally develop only in the particular circumstances outlined below and, once developed, have a characteristic behaviour which is quite distinct from that of normal convective storms (including those which are described and coded as severe).

(i) *Formation normally requires:*

(a) A supply of warm moist air at low levels, i.e. high values of surface wet-bulb potential temperature (θ_w), typically about 20°C.

(*b*) Great depth of instability.

(*c*) Great buoyancy, indicated by a large excess of θ_w over the saturation wet-bulb potential temperature (θ_s) in the middle and upper troposphere.

(*d*) Vertical wind shear, typically a veer with height throughout the convective layer. The convective layer is the entire troposphere for Severe Local Storms, and shear between the ground and the 500-mb level is usually in the range 30-60 kt.

(*e*) Trigger action, namely day-time surface heating, low-level convergence, or orographic uplift.

Note

These requirements are most likely to be met in the S'ly flow ahead of a major trough. Northward advection, at a height of 3000-7000 ft, of air warmed over Spain can then form a lid to small-scale surface convection such that low-level moisture is confined beneath this lid and high buoyancy can develop. Advective cooling aloft is meanwhile aided by large-scale ascent.

(ii) *Characteristics*

(*a*) Ordinary non-severe showers and thunderstorms usually consist of aggregates of convective cells each of which goes through three stages of evolution (Byers and Braham): (1) the cumulus stage (updraught alone), (2) the mature stage (updraught and downdraught together), and (3) the dissipating stage (downdraught alone). However, when a thunderstorm develops into a Severe Local Storm the ordinary mature stage, instead of lasting for the usual 15 to 30 min, evolves into a 'severe mature' stage that can persist for periods of up to several hours.

Ref: Byers and Braham,
Thunderstorm Rpt,
Washington, 1949.

(*b*) Whereas normal convective systems move roughly with the mid-tropospheric winds, once the severe mature stage is reached movement is some 20° either to the right of and slower than, or to the left of and faster than, the mid-tropospheric winds, i.e. progression is developmental not advective. In the northern hemisphere, Severe Local Storms that travel to the right of the winds (SR storms) are more common than those that travel to the left (SL storms).

(c) Hail, heavy rain and thunder occur, and very occasionally in this country damaging winds or tornadoes too.

(d) The normal sequence of development and the principal properties of the severe mature stage of SR storms, as observed by radar, may be summarized as follows:

Airflow	*3-D distribution of precipitation*
Broad updraught	Large unicellular radar echo (supercell).
↓	
Entrainment unimportant...........	Radar echo top reaches heights predicted by parcel theory. (1-2 km above the topopause in this country).
↓	
Intense updraught	Radar vault on R.H.I.
Rotating updraught	Radar hook echo on P.P.I.
Updraught travels to right of winds	Precipitation echo mainly on storm's left flank.
↓	
Evaporation of precipitation leads to intense downdraught reaching ground in rain area on storm's left flank	
↓	
Updraught and downdraught co-exist stably to give quasi-steady state organization	Quasi-steady 3-D radar echo configuration.

Ref: Browning and Ludlam, *Q. Jnl R. Met. Soc.*, 1962, p.117.
Pedgley, *Weather*, Nov. '65.
Carlson and Ludlam, *Met. Abstract* No. 231.
Browning, *Met. Abstract* No. 233.
Browning, *Jnl Atmos. Sci., Lancaster, Pa.*, 1965, p.664.

6.5. *FORM OF PRECIPITATION – SNOW AND SLEET*

6.5.1. *General approach*

Relevant parameters are:

(i)	Surface dry-bulb temperature	(6.5.2.)
(ii)	Freezing level	(6.5.3.)
(iii)	Wet-bulb freezing level	(6.5.4.)
(iv)	Downward penetration of snow	(6.5.5.)
(v)	1000 – 500 mb thickness	(6.5.6.)
(vi)	1000 – 700 mb thickness	(6.5.7.)
(vii)	1000 – 850 mb thickness	(6.5.8.)

Note

An investigation of the relative merits of parameters (i), (ii), (v), (vi) and (vii) for use in the U.K. showed that (ii) and a modified form of (vii) are the best.

Ref: Boyden, *Met. Mag.*, Dec '64.

6.5.2. *Surface dry-bulb temperature*

Temperature (°C)	1.2	>4.0
Probability that ppn will fall as snow (%)	50	approx. 0

Ref: Murray, *Met. Mag.*, Jan. '52.

6.5.3. *Freezing level*

Height of freezing level above ground (mb)	12	25	35	45	61
Probability that ppn will fall as snow (%)	90	70	50	30	10

Ref: Boyden, *Met. Mag.*, Dec. '64.

6.5.4. *Wet-bulb freezing level*

Height of wet-bulb freezing level	Form of precipitation
feet 3000 or above	Almost always rain; snow rare
2000 – 3000	Mostly rain; snow unlikely
1000 – 2000	Rain can readily turn to snow (see 6.5.5.)
Below 1000	Mostly snow; only light or occasional precipitation falls as rain. Near windward coasts moderate or heavy precipitation may persist as rain

Ref: HWF 16.7.6.1.

6.5.5. *Downward penetration of snow*

The relation between downward penetration of snow (below the 0°C level) and various values of the wet-bulb temperature is given below (See also 1.5.3.).

Type of precipitation	Wet-bulb temperature: To which snow will descend	Wet-bulb temperature: Below which snow is unlikely
	degrees Celsius	
Prolonged frontal precipitation	2.0	2.5
Extensive areas of moderate or heavy instability precipitation	3.0	3.5

Ref: Lumb, *Met. Mag.*, Jan. '63.

6.5.6. *1000 – 500 mb thickness (NW Europe)*

Area	Critical value for equal probability of rain and snow
	decametres
Average	527 (extremes 521 and 546)
Over established snowfields	536
At windward edge of snowfields	528
Over seas with temperature around 10° C, and over windward coasts	523

Ref: Lamb, *Q. Jnl R. Met. Soc.*, 1955, p. 172

6.5.7. *1000 – 700 mb thickness*

1000 – 700 mb thickness	Precipitation
decametres	
> 285	Snow rare
282 – 285	Snow uncommon
278	Rain and snow equally probable; sleet likely
< 276	Rain rare

Ref: Murray, *Met. Mag.* Jan. '52

6.5.8. *1000 – 850 mb thickness*

Boyden recommends adjusting the MSL pressure, p_0 (mb), and station height above MSL, h (gpm), by adding to the 1000 – 850 mb thickness the correction factor

$$\frac{p_0 - 1000}{4} - \frac{h}{30}$$

then:

Probability that ppn will fall as snow (%)	90	70	50	30	10
Modified 1000 – 850 mb thickness (gpm)	1281	1290	1293	1298	1303

Ref: Boyden, Met. Mag., Dec. '64.

6.6 SUMMER DRY SPELLS IN SE ENGLAND

(May to October: three days or more with no measurable rain.)

6.6.1. *Amplifying wave model* (Lowndes)

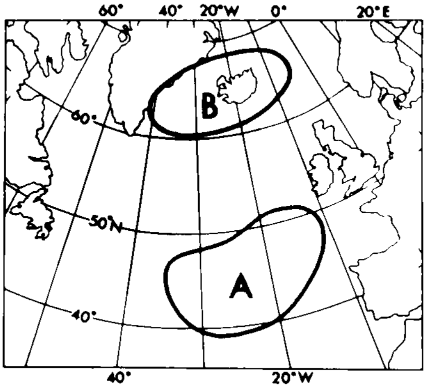


FIG. 6.3. *Significant areas for amplifying wave model (SE England)*

A dry spell will develop within one or two days provided that:

(i) There is a 500-mb trough with axis between 50°W and 60°W, the flow round its base is centred south of 52°N, and the flow ahead of the trough south of 57°N is from a point *not* E of S.

(ii) The next upwind trough is at least 32° longitude further W.

(iii) A surface high of 1024 mb or more is centred within area A in Fig. 6.3. and is *not* elongated in a N'ly direction towards Iceland or Greenland.

(iv) *No* surface high of 1016 mb or more is centred within area B in Fig. 6.3.

(v) The 500-mb height at Lajes minus that at Keflavik is less than 60 decametres.

Ref: Lowndes, *Met. Mag.*, Aug. '65.

6.6.2. *Right exit model* (Ratcliffe)

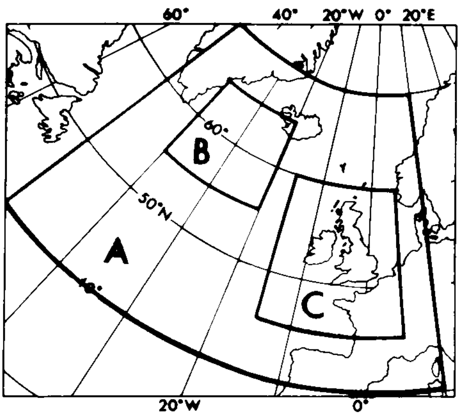


FIG. 6.4. *Significant areas, right exit model (SE England)*

A dry spell is about to begin (or exceptionally may follow in one or two days) provided that:

(i) The strongest 500-mb flow within area A in Fig. 6.4. is between 180° and 270° in direction and passes through some part of area B.

(ii) An equally strong flow between 270° and 310° does *not* exist immediately upstream of the strong flow in area B.

(iii) A 500-mb trough or vortex is *not* in evidence within area C. In applying this proviso, some discretion must be exercised; if a trough lies near the E of area C and is clearly progressive and confluent, this does not rule out a dry spell.

Ref: Ratcliffe, *Met. Mag.*, May '65.

6.6.3. *Blocking high model* (Ratcliffe and Collison)

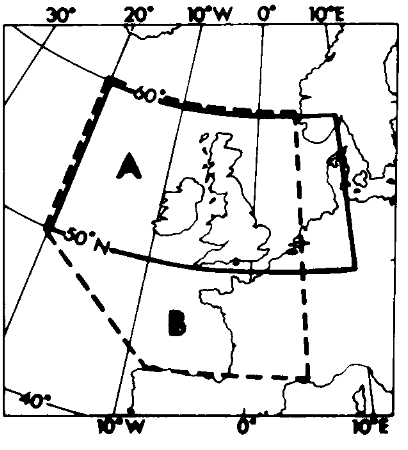


FIG. 6.5. *Significant areas, blocking high model (SE England)*

A dry spell has begun or is expected to begin within 24 h provided that:

(i) A closed contour high at 500 mb of 570 decametres or higher lies within area A.

(ii) A 500-mb trough or vortex is *not* in evidence within area B.

(iii) If the 500-mb wind at Shanwell, Long Kesh or Valentia is between S and W, it does *not* exceed 45 kt at any of these three stations.

Note

In dealing with provis› (ii) some discretion is necessary. Troughs over the southern North Sea and France which are embedded in a N'ly airstream are not important, but troughs over France which link up with an Atlantic trough extending SE towards Biscay are important. A strong W or NW flow behind an Atlantic trough may also vitiate the dry spell even if the trough axis lies just west of 20°W.

Ref: Ratcliffe and Collison, *Met. Mag.*, Aug. '67.

6.7. SUMMER DRY SPELLS IN SW SCOTLAND (Ratcliffe)

(May to October: three days with no measurable rain.)

