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HANDBOOK
OF
WEATHER FORECASTING

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PREFACE

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

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

CHAPTER 15

HUMIDITY

CONTENTS

Chapter 15 HUMIDITY

	<i>Page</i>
15.1. Introduction	1
15.2. Measures of humidity and their conservatism	1
15.3. Mechanisms for the distribution of moisture in the atmosphere .	3
15.4. Variability of humidity	4
15.5. Observed distribution of moisture in the atmosphere	6
15.6. Forecasting humidities	34
Bibliography	40

LIST OF DIAGRAMS

Figure		Page
1.	Temperature and humidity readings at 550 millibars over Abingdon-Gloucester area, 1100-1200 G.M.T., 19 April 1948	5
2.	Temperature and humidity readings at 700 millibars over Farnborough-Salisbury Plain area, 1100-1200 G.M.T., 27 April 1948	5
3.	Differences between relative humidities as observed by aircraft and radio-sonde	7
4.	Temperatures and humidities as measured (a) from an aircraft and (b) by radio-sonde when a layer of stratocumulus was present	8
5.	Mean variation of absolute humidity with height for each hour for March, June, September and December at Rye	10,11
6.	Diurnal variation of humidity at 1.1 and 106.7 metres for March, June, September and December at Rye	13
7.	Diurnal variation of humidity at 1.1 and 106.7 metres for clear and overcast days in winter and summer at Rye	14
8.	Schematic illustration of a dry tongue associated with a front	19
9.	Profile of temperature and frost-point at 800 millibars, 1615-1710 G.M.T., 4 January 1952	19
10.	Surface chart, 1500 G.M.T., 3 January 1952	20
11.	Surface chart, 1500 G.M.T., 4 January 1952	20
12.	700-millibar contour chart, 1500 G.M.T., 4 January 1952 and 24-hour trajectories at 700 millibars terminating at that time	21
13.	Cross-section, 1500 G.M.T., 4 January 1952	21
14.	Aircraft sounding, 1030 G.M.T., 26 April 1949	23
15.	Aircraft soundings illustrating the "moist-layer" effect	24
16.	Aircraft sounding, 1500 G.M.T., 24 May 1948	25
17.	Aircraft sounding, 0930 G.M.T., 5 July 1949	25
18.	Mean ascent curves; summer, winter and year	28
19.	Mean ascent curves for tropopauses with and without inversion of temperature	28
20.	Mean ascent curves; selected ascents; high tropopause and low tropopause	29
21.	Mean frost-points at levels above the tropopause	30
22.	Relation between tropopause pressure and frost-points at the 300-, 250- and 200-millibar levels	32
23.	Relation between temperature at the 500-millibar level and frost-point at the 250-millibar level	33
24.	Relation between ambient air temperature and frost-point depression observed on high-level ascents above 25,000 feet	34
25.	Calculation of decreased dew-point in Föhn conditions	37

CHAPTER 15

HUMIDITY

15.1. INTRODUCTION

It is hardly possible to exaggerate the importance of humidity in the atmosphere, for without water vapour there would be no cloud, rain or fog - in fact no weather. Furthermore the horizontal and vertical transport and the phase transformations of water are processes by which substantial quantities of energy in the form of heat are redistributed in the atmosphere. In other words, water plays an active part in the working of the atmospheric heat engine, particularly at lower and mid-tropospheric levels. At higher levels (and lower temperatures), where only minute quantities of water are normally present, phase changes of water may be thermodynamically much less significant but water may still have important effects on radiation, and ice particles at the higher tropospheric levels may lead to changes in the constitution of clouds at lower levels which are important for the release of precipitation. In addition, as was seen in Chapter 14, the absorption properties of water vapour for certain band widths of long-wave terrestrial radiation exert a marked effect on air temperature.

Although water is such an important constituent of the atmosphere, very few forecasts contain explicit values of the humidity to be expected. Nevertheless even quite general forecasts of clouds and precipitation cannot be reliably compiled without a careful assessment of humidity, whilst accurate forecasts of the fine detail of clouds and precipitation demanded by aircraft operators require very close estimates of the spatial and temporal distribution of humidity. When fog is likely, an accurate assessment of humidity may be the crucial factor determining the success or complete failure of the forecast. For some very specialized work also humidity is important. For example, it is a factor to be considered when forecasting the levels at which aircraft may form condensation trails and the vertical distribution of humidity is of importance when considering the propagation of radio (mainly short) waves.

Very few techniques dealing solely with the forecasting of humidity have been developed. The quantitative techniques which are available for forecasting those particular values of humidity (convective condensation level, fog-point etc.) which are needed for some elements of a forecast have been included in the relevant chapters (for example, Chapter 16 *Cloud and precipitation*, Chapter 17 *Visibility* etc.). A consequence of the deliberate exclusion of quantitative techniques from this chapter is that the text becomes qualitative and descriptive in character. However, the chapter has been designed to provide a broad framework of knowledge on humidity within which the forecaster can make his assessments of humidity based on physical reasoning and on an understanding of the probable accuracy of observations and of the moisture distribution in the atmosphere.

15.2. MEASURES OF HUMIDITY AND THEIR CONSERVATISM

15.2.1. Measures

There are many ways of expressing the humidity of the air. Most standard textbooks contain a section on the various properties of the several measures of humidity. The basic material is not repeated in this handbook and the reader should refer to the textbooks for proofs and fundamental relationships.

Handbook of Weather Forecasting

The formal definitions of the meteorological measures of humidity are contained in *World Meteorological Organization Publication No. 49*.^{1*} These are complicated by the fact that air and water vapour do not behave exactly as perfect gases. The mixing ratio is taken as the fundamental measure and the vapour pressure, dew-point and thermodynamic wet-bulb temperature are defined by mathematical relations with it. The commonly used measures of humidity are as follows:

Vapour pressure,
Relative humidity,
Humidity mixing ratio,
Dew-point,
Dry- and wet-bulb temperatures,
Dry- and wet-bulb potential temperatures.

With sufficient accuracy for most forecasting practice we may regard them as defined below.

15.2.1.1. *Vapour pressure.* The vapour pressure (e) is the partial pressure exerted by the water vapour present in the sample of air under consideration.

15.2.1.2. *Relative humidity.* If e is the vapour pressure actually present and e_s is the saturation vapour pressure at the same temperature then the relative humidity (RH) is defined as $RH = 100 \frac{e}{e_s}$ and is always expressed as a percentage.