6.7.1. *Right exit model*

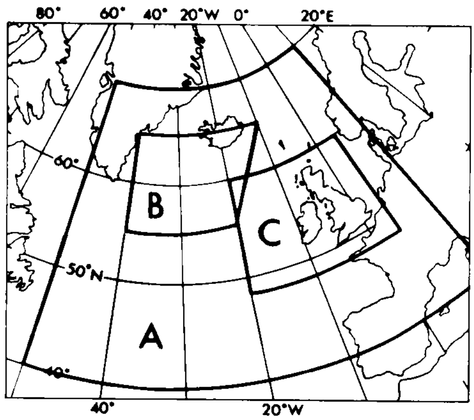


FIG. 6.6. *Significant areas, right exit model (SW Scotland)*

A dry spell has begun or is about to begin provided that:

(i) The strongest 500-mb flow in area A in Fig. 6.6. has a direction between 180° and 240°, passes through area B and does not extend closer to Scotland than the eastern boundary of area B.

(ii) An equally strong 500-mb flow from between 270° and 310° does *not* exist immediately upwind of the strong flow in area B.

(iii) A strong 500-mb ridge lies approximately N to S between 20° and 5°W, with the 570-decametre line reaching at least 55°N and with a wind of less that 30 kt across its axis at all points S of Stornoway and OWS I.

(iv) A 500-mb trough or vortex does *not* exist within area C.

Note

In dealing with proviso (iv), a trough embedded in a NE'ly flow over SE England may be ignored, but *not* a trough in an approximately S'ly flow in the western Channel or S of Ireland.

Ref: Ratcliffe, *Met. Mag.* Apr. '66

6.7.2. *Blocking high model*

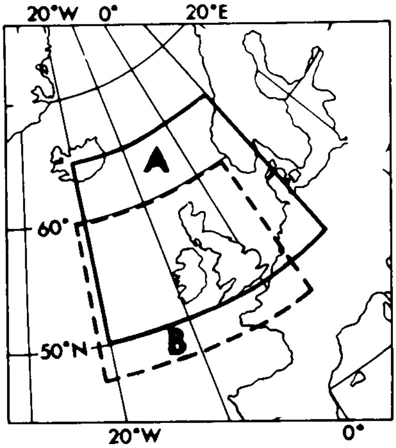


FIG. 6.7. *Significant areas, blocking high model (SW Scotland)*

A dry spell has begun or is about to begin provided that:

(i) A closed contour 500-mb high of 570 decametres or more exists within area A.

(ii) The 500-mb wind must be less than 40 kt within 400 n. miles of the centre and over the British Isles (but not beyond 10°E).

(iii) A 500-mb trough or vortex does *not* exist within area B.

Ref: Ratcliffe, *Met. Mag.*, Apr. '66

6.8. WET SPELLS IN SE ENGLAND

(Three consecutive days on which at least 6 mm of rain fell, with at least 1 mm on each day.)

6.8.1. *Left exit model* (Ratcliffe and Parker)

A wet spell has begun or is about to begin (exceptionally does not commence for 24 h) provided that:

(i) London lies near, or slightly to the left of the core of a long and fairly straight (or slightly cyclonically curved) jet, or downwind from the exit region if the jet is expected to propagate forwards along its path.

(ii) The jet is (normally) the strongest flow in the area bounded by 40°N 70°N 10°E 60°W and is about 1000 n. miles in length, with the greater part W of 10°W.

(iii) There is *no* 500-mb ridge over the British Isles, and *no* closed 500-mb vortex over Iberia.

Note

Particular care must be exercised when a 500-mb trough exists between 50°W and 60°W (cf. 6.6.1.) Provided such a trough is *not* well developed, and a further trough exists within 40° longitude upwind, the wet spell may be forecast when the three criteria above are satisfied; but if the trough at 50°W to 60°W is well developed, a wet spell should be forecast only if there is *no* indication of backing upper flow over the E Atlantic which could lead to ridge development over Britain.

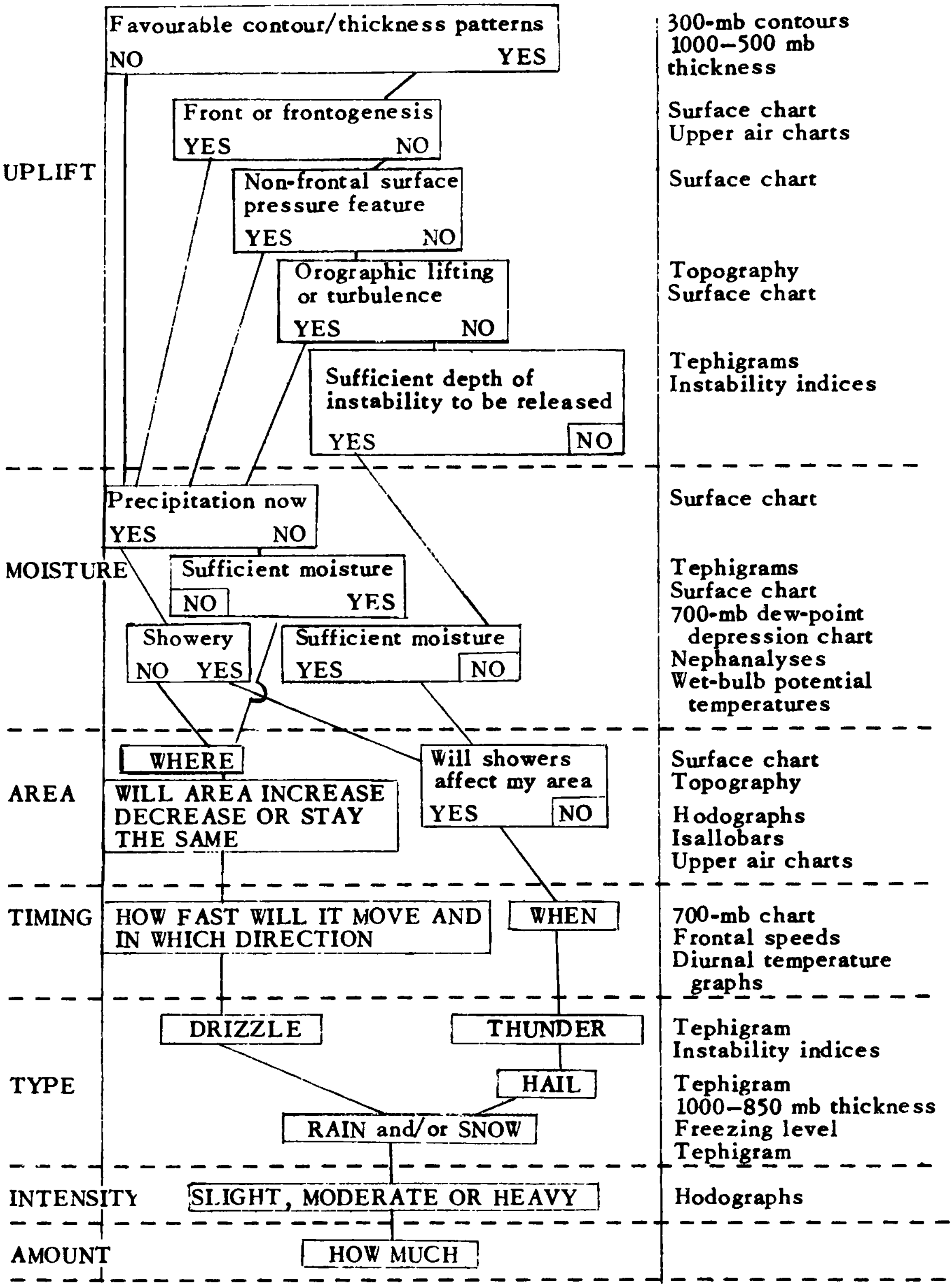
Ref: Ratcliffe and Parker, *Met. Mag.*, Jan.'68.

6.8.2. *Trough model* (Lowndes)

The model is based on the likelihood that wet spells occur on the forward side of upper troughs. Criteria, however, vary from season to season. Full details are given in the original article.

Ref: Lowndes, *Met. Mag.*, Oct. '62.

6.9. RAPID REFERENCE GUIDE TO FORECASTING OF PRECIPITATION



Ref: Met.O.TS. modification from FTBM No. 1

CHAPTER 7. TROPOPAUSE AND STRATOSPHERE

7.1. DEFINITION AND STRUCTURE OF THE TROPOPAUSE

7.1.1. *WMO definition*

(i) The first tropopause is defined as the lowest level at which the lapse rate decreases to 2 degC/km or less, provided also that the average lapse rate between this level and all higher levels within 2 km does not exceed 2 degC/km.

(ii) If above the first tropopause the average lapse rate between any level and all higher levels within 1 km exceeds 3 degC/km, then a second tropopause is defined by the same criterion as under (i). This tropopause may be either within or above the 1-km layer.

Note

A level otherwise satisfying the tropopause definition but occurring at a height below that of the 500-mb level will not be designated a tropopause unless it is the only level satisfying the definition and the lapse rate fails to exceed 3 degC/km over at least 1 km in any higher layer.

Ref: WMO EC – IX.

7.1.2. *Tropopause model*

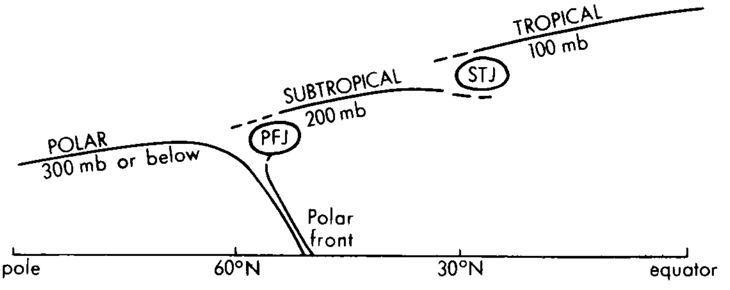


FIG. 7.1. *Schematic changes in height of tropopause from pole to equator*

PFJ Polar-front jet stream STJ Subtropical jet stream

Notes

(a) Positions and heights indicated are rough average values only.

(b) Tropopause 'breaks' occur in association with jet streams.

(c) In addition to the two breaks shown in the model, further breaks are sometimes found in both polar and subtropical tropopauses, e.g. the arctic front may disrupt the polar tropopause in winter.

Ref: Reiter, *Jet-stream meteorology*, University of Chicago Press, p.97.

7.2. TROPOPAUSE IDENTIFICATION

7.2.1. *Tropopause pressure/temperature relationship*

Strict application at radiosonde stations of the WMO definition in 7.1.1. will on occasions give misleading reports which, in particular, could omit the lowest tropopause. When it is suspected that this has happened the following table might assist identification on a tephigram.

TABLE 7.I. *Average pressure and temperature at the tropopause*

Month	Tropopause pressure (mb)					
	150	200	250	300	350	400
			degrees Celsius			
January	−72	−67	−60	−56	−53	(−50)
April	(−67)	−64	−59	−53	−48	(−44)
July	−63	−58	−53	−48	−44	–
October	−66	−61	−56	−52	−48	(−42)

Notes
(*a*) Values in brackets are based on fewer than 20 observations.
(*b*) Analysis is based on data from Stornoway 1957-66 and Crawley 1957-64.
(*c*) Standard deviations of individual values about the above means are about 2–4 degC.