15.2.1.3. *Humidity mixing ratio.* The humidity mixing ratio is defined as the ratio of the mass of water vapour present to the mass of dry air in a given volume of moist air. It is usually denoted by x and expressed in grammes per kilogram of dry air. If p is the pressure of the moist air and e is the vapour pressure then $x = \frac{622e}{p-e}$ (in grammes per kilogram). If x_s is the humidity mixing ratio at saturation then $100 \frac{x}{x_s}$ represents the relative humidity to a high degree of accuracy.

15.2.1.4. *Dew-point temperature.* The dew-point temperature of a sample of air at pressure p is defined as the temperature for which the saturation vapour pressure over a plane water surface is equal to the partial pressure of water vapour actually present in that sample at pressure p . When air is cooled at constant pressure to below its dew-point, condensation normally occurs.

15.2.1.5. *Frost-point temperature.* The frost-point temperature of a sample of air at pressure p is defined as the temperature for which the saturation vapour pressure over a plane ice surface is equal to the partial pressure of water vapour actually present in the sample at pressure p .

15.2.1.6. *Dry- and wet-bulb temperatures.* The dry- and wet-bulb temperatures of the air are the temperatures recorded by well ventilated dry- and wet-bulb thermometers. If T and T_w are the dry- and wet-bulb temperatures, x the humidity mixing ratio and x' the humidity mixing ratio of saturated air whose temperature is T_w , L the latent heat of vaporization at temperature T_w and C_p the specific heat of dry air at constant pressure, then, to a close degree of approximation,

$$T_w = T - \frac{L(x' - x)}{C_p} \quad (1)$$

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

Humidity

A derivation of this equation is contained in textbooks (see, for example, Petterssen²).

It should be borne in mind that, mainly because of the limitations in ventilation, the wet-bulb temperature recorded in the standard type of screen may differ from T_w as defined above. This difference is not trivial and the discrepancy between the theoretical wet-bulb temperature given by equation (1) (or obtained from the tephigram) and the value appropriate to a thermometer in a normal screen as given in the hygrometric tables may amount to 2°F. or more.

15.2.1.7. *Dry- and wet-bulb potential temperatures.* These potential temperatures of a sample of air which does not contain condensed water are the dry- and wet-bulb temperatures which the air would have if brought to a standard pressure - normally 1000 millibars - by an adiabatic process. These temperatures are readily obtained by means of the tephigram.

15.2.2. *Conservatism*

The various measures of humidity are, unfortunately, not conservative in many of the physical processes which occur in the atmosphere. They cannot therefore be used as "tracers" to identify parcels of air from one chart to another and forecast values must be obtained by modifying the observed values according to the physical processes which are expected to have taken place between the time of observation and the time of forecast. A discussion of the conservatism of the various measures of humidity is contained in most standard textbooks (see, for example, Pettersen,² Chapter 1) and a summary of their conservative properties only is given below.

15.2.2.1. *Humidity mixing ratio.* Conservative for dry-adiabatic temperature changes and for radiation processes so long as neither evaporation nor condensation takes place.

15.2.2.2. *Dew-point.* Conservative for isobaric temperature changes provided water vapour is neither added to nor withdrawn from the air.

15.2.2.3. *Wet-bulb temperature.* Conservative for evaporation of water into the air (for example, by falling rain) when the heat of evaporation is supplied by the air.

15.2.2.4. *Wet-bulb potential temperature.* Conservative for evaporation of water into the air (for example, from falling rain) when the heat of evaporation is supplied by air. Conservative for dry- and wet-adiabatic changes provided condensed water is removed and approximately conservative if water droplets remain in the air.

15.3. MECHANISMS FOR THE DISTRIBUTION OF MOISTURE
IN THE ATMOSPHERE

All moisture in the atmosphere emanates from the surface of the earth. For most of the time evaporation is taking place over the greater part of the earth's surface. The periods during which the dew-point of the air exceeds the surface temperature and water is being condensed directly on the ground or sea are much shorter than the periods of evaporation. Moisture is also supplied to the air by the transpiration of vegetation.

The vertical transport of moisture from the earth's surface is carried out on four scales: firstly by molecular diffusion through the boundary layer of the

Handbook of Weather Forecasting

atmosphere in contact with the surface; secondly by mechanical turbulence, mainly in the lowest thousand feet of the atmosphere; thirdly by free convection; and fourthly by the slower large-scale vertical currents of fronts and depressions. Horizontal redistribution on the synoptic scale depends entirely on the large-scale horizontal air movements.

During some of these processes of distribution moist air is subjected to cooling to such an extent that clouds form. Although in some clouds the cloud particles remain very small throughout the life of the cloud, in others some of the cloud particles grow to precipitation size. As these precipitation elements grow their fall-speeds under gravity increase so that they are carried out of the cloud towards the ground. When falling precipitation evaporates into dry air layers, further redistribution of moisture occurs. When the precipitation reaches the earth the moisture returns to replenish the source and the cycle is complete.

This cycle of evaporation of water from the surface and its return mainly as precipitation is a continuous and fairly rapid one. Quoting from a recent paper by Sutcliffe³ the mean annual rainfall of the earth as a whole may be put at about 90 centimetres and the water-vapour content of the atmosphere at about the equivalent of 2.5 centimetres of water. This is only one-fortieth of the annual rainfall, or about nine days' rainfall, even if all the water in the atmosphere were precipitated. These figures make it clear that the replenishment of water vapour to the atmosphere must be continually maintained at a fairly rapid rate.

15.4. VARIABILITY OF HUMIDITY

In view of the cycle of humidity and the processes which distribute water horizontally and vertically it is not surprising that the horizontal and vertical distributions of humidity should exhibit wide variations. On synoptic charts it is possible to recognize a broad uniformity of humidity in air masses and a semi-orderly variation of humidity across frontal surfaces. Some recognized patterns have been described in Chapter 7. Superimposed on these variations is a variability of humidity on a small scale (of the order of a few miles). This variability is seldom observed and cannot normally be identified from routine reports and charts. It introduces an element of uncertainty in forecasting and it is important that forecasters should have some knowledge of this variability.

In the lowest few hundred feet the patchiness of the dew-point of the air over the land is caused mainly by variations in the underlying surface. It is well known and easily understood that dew-points are greater in air over low-lying marshy ground or lush pasture than over hard dry surfaces almost devoid of vegetation. Forecasters can make some attempt to allow for these variations in very local forecasting. In addition to this low-level patchiness, flights by the Meteorological Research Flight^{4,5,6} at various levels from 940 millibars up to the tropopause have shown that, in cloud-free air, considerable variations in humidity occur over small horizontal distances. Variations of frost-point of 5°C. in 15 miles are common and changes of 15°C. or more are sometimes observed. Two horizontal patterns of frost-point and temperature examined by Frith⁴ are reproduced in Figures 1 and 2. Figure 1 shows variations which are common, that is, variations in temperatures and frost-point of the order of 1°F. and 10°F. (that is, approximately 0.5 and 5.0°C.) respectively over an area about 25 miles by 15 miles. Figure 2 shows variations in frost-point in excess of 25°F. (about 14°C.) over distances of 5 miles. Although Figure 2 shows a very close association between cold moist air and warm dry air other examples examined by Frith did not exhibit quite such a close association.