7.2.2. *Time changes of tropopause potential temperature and pressure*

(i) The tropopause is found to retain its potential temperature within the limits of measurement over 24 h. In areas well away from tropopause breaks, time changes of tropopause height at a point rarely exceed 2 mb/h.

(ii) The potential temperature of the tropopause will usually be discontinuous at a break, and time changes of tropopause level at a point are often about 10-12 mb/h.

Ref: Sawyer, *Geophys Mem.* 92.
Kantor, AFCRL, Envir. Res. Pap. No.41, Bedford, Mass., 1964.

7.3. SYNOPTIC AIDS TO THE CONSTRUCTION OF TROPOPAUSE CHARTS

(i) Tropopause funnels (lows) are often of relatively small horizontal dimensions, and their presence is therefore not necessarily indicated directly by the observations on a particular chart. Continuity usually gives a good guide to the position of all tropopause centres.

(ii) An independent check on the positions of tropopause centres may be obtained from the computed thermal winds at stations surrounding the centres. Tropopause funnels extending below the 300-mb level are normally aligned vertically with warm centres in the 300-200 mb thickness pattern, and tropopause domes (highs) extending above the 300-mb level are aligned with warm pools in the 500-300 mb thickness pattern. Lines drawn through the positions of the radiosonde stations perpendicular to the thermal winds for the appropriate layer will therefore intersect near the position of the tropopause centre.

(iii) Tropopause funnel development (due to the descent and adiabatic warming of stratospheric air) will accompany strong divergence at the maximum wind level; this in its turn is normally associated with vigorous deepening of surface depressions.

(iv) Tropopause isopleth gradients are uneven, and show bands of strong gradient along tropopause breaks; thus, typically there will be two such bands on all hemispherical charts, but on occasions the two bands can merge into one. The best positional guide for these bands of strong gradient will therefore be the jet-core positions as determined on a maximum wind chart (or as roughly indicated by a 300-mb chart).

(v) If the thickness field indicates that little vertical motion is taking place near a front at high level, the advective component of motion will be dominant, and tropopause isopleths will move as if tied to the front.

Ref: Ogden (unpublished).

7.4. THE STRATOSPHERE

7.4.1. *ICAO standard atmosphere (ISA)*

Layer	Altitude	Lapse rate
	gpkm	*degC/gpkm*
Troposphere	0-11	6.5
Lower stratosphere	11-20	Isothermal
Middle stratosphere	20-32	-1.0

For full ICAO standard atmosphere, see Appendix I.

Ref: ICAO Doc. 7488/2, *Manual of the ICAO standard atmosphere*, 1964.

7.5. THE LOWER AND MIDDLE STRATOSPHERE OVER THE BRITISH ISLES IN SUMMER

The overall picture of the stratosphere in summer (June to July) is that of a quiet region where little change occurs from day to day and from year to year. Table 7.II. gives the winds and temperatures from 500 to 10 mb at Crawley in July 1967 and may be taken as typical of summer months from 50 mb upwards. Tropospheric data are included for comparison purposes.

TABLE 7.II. *Winds and temperatures from 500 to 10 mb, Crawley, July 1967*

Pressure level	WIND					TEMPERATURE				
	Vector mean	SVD	RMS vector changes			Mean	SD	RMS changes		
			24 h	48 h	72 h			24 h	48 h	72 h
mb	*deg/kt*	*knots*				*degrees Celsius*				
500	253 22	18.8	21.9	27.4	27.9	-14.9	2.3	2.5	2.6	3.0
300	251 32	31.1	35.5	43.7	48.5	-42.1	2.8	2.6	3.3	3.8
200	253 33	28.6	28.3	36.9	41.8	-54.2	4.0	5.3	5.7	5.6
100	249 17	12.0	10.6	13.6	15.1	-53.9	2.5	2.4	3.0	3.3
50	123 05	5.5	5.8	6.9	7.1	-50.2	1.5	1.9	1.5	1.9
30	103 14	6.3	6.7	7.8	7.0	-45.2	2.2	3.2	2.7	3.1
20	103 19	7.8	9.9	11.3	9.7	-40.1	2.3	3.0	3.2	3.2
10	105 23	5.4	8.6	7.5	6.6	-29.1	3.0	-	-	-

The main features of the stratospheric régime when compared with the tropospheric régime are as follows.

WIND

Troposphere	Stratosphere
Westerlies normally increase with height to the level of maximum wind (LMW) in the high troposphere.	Westerlies decrease with height and reverse to easterlies between 100 and 50 mb, then increase slowly with height.
Maximum variability, i.e. high standard vector deviation (SVD), occurs at LMW.	SVD decreases with height in the low stratosphere. From 50 mb upwards values are not only small, at 5-8 kt, but smaller than root mean square vector (RMSV) differences over at least 1-3 days; representative climatological means, or the means of several recent values, will therefore give forecasts with smaller standard errors than will persistence.
RMSV changes reach a maximum at LMW. They increase with time up to 2-5 days and can be as large as 70 kt in some years. Out-of-date reports therefore have little current value.	RMSV differences decrease with height in the low stratosphere and from 50 mb upwards rarely exceed 10-12 kt. There is no signficant increase in RMSV changes in periods of up to a week, so that out-of-date reports may have considerable current value. From 50 mb upwards, changes in the vector mean from year to year are small. Values for a particular year are therefore useful.

TEMPERATURE

Troposphere	Stratosphere
Normally decreases with height.	Usually a small increase with height even in the lower stratosphere (cf. ICAO standard atmosphere).
The standard deviation (SD) changes little with height except for a limited region near LMW where it is a minimum.	Maximum values of SD are found in the low stratosphere above which they decrease to minimum values between 50 and 30 mb. At higher levels SD shows an apparent increase with height but this may be due, in part, to increasing errors of observation with height. As SD is smaller than RMS changes, means

Troposphere	Stratosphere
	give better forecasts than does persistence. There is some evidence to suggest that the 10-mb mean in Table 7.II. is 6–8 degC too warm.
	RMS changes with time from 50 mb upwards show no significant increase for periods up to one week. Out-of-date reports may therefore be of value.

Ref: Hawson (unpublished).

7.6. THE LOWER AND MIDDLE STRATOSPHERE IN WINTER

7.6.1. *Major disturbances in the stratosphere*

Stratospheric wind and temperature patterns during the winter are very much affected by occasional stratospheric disturbances. These are very different from the familiar tropospheric disturbances and the following distinguishing characteristics may be noted.

(i) Vigorous stratospheric disturbances are so large that they can influence directly half a hemisphere and indirectly the other half of the hemisphere.

(ii) Their effect can be so marked that the 'normal' winter thermal pattern is completely reversed, with the warmest air located over the pole (i.e. a summer pattern but with stronger gradients). A disturbance that reverses the normal cold pole régime to produce a relatively warm pole is classified by some workers as a *major disturbance*. Whereas in high latitudes the temperature at 10 mb is commonly –70°C to –80°C in undisturbed régimes, temperatures in excess of 0°C may occur in major disturbances. Thus, surprisingly, at high levels in high latitudes extreme high and low temperatures both occur in January. *STRATWARM ALERTS* are broadcast whenever the temperature at 10 mb exceeds –30°C (roughly ISA + 15 degC.

(iii) Movement of the systems is relatively slow, and their life cycle typically extends over 4 weeks or so; a disturbance may thus completely dominate the wind and temperature patterns for a month or more.

(iv) Whereas minor disturbances occur every winter, major disturbances do not affect the middle and lower stratosphere every year. Thus in the 17 winters from 1951/52 to 1967/68, there have been six major disturbances, starting in January of 1952, 1957, 1958, 1963 and December of 1965 and 1967. (There is some evidence that major disturbances occur more frequently in the upper stratosphere, from which level only a proportion extend their influence downwards to the middle and lower stratosphere.)

(v) Variations of temperature following the flow can be quite spectacular, much more so than time variations at a point. The high temperatures in an area of stratospheric warming are largely caused dynamically by descent through the pressure surfaces, and the areas of warmest air correspond to the lowest point of the trajectories. The centres of maximum temperature at each level are not aligned vertically; slopes of the axes joining warm centres at progressively higher levels of about 1:200 have been measured between 50 and 10 mb.

Ref: Hawson (unpublished).

7.6.2. *Mean values of wind and temperature over the British Isles*

As a result of the characteristics of stratospheric disturbances noted in 7.6.1. the frequency distributions are abnormal; values of both wind and temperature meaned over a number of winter seasons are therefore of little help to the analyst. Similarly, mean values for a particular winter may give little guidance about the values in any one month. Better guidance to winter stratospheric conditions is given by values of means, SD, and RMS changes for particular months from single years.

The figures for January 1967 given in Table 7.III. may be taken as representative of the winter stratosphere during a year in which no major disturbance occurs. The values of wind and temperature to be expected during a stratospheric disturbance will clearly depend on its position relative to the British Isles. The figures for January 1968 (during which a major disturbance was centred near Greenland), Table 7.IV., are typical only of conditions on the SE flank of a disturbance, some way from the centre; they do, however, indicate the possible departures from conditions in an undisturbed régime. In the development stage of a disturbance, W'ly winds of 150 kt have been recorded at 50 mb on the northern flank of the warm thermal centre, compared with the undisturbed normal of about 40 kt.

7.6.3. *Data for an undisturbed winter, January 1967 at Crawley*

Table 7.III. may be taken as typical of an undisturbed winter régime.

TABLE 7.III. *Winds and temperatures from 100 to 10 mb, Crawley, January 1967*

Pressure level	WIND					TEMPERATURE				
	Vector mean	SVD	RMS vector changes			Mean	SD	RMS changes		
			24 h	48 h	72 h			24 h	48 h	72 h
mb	*deg/kt*	*knots*				*degrees Celsius*				
100	295 26	27.2	16.6	20.0	25.7	-59.1	4.0	2.6	3.4	4.4
50	285 28	26.1	15.4	20.0	20.5	-60.8	4.0	3.2	4.2	4.6
30	276 36	29.3	14.4	19.4	21.1	-59.3	4.9	3.8	5.2	6.7
20	272 55	35.3	17.2	21.9	28.2	-57.0	6.9	4.7	6.6	8.5
10	271 88	48.4	19.1	27.7	46.5	-48.0	8.7	4.5	4.7	7.6

The main features compared with those of the troposphere are as follows.

WIND

Troposphere	Stratosphere
Predominantly W'ly	Also predominantly W'ly
Speed and SVD both increase to maxima at LMW	Speed and SVD both decrease to rather flat minima between 100 and 50 mb, then increase steadily, and are greater than RMSV changes. From 100 mb upwards climatological-type forecasts yield standard errors larger than those resulting from forecasts based on persistence for 24 h.
RMSV changes reach a maximum at LMW, and increase with time for about 5 days, reaching values of about 60-70 kt.	RMSV changes decrease with height to the minimum wind level, but then increase steadily upwards. Throughout the stratosphere, RMSV changes with time increase for a week or more and reach values exceeding 70 kt at 10 mb over 7 days.

TEMPERATURE

Troposphere	Stratosphere
Normally decreases with height.	Normally also decreases with height (but at a much reduced rate) to a rather flat minimum at

TEMPERATURE (contd)

Troposphere	Stratosphere
	about 30 mb. Above this level, temperature normally increases with height.
SD changes little with height except for a limited region near LMW where it is a minimum.	SD and RMS changes increase sharply to maxima in the low stratosphere at about 200 mb, then decrease to rather flat minima between 100 mb and 50 mb. Above this level, variability increases again to a maximum probably at about 5 mb where values are greater than at 200 mb. Persistence 24-h forecasts are better than those based on means.