Humidity

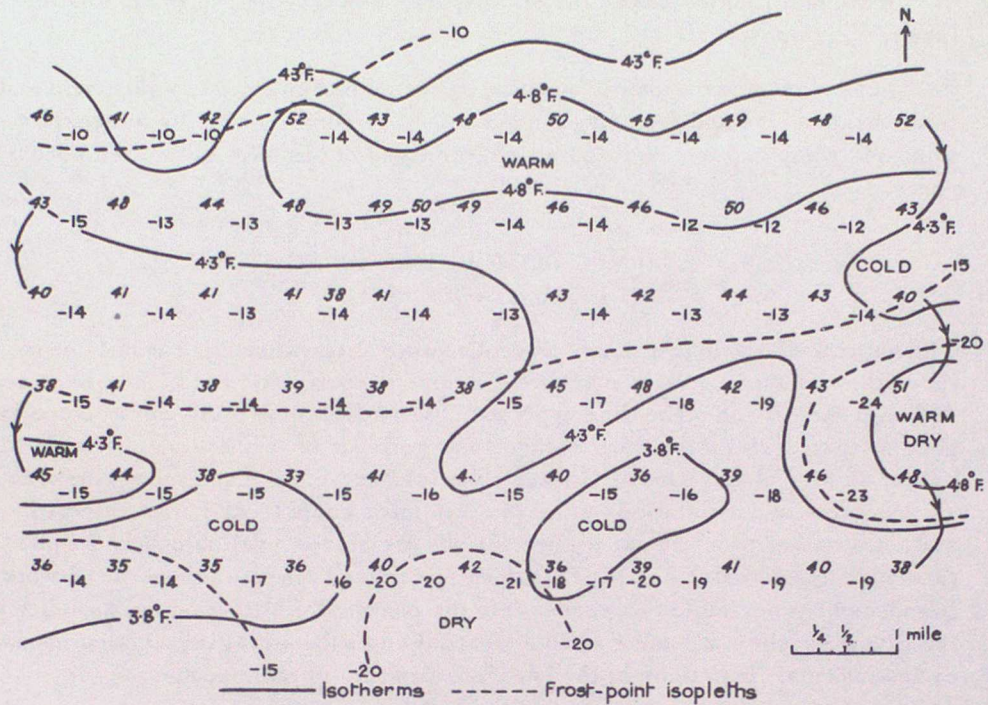


FIGURE 1 Temperature and humidity readings at 550 millibars over Abingdon-Gloucester area, 1100-1200 G.M.T., 19 April 1948

The rows of figures along the aircraft's tracks are:
 Italic (upper) figures: $(T \times 10)$, where T is the temperature in $^{\circ}\text{F}$.
 Lower figures: Frost-points in $^{\circ}\text{F}$.

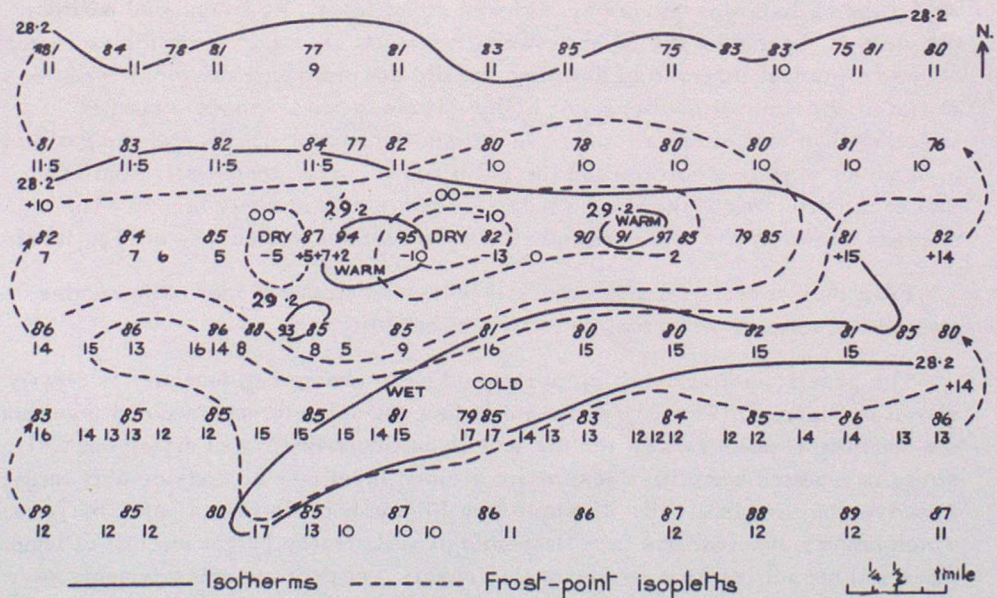


FIGURE 2 Temperature and humidity readings at 700 millibars over Farnborough-Salisbury Plain area 1100-1200 G.M.T., 27 April 1948

The rows of figures along the aircraft's tracks are:
 Italic (upper) figures: $(T - 20) \times 10$, where T is the temperature in $^{\circ}\text{F}$.
 Lower figures: Frost-points in $^{\circ}\text{F}$.

Handbook of Weather Forecasting

Frost-point fluctuations in the stratosphere appear to be somewhat smaller than in the troposphere.

These random variations in humidity are often present in the troposphere and lower stratosphere and their magnitudes should be borne in mind by forecasters when analysing and forecasting from observations of humidity in the free atmosphere.

15.5. OBSERVED DISTRIBUTION OF MOISTURE IN THE ATMOSPHERE

The network of stations making frequent routine observations of humidity near the surface is fairly good in north-west Europe over the land but is meagre at sea. Although the network of routine upper air observations near north-west Europe is good compared with many other areas of the world, it is very sparse compared with that of surface observations. Surface observations of humidity over a wide area are available at most stations at least every three hours (hourly from selected stations) whereas routine upper observations are seldom available more frequently than once every twelve hours. This open spatial and temporal upper air network introduces an inevitable coarseness into the observed distributions of humidity in the free atmosphere but there is an additional and substantial imprecision imposed by instrumental limitations of the humidity elements of radio-sondes.

15.5.1. *Instrumental limitations of the humidity elements of radio-sondes.*

Brewer and Harrison⁷ examined a series of observations of humidity made with a frost-point hygrometer installed in an aircraft which was flown in circles around radio-sondes which were operated at reduced lift to keep within the rate of climb of the aircraft. Plots of the humidities as observed by radio-sonde and aircraft on two occasions are shown in Figure 3. On the first ascent the large discrepancy at about 900 millibars is due to the aircraft selecting cloud-free air whilst the balloon was traversing a broken cloud layer. At around 600 millibars the aircraft observed a dry layer between two moist layers whereas the radio-sonde showed a gradual increase of humidity and did not reproduce the rapid variations shown by the aircraft readings. At higher levels the radio-sonde record is smoother than that of the aircraft. On the second ascent illustrated, the variations in humidity were less marked and the radio-sonde curve is generally similar to that of the aircraft but the radio-sonde values are considerably higher. The changes shown by the radio-sonde curve are smoothed particularly at high levels.

From their investigation Brewer and Harrison concluded that radio-sondes smooth out some of the actual variations of humidity.

The general instrumental lag in responding to changes in humidity is clearly shown in Figure 3. When the radio-sonde has ascended through a cloud layer and the humidity element is wet, the element is particularly slow in drying out and errors in reported humidities just above a cloud layer may be considerably larger. This type of error is vividly illustrated by Figure 4, taken from a paper by James⁸ which reports observations from Meteorological Research Flight aircraft of temperatures and humidities near stratocumulus cloud. Temperature measurements were made by a flat-plate thermometer, and frost-points were measured by a Dobson-Brewer manual hygrometer. Observations of temperatures and frost-points (or dew-points) were made at intervals of 250 feet during the ascent and subsequently frequent observations were made on level runs near the stratocumulus layers investigated. The aircraft measurements made on one flight are shown in Figure 4(a).

Humidity

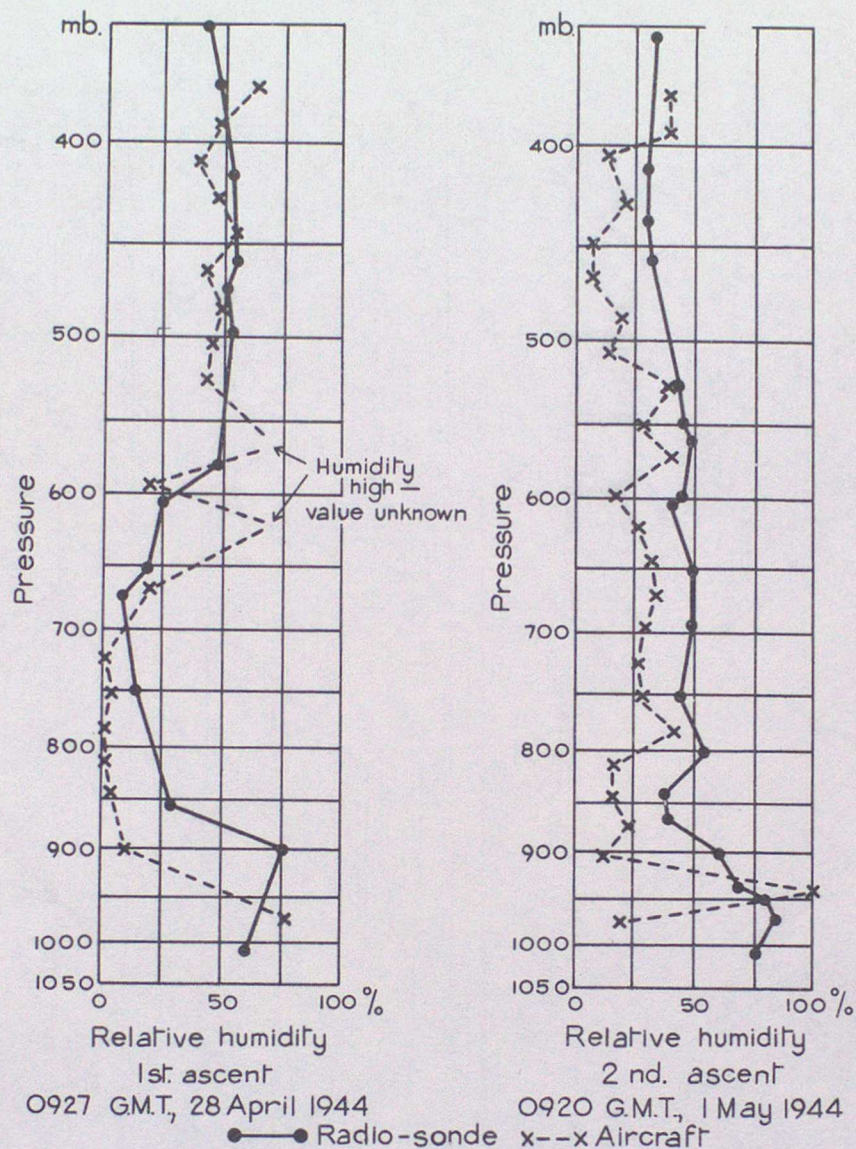


FIGURE 3 Differences between relative humidities as observed by aircraft and radio-sonde

Figure 4(b) shows the temperature and humidity structures of the air based on the values reported from the radio-sonde at Crawley at about the same time.

The large discrepancies in the humidities above cloud top as measured from the aircraft and by the radio-sonde are very striking. Furthermore the shapes of the profiles are fundamentally different in some layers (for example, just below cloud base and just above cloud top). If it can be inferred that a radio-sonde ascent has penetrated a cloud layer the forecaster should use considerable discretion when applying reported humidities to forecasting. The same sort of error is likely to occur when the radio-sonde penetrates a convective cloud but this type of occurrence can seldom be reliably inferred.

There is another instrumental deficiency of which forecasters should be aware. The sensitivity of humidity elements currently in common use on radio-sondes decreases with decreasing temperature. An indication of the probable accuracy of