7.6.4. *Data for a disturbed winter, January 1968 at Crawley*

Table 7.IV. refers to one month in a disturbed winter, but it is not necessarily typical of a disturbed régime. In this case the disturbance was centred near Greenland – hence the N'ly winds, in sharp contrast to those of January 1967 (see 7.6.3.).

TABLE 7.IV. *Winds and temperatures from 500 to 10 mb, Crawley, January 1968 – a disturbed winter*

Pressure level	WIND					TEMPERATURE				
	Vector mean	SVD	RMS vector changes			Mean	SD	RMS changes		
			24 h	48 h	72 h			24 h	48 h	72 h
mb	*deg/kt*	*knots*				*degrees Celsius*				
500	315 40	34.7	36.2	45.5	50.5	-23.4	4.1	4.[illegible]	5.8	5.4
300	322 58	48.5	52.7	67.8	68.4	-49.9	2.2	2.5	2.7	3.1
200	320 56	43.8	48.7	61.2	64.4	-61.4	6.7	6.4	8.2	8.3
100	319 27	23.7	18.6	25.0	30.9	-58.1	5.1	4.9	5.7	5.8
50	345 18	20.1	13.5	16.5	22.7	-56.7	5.2	3.7	5.4	6.2
30	352 22	30.2	16.2	26.8	33.6	-54.2	7.3	4.5	6.0	6.8
20	006 20	36.1	23.8	35.3	45.4	-51.1	9.6	4.8	7.6	8.8
10	013 20	38.3	27.9	42.1	46.0	-46.9	6.9	4.2	3.4	3.5

Note

RMSV wind changes increase throughout 6 or 7 days; this effect is typical of large, slow-moving and long-lived systems.

Ref: Hawson (unpublished).

7.7. THE STRATOSPHERE DURING THE TRANSITION SEASONS

7.7.1. *The change from summer to winter*

This is a gradual process caused by seasonal changes in the radiation balance. Changes first take place over the pole at high level, and then extend southwards and downwards. Changes in wind patterns result from the thermal changes.

Approximate dates	Thermal pattern over the British Isles	Wind pattern over the British Isles
Up to mid-August	←——— Summer	patterns ———→
By late August	Temperatures increase northwards below 10 mb. Pattern reversed above 10 mb.	E'ly in the middle stratosphere.
September	Generally weak patterns. Belt of relatively warm air in mid-latitudes with colder air to N and S.	Tropospheric pattern decreased but not reversed in the low stratosphere, then little change with height from 50 mb to 10 mb, i.e. stratospheric patterns are weak forms of tropospheric patterns.
By mid-October	Temperatures increase northwards below 50 mb. Pattern reversed above 50 mb.	W'ly in the middle stratosphere.
From early November	←——— Winter	patterns ———→

Note

The dates are only a rough guide; the change to winter patterns is brought forward if there is strong tropospheric W'ly flow which can dominate the lower stratosphere and effectively join (in the vertical) with the stratospheric W'ly flow which is developing downwards.

7.7.2. *The change from winter to summer*

Over the British Isles persistent E'ly wind components at 50 mb have first been observed as early as mid-March and as late as late May, whilst in 4 years out of 12 (1957-68) the first appearance occurred during the second half of May. In any one year the first appearance of the E'ly wind components was earlier the higher the level considered.

After a major disturbance which causes a warm pole to develop dynamically, the situation will generally revert to the normal cold-pole régime; in such years there is a tendency for the transition to be late. But if the disturbance forms in late winter (say, late January onwards) its life cycle will last through February into March. If the warm pole is still present giving stratospheric easterlies in mid-March, the cold pole will not re-form and the summer pattern is effectively established already.

CHAPTER 8. BUMPINESS IN AIRCRAFT

8.1. SOURCES AND SCALES OF TURBULENCE

The principal sources of turbulence are thermal, dynamical and orographic, acting separately or in combination. Turbulence of significance to the operation of aircraft depends to some extent on the aircraft size, weight and speed; eddy sizes of a few feet to a few hundreds of feet are usually experienced as high-frequency bumps whilst larger eddies, up to several thousands of feet, are experienced as heaving motion. Intermediate eddies give rough or jolting flight.

The ICAO 6th Air Navigation Conference (Montreal 1969) recommended that the following classification be used as a guide to the reporting of moderate and severe turbulence. (Light turbulence may be reported when the effects are less than those due to moderate turbulence, extreme turbulence when effects are more pronounced than those of severe turbulence.)

Description of turbulence	Effect in aircraft
Moderate	There may be moderate changes in aircraft attitude and/or altitude but the aircraft remains in positive control at all times. Usually, small variations in airspeed. Changes in accelerometer readings of 0.5 g to 1.0 g at aircraft's centre of gravity. Difficulty in walking. Occupants feel strain against seat belts. Loose objects move about.
Severe	Abrupt changes in aircraft attitude and/or altitude; aircraft may be out of control for short periods. Usually, large variations in airspeed. Changes in accelerometer readings greater than 1.0 g at the aircraft's centre of gravity. Occupants are forced violently against seat belts. Loose objects tossed about.

Ref: ICAO Doc.8812, AN-CONF/6, 1969.

8.2. THERMAL TURBULENCE

8.2.1. *Occurrence*

At the edges of upcurrents and downcurrents in a convective régime, as follows:

(i) In cloud.

(ii) Outside Cb, particularly in clear air around anvil and just above storm top.

(iii) In dry thermals below cloud base, or in a cloudless atmosphere over any heated land mass. Over deserts, dry convection may reach 15 000 - 20 000 ft.

(iv) In downdraughts appearing with the onset of precipitation; these spread out at the surface to produce a line-squall close to the shower area (see 4.7.).

Thermal turbulence can reach heights exceeding 40 000 ft in the U.K. and 60 000 ft in the U.S.A. and some tropical areas.

Recent evidence suggests that severe turbulence is unlikely to be encountered more than about 15 n.miles from the edges of storm radar echoes.

Ref: Burns, Harrold, Burnham and Spavins, *ESSA Tech. Mem.* 1 ERTM-NSSL 30, 1960.

8.2.2. *Typical vertical currents* (based on aircraft reports)

Régime	Vertical velocity m/s	Degree of turbulence
Cu hum, Cu med	1-3	Light
Cu cong	3-10	Moderate
Cb	10-25	Severe
Severe storms (U.S.A.)	20-100	Extreme
Dry thermals	1-5	Light to moderate
Downdraughts	3-15	Up to severe
Downdraughts (U.S.A.)	Up to 40	Extreme

8.2.3. *Forecasting intensity of turbulence in cloud from the tephigram* (George)

ΔT	Cloud	Turbulence
degC		
0-3	Sc, Cu hum, Cu med	Light
4-6 7-9	Cu cong, Cb (air-mass type)	Moderate Severe
7-9	Cb (frontal or line-squall type)	Extreme

ΔT = Maximum temperature difference up to 400 mb between parcel and environment curves.

Note

Investigation was carried out in U.S.A. but results are probably applicable elsewhere in temperate latitudes.

Ref: George, *Weather forecasting for aeronautics*, New York, 1960, p.433.

8.3. DYNAMICAL TURBULENCE

Turbulence produced by strong vertical wind shear generally occurs between 500 mb and 200 mb in layers a few hundred feet thick and of mesoscale dimensions, but occasionally may be much deeper and more widespread. This type of turbulence usually occurs in clear air (CAT) and is normally light or moderate. Severe CAT may be encountered once in several hundred flying hours.

8.3.1. *Association with synoptic features*

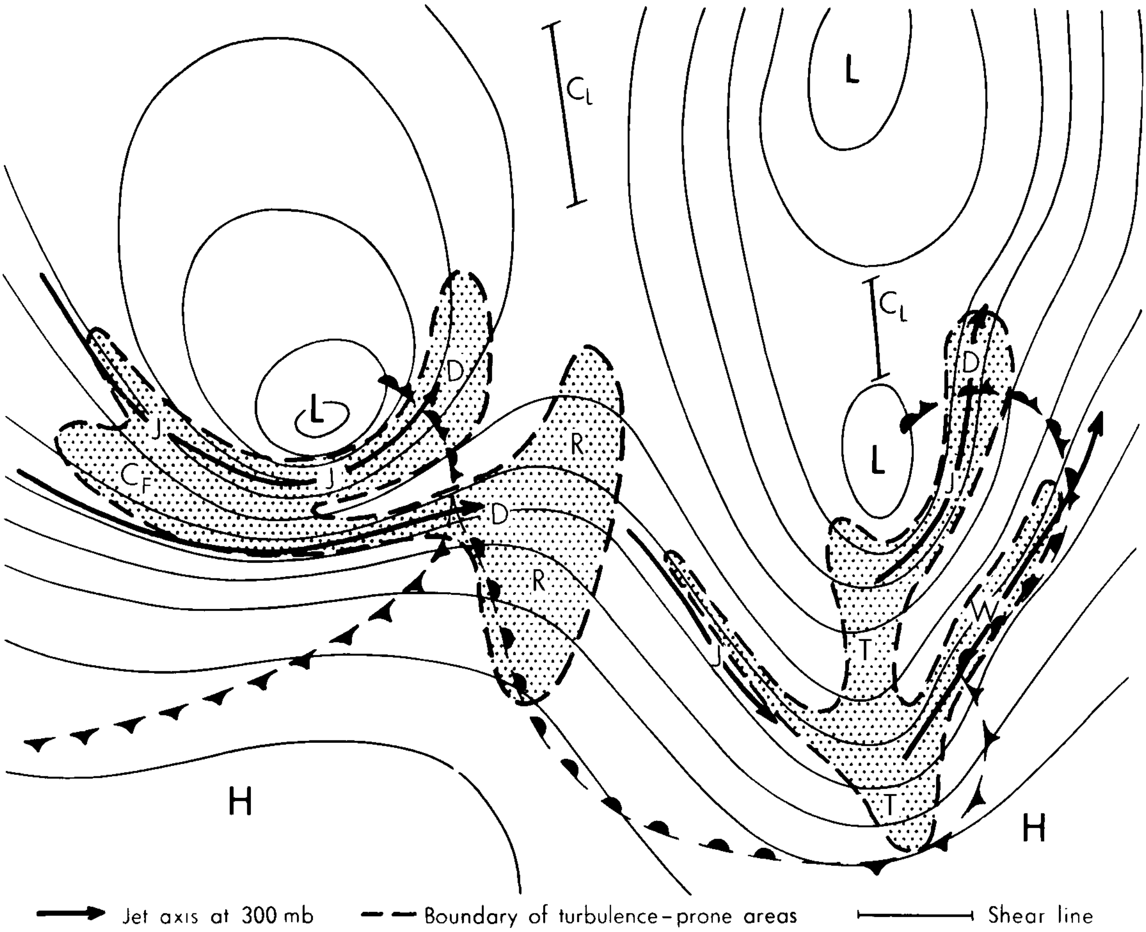


FIG.8.1. *Main turbulence-prone areas between 500 and 200 mb as related to features of the 300-mb chart*

—— 300-mb contours. Fronts marked are at surface.
C_F – Region of confluence between two jet streams.
C_L – Upper air col. Turbulence occurs in narrow bands along marked shear line.
D – Diffluent region of jet stream.
J – Jet-stream turbulence on low-pressure side.
R – Developing upper ridge.
T – Sharp upper trough.
W – Developing wave depression.