Handbook of Weather Forecasting

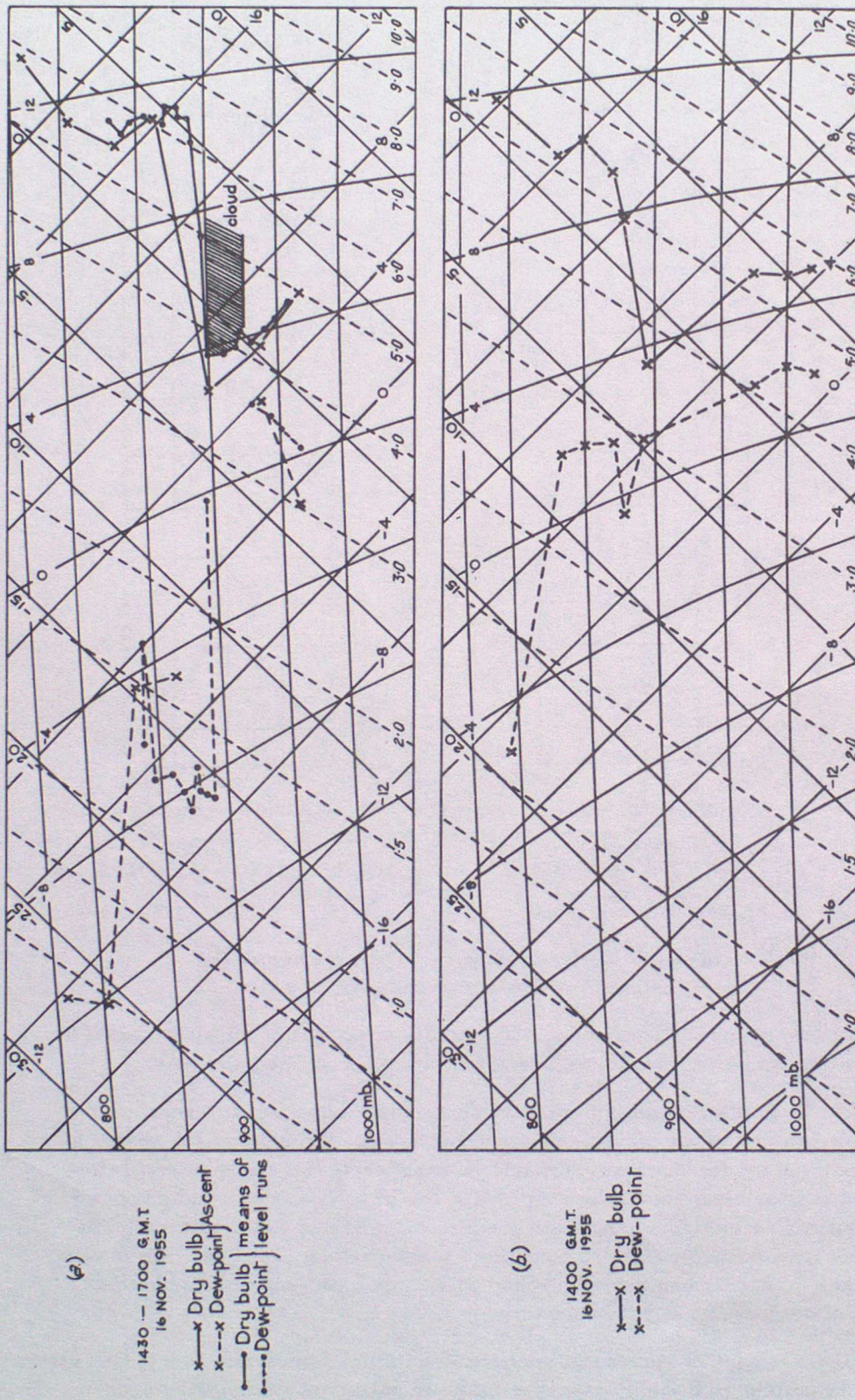


FIGURE 4 Temperatures and humidities as measured (a) from an aircraft and (b) by radio-sonde when a layer of stratocumulus was present

Humidity

humidities measured by radio-sonde is given by the following quotation from *World Meteorological Organization Publication No. 8*:⁹

"The accuracy of humidity units of radio-sondes leaves much to be desired. Their errors arise mainly from their low rate of response at low temperatures and their insensitivity near the extremes of humidity. Since the variation of humidity with height is generally much more irregular than that of temperature it is not very practicable to apply lag corrections on a routine basis. With the type of hygrometer element at present in common use an accuracy of the order of 5% can be expected at temperatures down to about 0°C. From about -20° to -40°C. the indications are only useful for qualitative information and at still lower temperatures, where the concentration of water vapour is minute, the elements must be regarded as useless."

It is clear that the open spatial and temporal network of reporting stations together with the marked instrumental defects of the humidity elements of radio-sondes impose severe limitations on the accuracy with which the forecaster can determine the horizontal and vertical distribution of humidity in the atmosphere as routine. These limitations are a great handicap to the practising forecaster. Lack of accurate observations of humidity has also been a factor impeding progress in those researches in which humidity is an important parameter. In some research projects special efforts have been made to obtain accurate and detailed measures of humidity in the hope that the accurate data would materially advance knowledge and understanding of the distribution of humidity and lead to a quantitative treatment of the role of humidity in various atmospheric processes. The account of the distribution of humidity which is given in sections 15.5.2 to 15.5.5 which follow, has been compiled partly from analyses of some of those special observations and partly from a statistical treatment of both special and routine observations of humidity. Although sections of the text include a degree of detail which is beyond the compass of most stations relying on routine observations only, the text should increase the forecasters background knowledge and so contribute towards a sound day-to-day assessment of probable humidity distributions and variations.

15.5.2. *Humidity in the lower layers of the atmosphere*

The meteorology of the lowest layers of the atmosphere very close to the ground (say below screen level) forms a very specialized branch of meteorology and a consideration of it is not appropriate to the handbook. If the reader wishes to study the humidity structure of this lowest layer he should consult the original papers and more specialized texts which are available. Amongst the latter may be mentioned Sutton's¹⁰ textbook which contains an advanced mathematical treatment of evaporation and diffusion and Geiger's¹¹ book which gives a mainly descriptive account of the microclimate near the ground.

The most detailed measurements of temperature and humidity in the lower layers of the atmosphere over the United Kingdom are those made over a period of three years at Rye, Sussex. By means of instruments installed in a normal screen and at three levels on a lattice tower, almost continuous records of temperatures and humidities were obtained at four heights up to 106.7 metres (350 feet approximately). The site is towards the western end of Romney Marsh and 5 kilometres north-north-east from the nearest point of the coast of the English Channel. From east-north-east to west-south-west, through south, the ground is practically flat to the coast; to the north and north-east it is 15 to 20 kilometres before rising ground is encountered in the Downs and the Weald, and to the west and north-west it is 4 to 6 kilometres before the land rises fairly sharply to about 60 metres (200

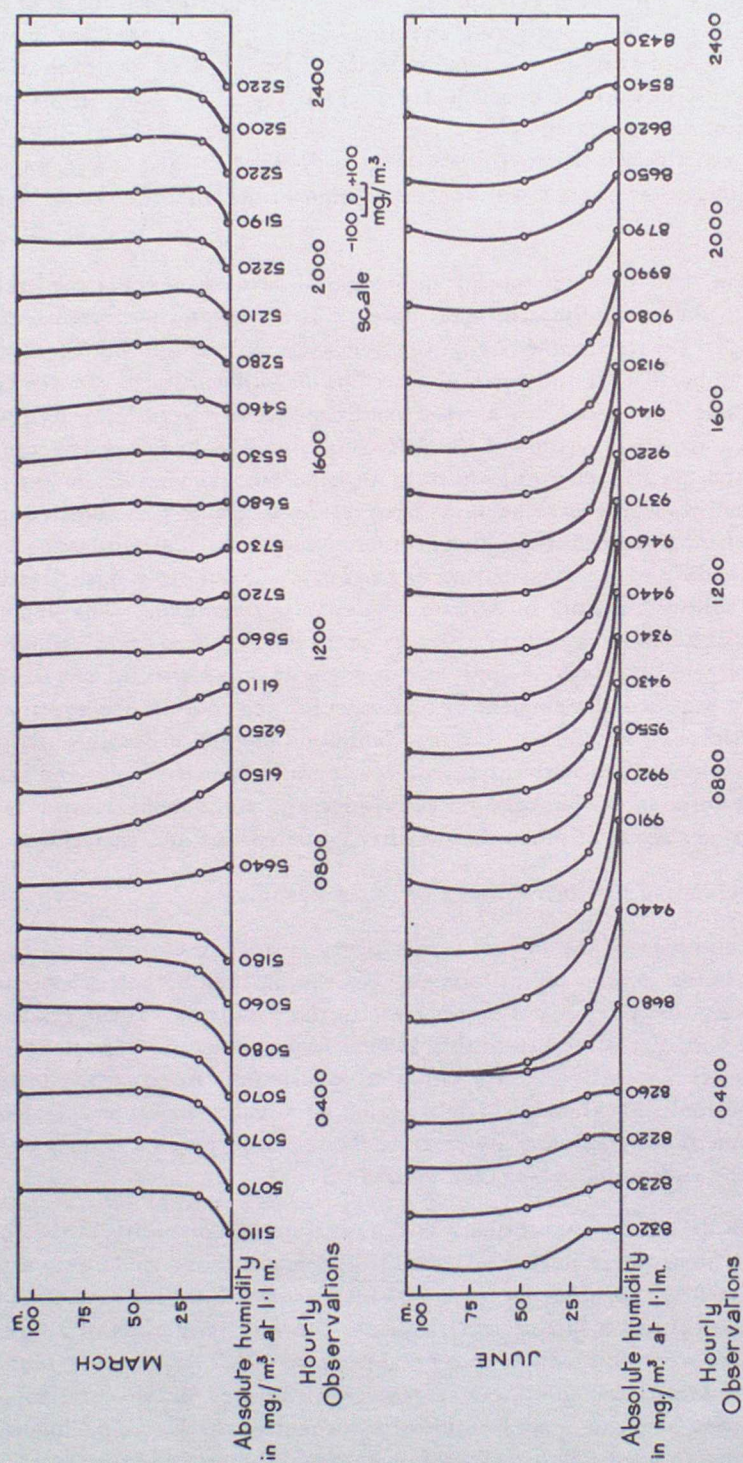


FIGURE 5 Mean variation of absolute humidity with height for each hour for March and June at Rye

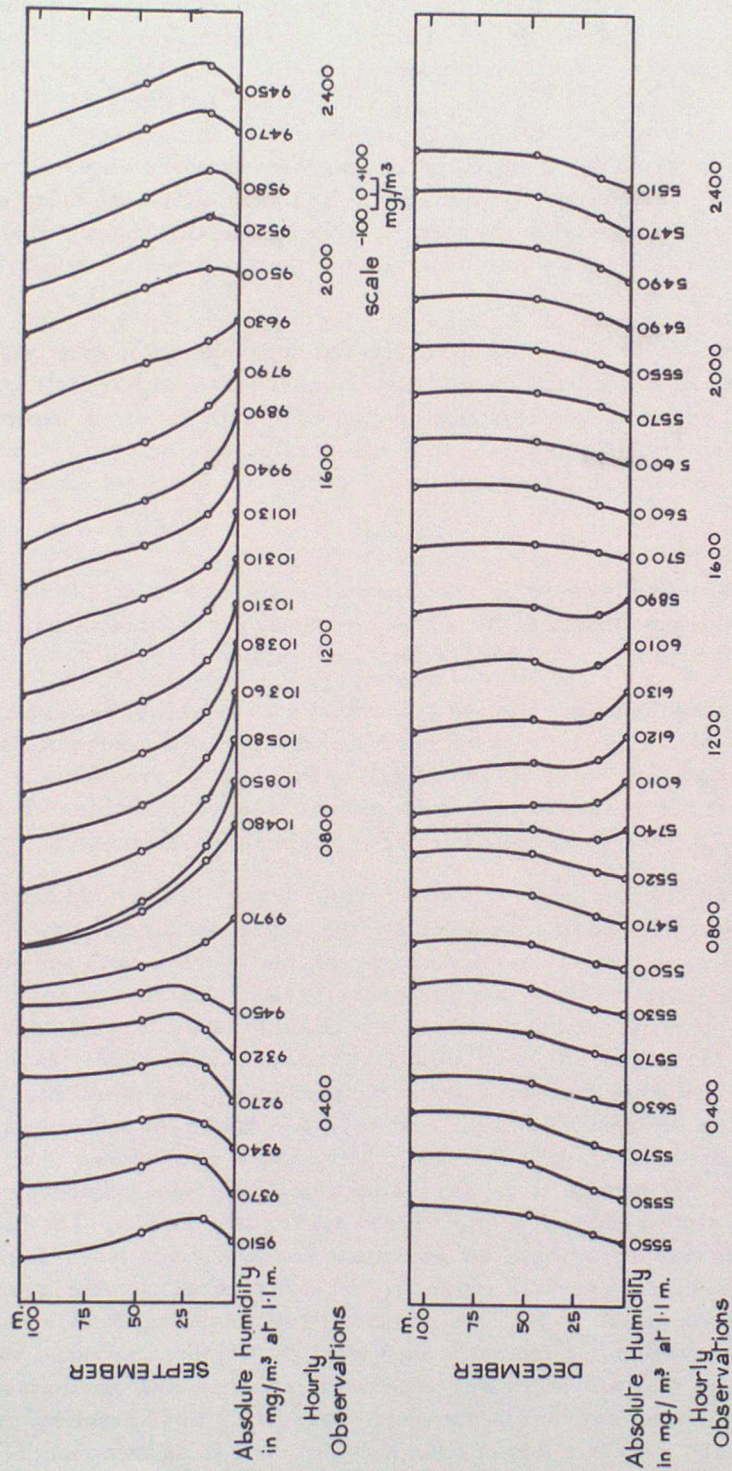


FIGURE 5 (CONTD.) Mean variation of absolute humidity with height
for each hour for September and December at Rye

Handbook of Weather Forecasting

feet). The marsh is predominantly grassland with a little arable; there are few trees and no woodlands.