(i) Turbulence within these areas is most likely to occur near the tropopause or other stable layers.

(ii) In jet streams, turbulence is generally found in the baroclinic zones above and below the core. There is some tendency for turbulence to be concentrated in the lower zone of cyclonically curved jets and in the upper zone of anticyclonically curved jets.

Ref: HWF 20.2.5.

(iii) There is some qualitative evidence that turbulence is more likely to be severe within turbulence-prone areas if rapid change or development in the upper air pattern is occurring locally than if the pattern is relatively static.

Ref: Roach, *Met.Mag.*, Mar. '69.
Lennie, *Met.Mag.*, Jan. '69.

8.3.2. *Effect of topography*

The incidence of CAT is more frequent over land, particularly over mountains, than over the sea. It is not possible to give precise quantitative values to this difference but it is suggested that a factor of 3 or more over most land areas, increasing perhaps to 10 over very mountainous areas, is reasonable.

Ref: Turner, *Met.Mag.*, Feb. '59.
Clodman and Ball, New York Univ., Coll. Eng. Res. Div., AFCRL Contract No. AF 19(604)-3068, 1959.
Colson, *WMO Tech. Note* 95, 1969.

8.4. OROGRAPHIC TURBULENCE

8.4.1. *Low-level turbulence*

The following table gives a rough guide to the subjective intensities of turbulence to be expected in the lowest few hundred feet.

Surface wind	Turbulence over:		
	Sea	Flat country	Hill country
knots 15-35	Light—moderate	Moderate	Severe
>35	Moderate—severe	Severe	Extreme

Note

Strong sunshine added to strong wind may increase handling difficulties with aircraft on landing and take-off.

Ref: Zbrozek, *RAE Tech. Note AERO* 2681, 1960.
Pasquill (unpublished).

8.4.2. *Squalls near the surface*

Violent squalls create the most dangerous low-level hazard to aircraft, and may arise as follows:

(i) During or preceding the passage of an active cold front.

(ii) During or preceding the passage of a thunderstorm. (see 4.7.)

(iii) In hilly or mountainous country in association with 'rotor streaming', which requires:

(a) Strong winds at low levels (20-30 kt or more).

(b) Winds decreasing sharply or even reversing direction at 1½ to 2 times the height of the hills.

(c) A stable air mass (apart from stirring in the lowest layers).

Ref: *Met. Rep.* No. 18.

8.4.3. *Mountain waves*

Mountain waves and associated turbulence may be pronounced when the following conditions are all satisfied:

(i) A wind blowing within 30° of normal to a substantial range of hills.

(ii) A wind speed of 20 kt or more at hill-crest level, with speed increasing with height but with little change of direction (strong waves are often associated with jet streams).

(iii) A stable layer somewhere between hill-crest level and a few thousand feet above.

Notes

(a) Turbulence-prone areas are generally near wave crests and troughs.

(b) These rules may be applied in a quantitative manner by using Scorer's parameter, l^2.

Ref: *Met. Rep.* No. 18, App. I.
Casswell, *Met.Mag.*, Mar. '66.

8.4.4. *Rotor-zone turbulence*

A rotor zone (not the same phenomenon as 'rotor streaming' of 8.4.2.(iii)) is a region of violent turbulence sometimes found somewhat above and to the lee of a large mountain ridge; its formation requires a well-defined lee escarpment and strongly developed lee waves. The rotor may, or may not, be indicated by a long roll of ragged cumulus or stratocumulus parallel to the main ridge. Rotor clouds are rare in the U.K. but the 'Helm Bar' of Crossfell is one example.

Ref: *Met. Rep.* No. 18.

CHAPTER 9. ICE ACCRETION ON AIRCRAFT

9.1. TYPES OF ICING

Type	Source	Formation and properties
Hoar frost	Vapour	Direct deposition on surface with temperature below frost-point of ambient air. White, crystalline coating.
Rime	Supercooled drops	Impact on surface below 0°C. Variable properties; two extreme forms described below: *(a)* Opaque rime: drops freeze without much spreading; light porous texture with large amount of entrapped air. *(b)* Clear or glaze ice: drops spread before freezing; smooth, glassy; sticks strongly to surfaces.
Rain ice	Supercooled rain	Formation as clear ice. Substantial deposit may form over extensive area. Rare.
Pack snow	Supercooled drops and snowflakes	Drops freeze on impact, embedding snowflakes.

Ref: HWF 18.2.

9.2. GENERAL REMARKS

Icing is very variable in both space and time in any cloud system, and any forecast must necessarily be of probabilities. The following remarks are therefore of a statistical nature.

(i) Since the number of water droplets intercepted depends upon the shape and speed of the aircraft, so will the icing. However, it is useful to generalize by regarding supercooled liquid water contents of clouds of 0.5 and 1.0 g/m^3 as separating 'light', 'moderate' and 'severe' icing.

(ii) Rain or cloud encountered by a cold aircraft descending may produce icing until the aircraft has warmed up. Intensity may be severe, particularly on forward-facing transparencies.

9.3. KINETIC HEATING EFFECTS

(i) Kinetic heating may prevent icing, even when the temperature of the ambient air is well below freezing-point, by raising the temperature of the aircraft above 0°C. The magnitude of the effect increases with the speed of the aircraft and with the density of the air, but decreases with increasing liquid water or ice-crystal content of the cloud.

Fig.9.1. shows for 500 mb the theoretical value of the threshold temperature of the ambient air above which icing will not occur, for clouds containing 0.5 g/m^3 of liquid water or ice crystals.

Ref: Monday Discussion, *Met. Mag.*, Feb.'60.

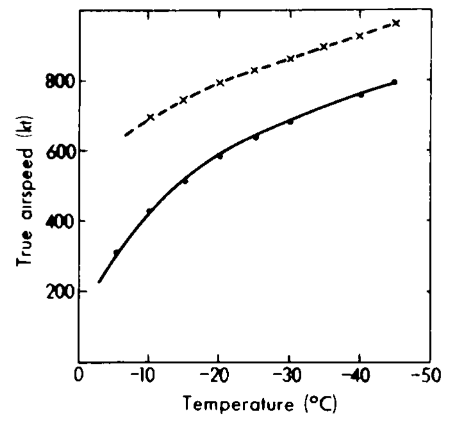


FIG.9.1. *Threshold temperatures for water cloud and ice-crystal cloud*

•——• Water cloud ×— — —× Ice-crystal cloud

Note. Values appropriate to mixed clouds will lie between the two curves.

(ii) The effect of aircraft speed on the threshold temperature for varying liquid water contents is shown in Table 9.I.

TABLE 9.I. *Threshold temperature*

Liquid water content	Aircraft speed (kt) 500	700
g/m^3	*degrees Celsius*	
0.0	−28.7	−56.3
0.5	−15	−31
2.0	−11.5	−24

Ref: Met.O. 768.

9.4. VARIATION WITH CLOUD TYPE

Type of cloud	Probability of icing	Intensity	Water content
			g/m^3
Cu Cb Ns	High	May be severe	0.2–4.0
Sc Ac Ac with As	≈ 50%	Rarely more than moderate	0.1–0.5
As	Low	Moderate or light	0.1–0.3
St	Low	Light	0.1–0.5

Ref: HWF 18.4.2.
Mason, B.J., *Physics of clouds*, 1957.

Note

In supercooled Sc relatively large liquid water contents may persist over many tens of miles near the upper boundary of the cloud layer. Moderate or severe icing can occur in these circumstances if an aircraft remains at this particular height.

9.5. CLOUD PROPERTIES AND ICING RISK

9.5.1. *Convective clouds (Cu and Cb)*

At temperatures of –40°C or below water droplets are very rare and the probability of icing is very low.

Between –40 and –20°C ice crystals greatly predominate and slight icing is likely, but moderate icing may occur for short periods in new, vigorously developing cloud cells.

Between –20 and 0°C there are few ice crystals and icing may be severe over a substantial depth of cloud and for a wide range of cloud-base temperatures. Use Fig.9.2. to determine intensity of icing in these cases.

Ref: HWF 18.4., 18.5.
Met.O.768.

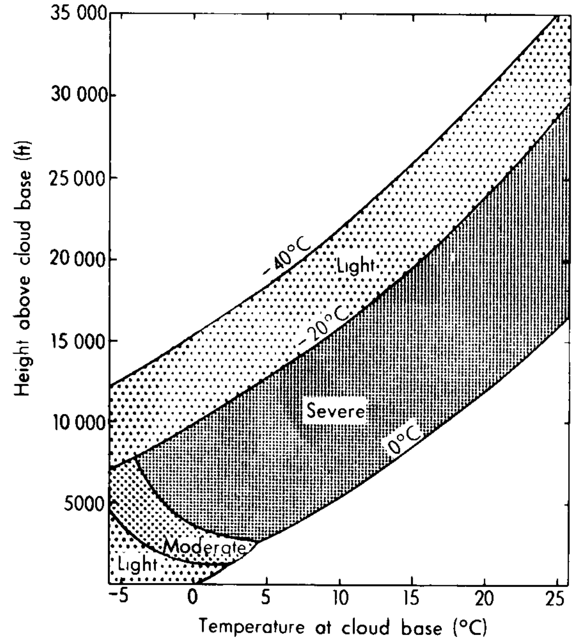


FIG. 9.2. *Approximate thickness of layers within which various degrees of icing may occur in convective cloud*

Base of cloud 950 mb. Ambient relative humidity 70%.

9.5.2. *Non-frontal layer clouds (St, Sc and Ac)*

These types are mainly formed by turbulent mixing and are rarely more than 3000 ft deep. In most cases water droplets predominate. To estimate the maximum liquid water content using the tephigram, plot the pressure and temperature at the cloud base, ascend along the saturated adiabatic to the cloud-top level, and note humidity mixing ratios (g/kg) at the top and bottom of the cloud. The difference is approximately (within 20-25% from the surface to 600 mb) numerically equal to the theoretical liquid water content (in g/m^3) near the top of the cloud. In practice, however, mixing with dry air at the cloud top reduces the maximum concentration to about half the theoretical value. For example, in a cloud layer about 3000 ft thick and with base at 850 mb, moderate icing is probable with cloud-top temperatures between 0 and -10°C whilst only light icing is likely below -10°C.

Note

Stratocumulus formed over the sea, especially in winter, is often formed by convection, the air being unstable to the sea surface temperature. In this case, and in that of stratocumulus formed by the spreading of cumulus, liquid water contents will be higher and the severity of icing greater.

9.5.3. *Frontal layer clouds*

These are formed by the slow ascent of a large mass of air and may be many thousands of feet deep. When associated with widespread continuous light precipitation they are mainly composed of ice crystals and snowflakes. Light icing is probable but is confined to a shallow layer (usually about 1000 ft deep) immediately above the 0°C surface.

The chance of severe icing is much increased with the more active fronts, especially where orographic uplift is present.