In many respects the site is unrepresentative and the results of the Rye data cannot be applied directly to other areas. However there is no data comparable in quantity, frequency or detail for other sites in the United Kingdom and it was considered preferable to present observational material rather than purely descriptive accounts. The observations were analysed by Best, Knighting, Pedlow and Stormonth¹² and their results contain a massive series of statistics. A small selection has been made and included in this handbook so that forecasters will have before them the broad pattern of events at Rye. For a thorough study forecasters should consult the original paper. It is important that the limitations of the site at Rye should be constantly borne in the minds of forecasters when interpreting the following account.

In the original paper humidities are expressed in milligrams of water vapour per cubic metre. Absolute humidity is seldom used in practical forecasting and, where possible, approximately equivalent values of humidity of direct importance to forecasting (for example, dew-point variations) have been included. Absolute humidity may be converted to vapour pressure in millibars by use of the equation

$$e = \frac{aT}{216700}$$

where a is the absolute humidity in milligrams per cubic metre and T is the temperature in degrees Absolute. Values of corresponding dew-points may then be obtained from hygrometric tables¹³ or from the humidity slide rule.

Curves showing the mean variation of humidity with height for each hour in each month are included in the original paper and those for the months of March, June, September and December are reproduced in Figure 5. Curves for other months show interesting variations from these but it seemed desirable to limit the selection as the curves are seldom directly applicable to forecasting.

Taking the curves as a whole the most obvious feature is the rapid decrease of the vertical humidity gradient with height; the largest variations occur in the lower layers which are nearest to the boundary surface between earth and atmosphere. In the upper layers there is a fairly constant mean gradient, generally a lapse. Around the middle of the day there is a humidity lapse near the surface in all months and this lapse is much greater in summer than in winter. In the later afternoon this lapse decreases and is replaced by an "inversion" of humidity which persists until dawn the next day. After dawn the inversion is broken down and a lapse is re-established. This sequence is common to all mean monthly curves. The periods of the day during which inversions or lapses exist varies with the season of the year in a more or less orderly manner. The variations of the magnitude and the height of the inversions are not so orderly. In February and June the inversion is not well marked (in the mean curves) and the lapses reach their greatest values in July and August. Best, Knighting, Pedlow and Stormonth draw attention to a feature of the curves for October, November and December. During that part of the day when there is a lapse near the surface the curves show a different structure in the middle layer (15.2 to 47.2 metres, that is, about 50 to 156 feet) from that of other months: there is an inversion, or greatly reduced lapse of humidity, above the lapse rate near the surface. There is no parallel feature in the records for temperatures at the same heights and the authors offer no physical explanation.

Statistics are also available showing the diurnal variation of humidity at each of the four heights for each month. Curves for the lowest and top heights

Humidity

(1.1 and 106.7 metres) for only four months are reproduced in Figure 6 as they indicate generally the time and magnitude of the diurnal change. The curves for other heights and other months show a broad similarity. It is interesting to note that the times of maximum humidity show a slight lag with increasing height up to 100 metres but that this lag does not seem to exceed 30 minutes. In the mean, the actual times of maximum humidity vary during the year from about 0730 to 1300 G.M.T.

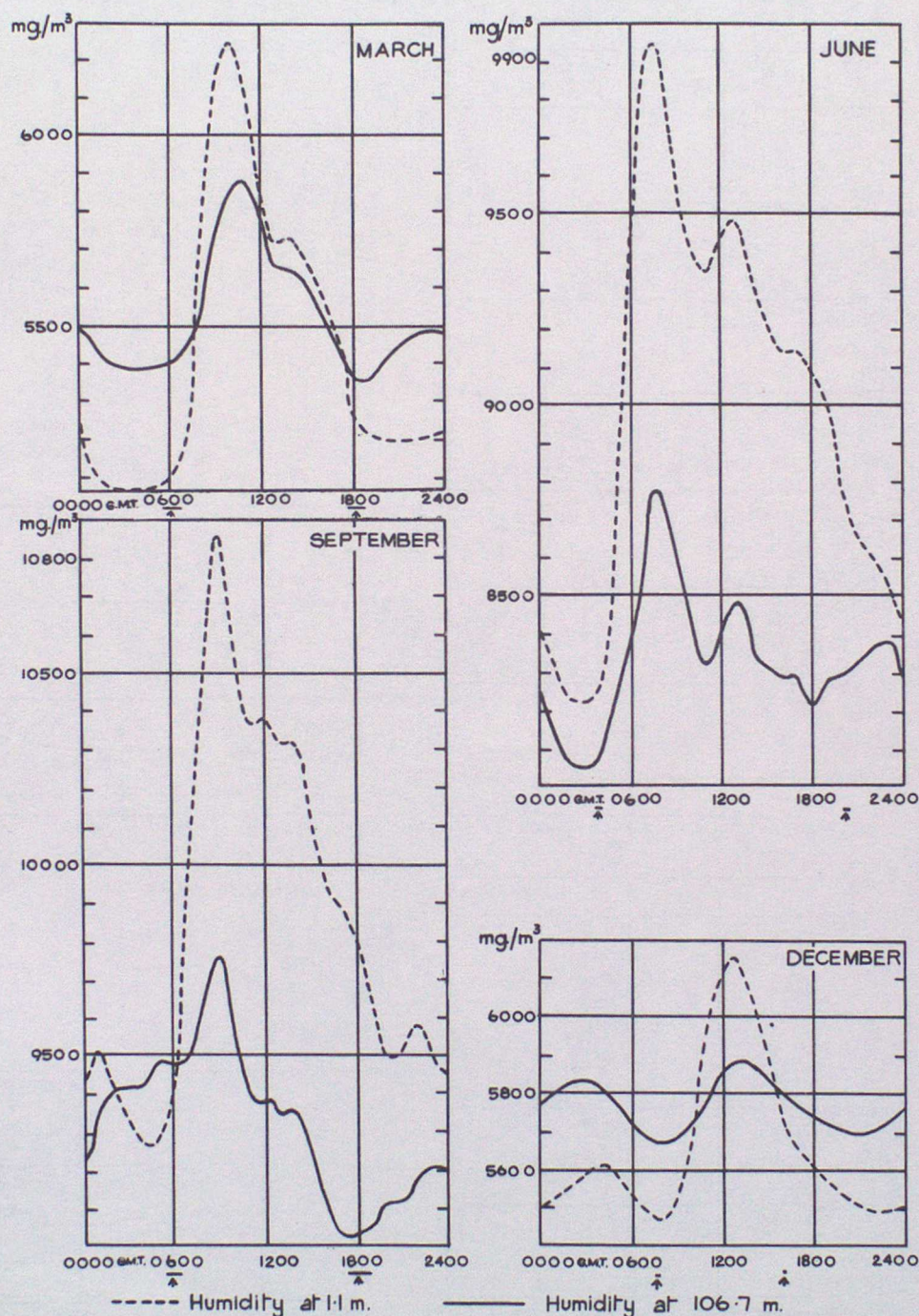


FIGURE 6 Diurnal variation of humidity at 1.1 and 106.7 metres for March, June, September and December at Rye

The arrows indicate times of sunrise and sunset, the limits for each month being shown by the length of the bar immediately above the arrow.