9.5.4. *Ice-crystal clouds*

These are not normally considered an icing hazard. However, a few cases of airframe icing have been reported, possibly arising from kinetic heating near the 'stagnation' points (the leading edges of structures where a small element of the flow is reduced to rest) of high-speed aircraft. This possibility must be increasingly borne in mind as aircraft speeds rise.

As the water content of cirriform clouds is quite small, icing will be only light but may still constitute a hazard as the cloud is often difficult to detect from an aircraft and because quite small amounts of icing of intakes, struts and guide vanes can cause compressor stall and flame-out of axial-flow jet engines.

A similar effect can occur when there is an accumulation of ice crystals or snowflakes in a jet engine intake, but this is a hazard only when the concentration of ice crystals is particularly high as in vigorous Cb in the tropics.

9.5.5. *Orographic cloud*

In cloudy air forced to rise over high ground the liquid water content is higher than in similar air elsewhere.

Air which is stable elsewhere may become unstable over and near high ground ('potential instability'). Thus a weak front may become active over high ground and deep layer cloud may contain vigorous convective cells in which the water content is very high because of the protective cloudy environment. Severity of icing is increased in these cases.

A further effect which should be watched for when a stable layer is lifted bodily is the lowering of the 0°C level which may occur as a result of adiabatic cooling.

Ref: HWF 18.4., 18.5.

9.6. ICING ASSOCIATED WITH FRONTS AND PRECIPITATION

Icing associated with fronts and depressions is very variable and each case should be treated on its merits. However, the following rules give some general guidance.

9.6.1. *Warm fronts*

In active warm fronts the possibility of severe icing should be indicated within 100 miles of the surface front; if the front is particularly active, the distance should be increased to 200 miles.

Notes

(a) If upper-level instability can be inferred the probability of severe icing should be increased.

(b) Severe rime icing may be encountered in As/Ns as much as 300 miles ahead of the surface front.

(c) Freezing rain (rain ice) ahead of a warm front can produce severe icing, but occurs only rarely.

9.6.2. *Cold fronts*

Severe icing may occur within 50-100 miles of the surface front, especially if it is active and there is a copious supply of moisture ahead.

9.6.3. *Depressions with ill-defined fronts*

In the thick cloud associated with the centre of a depression where the fronts are ill defined, severe icing may occur within 250 miles of the centre, particularly in the northern and western sectors.

Ref: HWF 18.5.3., 18.5.4.
Jones, MRP 1017.

9.7. ENGINE ICING

(i) Older types of piston engine are prone to carburettor icing because of adiabatic cooling of the air in the venturi and evaporation of the fuel. Increasing use of fuel injection systems makes this rare.

(ii) Turbine and pure jet engines are affected by icing of the intake rim, struts across the intake and guide vanes. Centrifugal engines are relatively insensitive but quite light icing can cause compressor stall in axial-flow engines (see 9.5.4. above).

(iii) Adiabatic expansion in the intake of jet engines at low forward speeds (e.g. on the approach or during take-off) may cause a drop in the temperature of intake air of around 5 degC. Engine icing may then occur at ambient temperatures around or above 0°C in a potentially hazardous situation.

Ref: Met.O.768.

9.8. HELICOPTER ICING

The effects of icing on helicopters are potentially more serious than on fixed-wing aircraft because:

(i) Wing (i.e. blade) loading is higher and aerodynamic effects therefore greater.

(ii) Extra blade weight due to icing may increase centrifugal forces on the rotor hub to unacceptable limits.

(iii) Slight asymmetries in the amount of icing on individual blades may set up severe vibration of the rotor assembly.

(iv) Icing may interfere with the operation of the cyclic-pitch control and hence control of the aircraft.

Note that since blade airspeed varies along the rotor blade, so does the severity of icing.

Ref: Met.O.768.

CHAPTER 10. CONDENSATION TRAILS (CONTRAILS)

10.1. FORECASTING EXHAUST CONTRAILS FROM JET AIRCRAFT

10.1.1. *Helliwell and MacKenzie's method*

The 'MINTRA' temperature (the ambient air temperature above which a Spitfire III piston-engined aircraft was unlikely to form exhaust trails of ice crystals at the level concerned) is used as a reference temperature, T_m. Values of T_m are shown by the MINTRA line on the 1956 and 1963 versions of the tephigram.

Forecast T_A, the ambient temperature for aircraft path; then

(i) If $T_A > (T_m - 11)$, trails unlikely.

(ii) If $(T_m - 11) \geqslant T_A > (T_m - 14)$, short non-persistent trails likely.

(iii) If $T_A < (T_m - 14)$, long persistent trails likely.

For borderline cases, account should be taken of the ambient humidity.

Persistent trails are more likely with high humidities.

Ref: Helliwell and MacKenzie, MRP 979.

10.1.2. *Appleman's method*

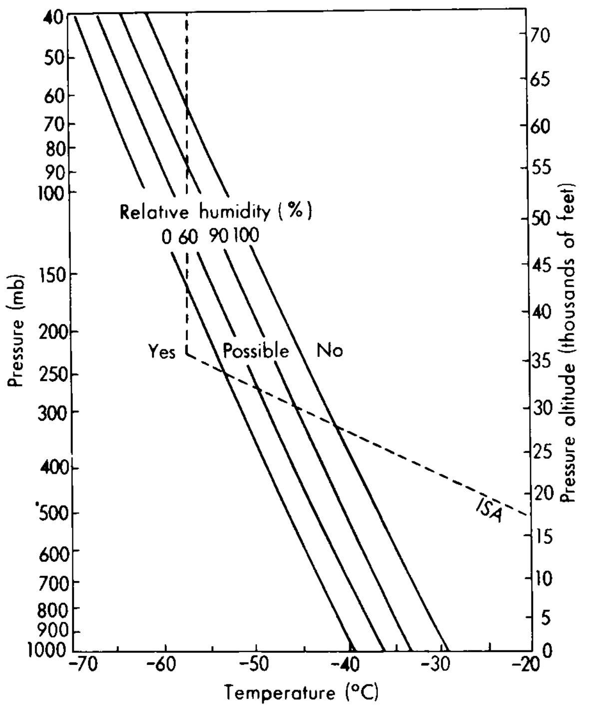


FIG. 10.1. *Graph of the relative humidity, with respect to liquid water, required for jet-aircraft contrail formation, expressed as a function of the pressure and temperature of the environment*

Fig.10.1. divides to three separate areas as follows:

(i) If the temperature at the given pressure lies to the left of the 0% R.H. line, trails should form whatever the humidity.

(ii) If the temperature lies to the right of the 100% R.H. line, trails should never form.

(iii) If the temperature lies between the 0% and 100% lines trails should form only if the ambient humidity with respect to water is at least equal to the value indicated at the corresponding point of the diagram.

Ref: Appleman, *Bull. Am. Met. Soc.*, 1953, p.14.
AWS Tech. Rep. 105-145, Washington, 1957.

10.2. ESTIMATING THE HUMIDITY

For use with Appleman's method, and also in borderline cases with the method in 10.1.1., estimates of average relative humidities with respect to water are given in Table 10.I.

TABLE 10.I. *Average relative humidities in and out of cloud*

	Troposphere	Stratosphere
	per cent	
In thin cloud	70	–
Close association with cloud	60	40
No close association with cloud	40	–
just above tropopause	–	10
5000 ft above tropopause	–	5
• 8000 ft above tropopause	–	5

APPENDIX 1. TABLES

TABLE 1. SOLAR ALTITUDE AT MIDDAY AND LOCAL MEAN TIMES OF SUNRISE AND SUNSET AT LATITUDE 52°N ON THE MIDDLE DAY OF EACH MONTH

	Jan.	Feb.	Mar.	Apr.	May.	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
	degrees											
Solar altitude	16.8	24.9	35.8	47.7	56.8	61.3	59.5	51.1	41.0	29.5	19.5	14.7
	hours											
Sunrise	0802	0716	0615	0505	0407	0339	0357	0443	0533	0624	0720	0801
Sunset	1617	1713	1804	1856	1946	2021	2013	1925	1817	1708	1609	1549
	minutes											
Correction per deg. lat.	+5.5	+2.9	+0.4	−2.3	−5.1	−7.4	−6.4	−3.6	−1.2	+1.7	+4.5	+6.5
Local correction												

For each degree of latitude *north* of 52°N

add / subtract — the correction to get the time of — sunrise / sunset

For each degree of latitude *south* of 52°N

subtract / add — the correction to get the time of — sunrise / sunset

The solar altitude at midday in latitude L may be obtained by adding the algebraic value of 52 − L to the figures given in the first row of the table.

TABLE II. CORIOLIS PARAMETER ($f = 2\Omega \sin \phi$)

Latitude ϕ	degrees 0	15	30	45	60	75	90
$f\ (h^{-1})$	0.000	0.136	0.262	0.372	0.453	0.508	0.524
$f\ (10^{-3} s^{-1})$	0.000	0.038	0.073	0.103	0.126	0.141	0.146
$\frac{\partial f}{\partial y} = \frac{2\Omega \cos \phi}{a}\ (10^{-11} m^{-1} s^{-1})$	2.29	2.12	1.98	1.62	1.14	0.59	0.00

a = earth's radius

TABLE III. ROSSBY LONG WAVES

Latitude degrees	Mean zonal wind speed (kt)									
	knots					*knots*				
	10	20	30	40	50	10	20	30	40	50
	L_s *kilometres*					L_s *degrees of longitude*				
70	5100	7200	8800	10200	11400	134	190	232	268	300
60	4200	6000	7300	8400	9400	76	108	132	152	170
50	3700	5300	6400	7400	8300	52	74	90	104	116
40	3400	4800	5900	6800	7600	40	57	69	80	90
30	3200	4500	5600	6400	7200	33	47	57	66	74

L_s = wavelength of stationary waves

TABLE IV *(a)*. TEMPERATURE CONVERSION

°F	−103	−94	−85	−76	−67	−58	−49	−40	−31	−22	−13	−4	
°C	−75	−70	−65	−60	−55	−50	−45	−40	−35	−30	−25	−20	
°F	5	14	23	32	41	50	59	68	77	86	95	104	113
°C	−15	−10	−5	0	5	10	15	20	25	30	35	40	45

TABLE IV *(b)*. DIFFERENCES

°F	1	2	3	4	5	6	7	8	
°C	0.6	1.1	1.7	2.2	2.8	3.3	3.9	4.4	
°C	0.5	1.0	1.5	2.0	2.5	3.0	3.5	4.0	4.5
°F	1	2	3	4	4.5	5	6	7	8

TABLE V. WATER-VAPOUR DENSITY AND PRESSURE IN SATURATED AIR

Temp. (°C)	40	30	20	10	8	6	4	2	0	
w_s over water	51.2	30.4	17.3	9.40	8.27	7.26	6.36	5.56	4.85	
e_s over water	73.8	42.4	23.4	12.3	10.7	9.3	8.1	7.1	6.1	
Temp. (°C)	−2	−4	−6	−8	−10	−12	−14	−16	−18	−20
w_s over water	4.22	3.66	3.17	2.74	2.36	2.03	1.74	1.48	1.26	1.07
over ice	4.14	3.52	2.99	2.53	2.14	1.80	1.51	1.27	1.06	0.88
e_s over water	5.27	4.54	3.91	3.35	2.86	2.44	2.08	1.76	1.49	1.25
over ice	5.17	4.37	3.69	3.10	2.60	2.17	1.81	1.51	1.25	1.03
Temp. (°C)	−22	−24	−26	−28	−30	−32	−34	−36	−38	−40
w_s over ice	.734	.607	.501	.413	.339	.277	.225	.183	.148	.119
e_s over ice	.850	.699	.572	.467	.380	.308	.249	.200	.161	.129

w_s = water vapour in gm m^{-3},
e_s = water vapour pressure in mb.

TABLE VI. ICAO ATMOSPHERE (DRY AIR)

Pressure	Temperature		Density	Height		Thickness of 1 mb layer	
mb	°C	°F	$g\ m^{-3}$	m	ft	m	ft
1013.2	15.0	59.0	1 225	0	0	8.3	27
1000	14.3	57.7	1 212	111	364	8.4	28
950	11.5	52.7	1 163	540	1 773	8.8	29
900	8.6	47.4	1 113	988	3 243	9.2	30
850	5.5	41.9	1 063	1 457	4 781	9.6	31
800	2.3	36.2	1 012	1 949	6 394	10.1	33
750	−1.0	30.1	960	2 466	8 091	10.6	35
700	−4.6	23.8	908	3 012	9 882	11.2	37
650	−8.3	17.0	855	3 591	11 780	11.9	39
600	−12.3	9.8	802	4 206	13 801	12.7	42
550	−16.6	2.1	747	4 865	15 962	13.7	45
500	−21.2	−6.2	692	5 574	18 289	14.7	48
450	−26.2	−15.2	635	6 344	20 812	16.1	53
400	−31.7	−25.1	577	7 185	23 574	17.7	58
350	−37.7	−36.0	518	8 117	26 631	19.7	65
300	−44.5	−48.2	457	9 164	30 065	22.3	75
250	−52.3	−62.2	395	10 363	33 999	25.8	85
200	−56.5	−69.7	322	11 784	38 662	31.7	104
150	−56.5	−69.7	241	13 608	44 647	42.3	139
100	−56.5	−69.7	161	16 180	53 083	63.4	208
90	−56.5	−69.7	145	16 848	55 275	70.5	231
80	−56.5	−69.7	128	17 595	57 726	79.3	260
70	−56.5	−69.7	112	18 442	60 504	90.6	297
60	−56.5	−69.7	96	19 419	63 711	105.7	347
50	−55.9	−68.8	80	20 576	67 507	127.0	417
40	−54.5	−66.0	64	22 000	72 177	160	525
30	−52.7	−62.8	47	23 849	78 244	215	706
20	−50.0	−58.0	31	26 481	86 881	326	1 072
10	-45.4	−49.7	15	31 055	101 885	669	2 195

APPENDIX II. CONSTANTS AND FUNDAMENTAL EQUATIONS

List of symbols not defined in the text.

m	= mass
p	= pressure
T	= temperature (in degrees Kelvin unless otherwise stated)
ρ	= density
V	= specific volume = $1/\rho$
e	= partial pressure of water vapour
e_s	= vapour pressure of saturated air
c_v	= specific heat of dry air at constant volume
c_p	= specific heat of dry air at constant pressure
$\mathrm{d}q$	= quantity of heat supplied to unit mass of gas
x,y,z	= Cartesian co-ordinates of a point
$\frac{\partial}{\partial x}$ etc.	= partial differentiation with respect to x etc.
u,v,w	= components of velocity along the axes of x,y,z
$\mathbf{V}$	= wind vector
X,Y,Z	= components of external forces acting on a unit mass of air
λ	= wavelength of radiation
a_λ	= absorptive power at wavelength λ
e_λ	= emissive power at wavelength λ
E_λ	= black-body radiation at wavelength λ
E	= total black-body radiation
I	= intensity of parallel radiation

1. GEOPOTENTIAL

a = earth's radius = 6.378×10^6m = 3963 miles at the equator
= 6.357×10^6m = 3950 miles at the poles

Ω = earth's angular velocity $\approx 2\pi/24$ hours = 7.3×10^{-5} radians s^{-1}

g = gravity at mean sea level
= $9.80616(1 - 0.0026373 \cos 2\phi + 0.0000059 \cos^2 2\phi)$ m s^{-2}
where ϕ = latitude

g_z = gravity at height z above mean sea level
= $ga^2/(a + z)^2 \approx g(1 - 2z/a) = g - 0.00000309z$,
if z is in metres

Φ = geopotential at height h =

$$\int_0^h g_z \mathrm{d}z = g\,a^2 \int_0^h \frac{\mathrm{d}z}{(a+z)^2}$$

$$= g\,h/(1 + h/a) \approx g\,h(1 - h/a)$$

The unit of geopotential is the potential energy acquired by unit mass on being raised through unit distance in a gravitational field of unit strength. It is measured in m^2s^{-2}. In the earth's gravitational field the unit of geopotential corresponds to a vertical distance of:

$1\ m^2s^{-2}g^{-1}$, approximately 0.102 m.

The dynamic metre is defined as 10 times this value, i.e. $\approx$ 1.02 m.

For practical purposes the geopotential metre is used defined as follows:

1 geopotential metre = 0.98 dynamic metre.

2. EQUATION OF STATE

R^* = universal gas constant = 83.1432×10^6 erg $mol^{-1}deg^{-1}$
= 8.31432 J $mol^{-1}deg^{-1}$

$pv = \dfrac{R^*T}{M} = RT$ where M = molecular weight (on scale ^{12}C = 12.0000)
= 28.9644 for dry air
= 18.0153 for water vapour

$p = R\rho T$

If p is in millibars $\quad R = 2.870 \times 10^3$ for dry air

$R_w = 4.615 \times 10^3$ for water vapour.

For moist air $p = R'\rho T$

where $R' = R/(1 - \overline{1 - \epsilon}\, e/p)$ and $\epsilon = 18.02/28.96$

$= 0.622 \approx 5/8$.

Alternatively, $p = R\rho T_v$

where T_v = virtual temperature $= T/(1 - \overline{1 - \epsilon}\, e/p)$

$= T(1 + \frac{3}{8}\frac{e}{p})$

For practical purposes, $T_v = T + \dfrac{x}{6}$

where x = mixing ratio in g kg^{-1}.

3. HYDROSTATIC EQUATION

$$\frac{\partial p}{\partial z} = -g\rho = -gp/RT$$

In an isothermal atmosphere

$$\log_e p_0 - \log_e p = gz/RT$$

whence

$$z = \frac{RT}{g} \log_e 10(\log_{10} p_0 - \log_{10} p)$$

$$= 67.4\ T\,(\log p_0 - \log p) \text{ metres.}$$

With a uniform temperature lapse β

we have $\quad T = T_0 - \beta z$

where T_0, T are the temperatures at the surface and at height z

$$\therefore \quad p/p_0 = (1 - \beta z/T_0)^{g/R\beta}$$

$$= (1 - \beta z/T_0)^{34.2/\beta}$$

if z is in km and β is in degC per km.

In the ICAO atmosphere up to 11km

$p_0 = 1013.2$ mb $\quad T_0 = 288°\text{K} \quad \beta = 6.5$ degC per km

$\therefore \ p/p_0 = (1 - 0.02257z)^{5.256}$ where z is in km.

In general,

$$\int_{z_1}^{z_2} dz = \int_{p_1}^{p_2} -\frac{dp}{\rho g} = \int_{p_1}^{p_2} -\frac{RT}{pg}\,dp = \frac{R}{g}\int_{p_2}^{p_1} T\,d(\log p) = \frac{R\,\overline{T}}{g}\log\frac{p_1}{p_2}$$

where $\overline{T}$ is the mean temperature in the layer between p_1 and p_2. This equation establishes the equivalence of thickness lines and mean isotherms.

4. THERMODYNAMICS

$$dq = c_v dT + p dv = (c_v + R)dT - v dp = c_p dT - v dp$$

$$c_v = 0.7147 \times 10^7 \text{ erg g}^{-1} \text{ deg}^{-1} = 0.7147 \times 10^3 \text{ J kg}^{-1} \text{ deg}^{-1}$$

$$c_p = 1.0031 \times 10^7 \text{ erg g}^{-1} \text{ deg}^{-1} = 1.0031 \times 10^3 \text{ J kg}^{-1} \text{ deg}^{-1}$$

$$\gamma = c_p/c_v = 1.403 \qquad R = c_p - c_v = c_p(\gamma - 1)/\gamma$$

In adiabatic conditions, $dq = 0$ and hence $dp = c_p \rho dT$

$$\therefore \Gamma_{dry} \equiv -\frac{dT}{dz} = -\frac{1}{c_p\rho}\frac{dp}{dz} = \frac{g}{c_p} = 9.8 \text{ degC km}^{-1}$$

Also, $\dfrac{dT}{T} = \dfrac{v dp}{c_p T} = \dfrac{R}{c_p}\dfrac{dp}{p} = \dfrac{\gamma-1}{\gamma}\dfrac{dp}{p}$

$$\therefore T/p^{\gamma-1/\gamma} = \text{constant, or } pv^{\gamma} = \text{constant.}$$

Potential temperature $\theta = T(p_0/p)^{\gamma-1/\gamma} = T(p_0/p)^{0.288}$

$$\therefore d(\log\theta) = \frac{dT}{T} - \frac{\gamma-1}{\gamma}\frac{dp}{p} = \frac{dT}{T} - \frac{R}{c_p}\frac{v}{RT}dp = \frac{dq}{c_p T}$$

$$\therefore \text{Entropy } (S) \equiv \int \frac{dq}{T} = c_p \log\theta + \text{constant}$$

In adiabatic conditions, with $dq = 0$, θ and S are constant.

5. WATER VAPOUR

$$x = \text{humidity mixing ratio} = \frac{\text{mass of vapour}}{\text{mass of dry air}} = \rho_w/\rho_d$$

$$= \frac{e}{R_w T} \Big/ \frac{p-e}{RT} = \frac{\epsilon e}{p-e} \approx \frac{\epsilon e}{p} \text{ where } \epsilon = 0.622 \text{ (See section 2)}$$

$$q = \text{specific humidity} = \frac{\text{mass of vapour}}{\text{mass of dry air + mass of vapour}} = \frac{\rho_w}{\rho_d + \rho_w}$$

$$= \frac{1}{1 + 1/x} = \frac{\epsilon e}{p - (1-\epsilon)e} \approx \frac{\epsilon e}{p}$$

T_v = virtual temperature = $T/(1 - \overline{1-\epsilon}\; e/p)$ (See § 2)

$$= T/(1 - \frac{1-\epsilon}{\epsilon} x) \approx T(1 + \tfrac{3}{5} x)$$

L_w = latent heat of evaporation

$= 597.3 - 0.56T$ cal g^{-1} $= (2500.9 - 2.34T) \times 10^3$ J kg^{-1}
where T is in °C

L_i = latent heat of melting ice

$= 79.7$ cal g^{-1} $= 333.7 \times 10^3$ J kg^{-1}

Clausius-Clapeyron equation is:

$$\frac{1}{e_s}\frac{de_s}{dT} = \frac{L}{R_w T^2} = \frac{\epsilon L}{RT^2}$$

$\Gamma_s \equiv$ saturated adiabatic lapse rate

$$= \Gamma_{dry} \frac{p + \epsilon L e_s / RT}{p + \epsilon^2 L^2 e_s / c_p RT^2}$$

6. EQUATION OF CONTINUITY

In Cartesian co-ordinates

$$-\frac{1}{\rho}\frac{d\rho}{dt} = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = \text{div } V$$

If the vertical co-ordinate is p,

$$-\frac{\partial p}{\partial p} = \left(\frac{\partial u}{\partial x}\right)_p + \left(\frac{\partial v}{\partial y}\right)_p = \text{div}_p \text{ V (or } \nabla_p \cdot \text{V)}$$

where $\dot{p} \equiv \frac{dp}{dt} = \frac{\partial p}{\partial t} + \frac{u\partial p}{\partial x} + \frac{v\partial p}{\partial z} + \frac{w\partial p}{\partial z}$

In mid-troposphere the last term $\frac{w\partial p}{\partial z}$, is frequently the largest and, since $\frac{\partial p}{\partial z} = -\rho g$, leads to the approximate relationship,

$$\dot{p} = -\rho g w$$

7. EQUATIONS OF MOTION

$$\frac{du}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial x} - 2\,\Omega\,(w\cos\phi - v\sin\phi) + X$$

$$\frac{dv}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial y} - 2\,\Omega\,u\sin\phi + Y$$

$$\frac{dw}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial z} + 2\,\Omega\,u\cos\phi - g + Z$$

If w is negligible compared with v, and if all external forces except friction are negligible, then

$$\frac{du}{dt} = f(v - v_g) + F_X$$

$$\frac{dv}{dt} = -f(u - u_g) + F_Y$$

where

$$u_g = -\frac{1}{f\rho}\frac{\partial p}{\partial y}; \quad v_g = \frac{1}{f\rho}\frac{\partial p}{\partial x}$$

are the components of the geostrophic wind V_g and F_X, F_Y are the horizontal components of the frictional force.

Rearranging, and putting $v' = v - v_g$, $u' = u - u_g$, then

$$fv' = \frac{du}{dt} - F_X = \frac{du_g}{dt} + \frac{du'}{dt} - F_X$$

$$fu' = -\frac{dv}{dt} + F_Y = -\frac{dv_g}{dt} - \frac{dv'}{dt} + F_Y$$

Expanding $\frac{du_g}{dt}$ and $\frac{dv_g}{dt}$ in terms of the partial derivatives, the equations become

$$fv' = \frac{\partial u_g}{\partial t} + \frac{u\partial u_g}{\partial x} + \frac{v\partial u_g}{\partial y} + \frac{w\partial u_g}{\partial z} - F_X + \frac{du'}{dt}$$

$$fu' = -\frac{\partial v_g}{\partial t} - \frac{u\partial v_g}{\partial x} - \frac{v\partial v_g}{\partial y} - \frac{w\partial v_g}{\partial z} + F_Y - \frac{dv'}{dt}.$$

The first term in each equation represents the isallobaric acceleration: The components of the isallobaric wind are

$$\frac{1}{f}\frac{\partial u_g}{\partial t} \quad \text{and} \quad -\frac{1}{f}\frac{\partial v_g}{\partial t}.$$

Further,

$$\frac{1}{f}\frac{\partial u_g}{\partial t} = \frac{1}{f}\frac{\partial}{\partial t}\left(\frac{1}{\rho f}\frac{\partial p}{\partial y}\right) \approx \frac{1}{\rho f^2}\frac{\partial}{\partial y}\frac{\partial p}{\partial t}$$

with a similar equation for $\frac{1}{f}\frac{\partial v_g}{\partial t}$

i.e. the isallobaric wind depends upon the gradient of the isallobars $\left(\text{isopleths of } \frac{\partial p}{\partial t}\right)$

The second and third terms in each equation represent the horizontal advective acceleration, $V.\,\nabla_h V_g$: The corresponding wind components are

$$\frac{1}{f}\left(\frac{u\partial u_g}{\partial x} + \frac{v\partial u_g}{\partial y}\right) \text{and} \quad -\frac{1}{f}\left(\frac{u\partial v_g}{\partial x} + \frac{v\partial v_g}{\partial y}\right)$$

The fourth terms in each equation represent the components of acceleration arising from vertical motion, the y- and x- components of the wind being

$$\frac{w}{f}\frac{\partial u_g}{\partial z} \quad \text{and} \quad -\frac{w}{f}\frac{\partial v_g}{\partial z}$$

The fifth terms represent the frictional forces, while the last terms are the components of the ageostrophic acceleration.

On isobaric contour charts where the vertical co-ordinate is p

$$\frac{\partial p}{\partial y} = -\frac{\partial p}{\partial z}\frac{\partial z}{\partial y} = g\rho\frac{\partial z}{\partial y} \quad \left(\frac{\partial p}{\partial z} = -g\rho\right)$$

hence, from the components of the geostrophic wind

$$u_g = -\frac{g}{f}\left(\frac{\partial z}{\partial y}\right)_p \qquad v_g = \frac{g}{f}\left(\frac{\partial z}{\partial x}\right)_p$$

or

$$u_g = -\frac{g}{f}\frac{\partial h}{\partial y} \qquad v_g = \frac{g}{f}\frac{\partial h}{\partial x},$$

where h is the height of the pressure surface above datum level.

8. VORTICITY

Vorticity, $\zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}$ i.e. the component of spin about a vertical axis, or, in polar co-ordinates, $\zeta = \frac{V}{r} + \frac{\partial V}{\partial n}$, where r is the radius of curvature of the streamline and n is directed normal to the streamline and is positive to the right of it.

Absolute vorticity $\zeta_a = \zeta + f$

Vorticity equation, in simplified form, and related to an isobaric surface, relates the change of vorticity with time to divergence, $\mathrm{div}_p V$,

$$\frac{d(\zeta + f)}{dt} = -(\zeta + f)\,\mathrm{div}_p V$$

or

$$\frac{1}{\zeta_a}\frac{d\zeta_a}{dt} = -\mathrm{div}_p V.$$

When $\mathrm{div}_p V$ is small, in particular at the 'level of non-divergence,'

$$\frac{d\zeta_a}{dt} = 0$$

Thus absolute vorticity ($\zeta + f$) is conserved following the flow. Northward-moving air acquires an increasing value for f and hence a decreasing value for ζ. The air must curve anticyclonically following a constant absolute vorticity trajectory. Similarly, southward-moving air must curve cyclonically.

Potential vorticity. If $\mathrm{div}_p\, V$ is not small then

$$\frac{d}{dt}\left(\frac{\zeta_a}{\Delta p}\right) = 0$$

or potential vorticity $\dfrac{\zeta + f}{\Delta p}$ = constant, where Δp is the pressure difference between the isentropic surfaces representing the top and bottom of an air column.

Development. In the usual case where $f \gg \zeta$

$$\frac{d}{dt}(\zeta + f) \approx -f\, \mathrm{div}_p\, V.$$

By using this equation near the surface – say at 1000 mb – and at some upper level and subtracting them to obtain the relative divergence it can be shown that

$$f(\mathrm{div}_p\, V - \mathrm{div}_p\, V_0) \approx -V' \frac{\partial}{\partial s}(\zeta' + 2\zeta_0)$$

where V' is the thermal wind, V_0 the surface wind, ζ' the vorticity of the thickness pattern and ζ_0 the vorticity of the surface pattern and $\partial/\partial s$ denotes differentiation along a thickness line.

If $\mathrm{div}_p\, V = 0$, i.e. the upper level is a level of non-divergence – say at 500 mb – then

$$\mathrm{div}_p\, V_0 = \frac{V'}{f} \frac{\partial}{\partial s}(\zeta' + 2\zeta_0)$$

relating the surface divergence to the thickness pattern.

9. RADIATION

Planck's law: $E_\lambda = c_1/\lambda^5(e^{c_2/\lambda T} - 1)$

where $c_1 = 3.74 \times 10^{-5}$ erg cm^{2}s^{-1} = 374×10^{-18} J m^{2}s^{-1}

and $c_2 = 14\,385$ μm degK = 1.4385×10^{-2} m degK

Wien's law: $\lambda_{max} = 2900/T$ μm

Kirchoff's law: $a_\lambda = e_\lambda/E_\lambda$

Stefan's law:

$$E = \int_0^\infty E_\lambda d\lambda = \sigma T^4$$

where $\sigma = 5.67 \times 10^{-12}$W cm^{-2} deg^{-4}
$= 56.7 \times 10^{-9}$W m^{-2} deg^{-4}
$= 1.35 \times 10^{-12}$ cal cm^{-2} deg^{-4}s^{-1}
$= 56.7 \times 10^{-9}$J m^{-2} deg^{-4}s^{-1}

Beer's law: $I = I_o e^{-k_\lambda m}$

where k_λ is the absorption coefficient at wavelength λ
and m is the mass of the absorbing medium

Ångström's formula for downward radiation:

$$R = \sigma T^4(A - B\ 10^{-\partial e})$$

where $A = 0.81$, B = 0.24, $\partial = 0.052$ and e = surface vapour pressure (mb)

Brunt's law: $R = \sigma\ T^4(a + b\sqrt{e})$

where a = 0.44 and b = 0.80

Nocturnal radiation: $N = \sigma\ T^4 - R$

Solar constant $= 2.00$ cal cm^{-2}min^{-1}
$= 1.395 \times 10^3$ J m^{-2}s^{-1}
$= 1.395$ kWm^{-2}

10. MISCELLANEOUS CONSTANTS AND RELATIONS

$e = 2.7183$ $\log_{10}e = 0.43429$ $\log_e 10 = 2.3026$

1 Å (angström unit) = 10^{-8} cm = 10^{-4} μm = 10^{-10} m

1 calorie (at 15°C) = 4.1855×10^7 erg = 4.1855 J

1 watt = 0.23892 cal s^{-1} = 14.335 cal min^{-1} = 1 J s^{-1}

1 langley = 1 cal cm^{-2} = 41.855×10^3 J m^{-2}

1 millibar = 1000 dyn cm^{-2} = 10^2 N m^{-2}

1 newton (N) = 1 kg m s^{-2}

1 knot = 1.689 ft s^{-1} = 1.853 km h^{-1} = 0.5148 m s^{-1}

1000 ft min^{-1} = 9.73 kt = 5.1 m s^{-1}

1 cm s^{-1} = $3600\, g\rho \times 10^{-3}$ mb h^{-1}
= 3.2 mb h^{-1} at 700 mb
= 1.6 mb h^{-1} at 300 mb
= 0.6 mb h^{-1} at 100 mb

Velocity of sound in dry air = $\sqrt{(\gamma RT)}$ (T in degrees Kelvin)
= $331\sqrt{(1 + T/273)}$ m s^{-1} if T is in degrees Celsius

Area of earth's surface = 510×10^6 km^2
Area of ocean surface = 360×10^6 km^2 (70.8% of earth's surface)
Area of land surface = 150×10^6 km^2 (29.2% of earth's surface)

Mass of atmosphere = 5.27×10^{21} g

Richardson number $Ri = g\beta / \left(\frac{\partial u}{\partial z}\right)^2$ where $\beta = \frac{1}{T}\left(\Gamma + \frac{\partial T}{\partial z}\right)$

Scorer's parameter $l^2 = \frac{g\beta}{U^2} - \frac{1}{U}\frac{\partial^2 U}{\partial z^2}$

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NOTES