Handbook of Weather Forecasting

From the data obtained during the three-year period of observation, mean diurnal variations were obtained for clear and overcast days in summer and winter. These values give an indication of probable maximum and minimum ranges in summer and winter. Curves for 1.1 and 106.7 metres are reproduced in Figure 7.

On clear summer days the maximum humidity at 1.1 metres is about 12,600 mg.m^{-3} occurring at 0800 G.M.T., the corresponding maximum for all states of sky

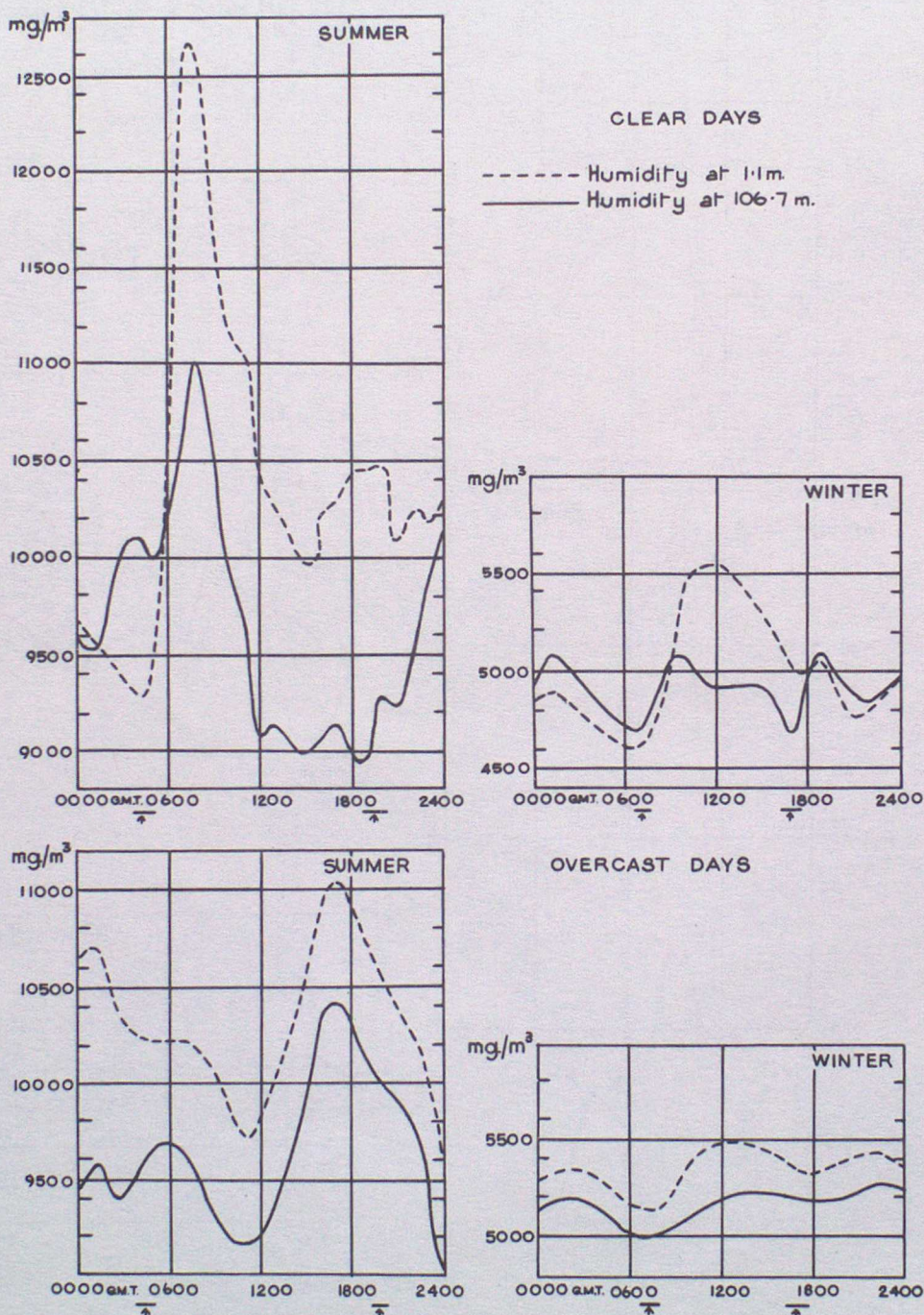


FIGURE 7 Diurnal variation of humidity at 1.1 and 106.7 metres for clear and overcast days in winter and summer at Rye

The arrows indicate times of sunrise and sunset, the limits for each graph being shown by the length of the bar immediately above the arrow.

Humidity

being 11,690 mg.m.⁻³ in August. On overcast days the maximum is about 11,000 mg.m.⁻³ at 1700 G.M.T. On clear days there is no second maximum of humidity in the early afternoon (see June curve of Figure 6) and this shows that on clear days the increased mechanical turbulence and convection are more effective in transporting moisture upwards. On overcast winter days there is a small range of humidity with a flat minimum about dawn and a flat maximum a little after midday. The graphs for overcast summer days and clear winter days are based on only four and three occasions respectively and the curves show some erratic features which are probably due to the smallness of the sample.

The diurnal range of humidity at 1.1 and 106.7 metres for each month is shown in Table 1. For practical use the approximate range in dew-points has been included. Values for clear and overcast days in both summer and winter have also been included. It should be noted that the maximum range at 1.1 metres occurs in the mean in August with a value of about 8°F., corresponding closely to the mean for clear days in summer at the same height. The range at 1.1 metres for clear days in winter is 6°F. which is approximately twice the mean for all days in winter months.

TABLE 1 *Diurnal range of humidity at 1.1 and 106.7 metres at Rye*

	1.1 metres (3 ft. 6 in.)		106.7 metres (330 ft.)	
	Humidity	Approx. range in dew-point	Humidity	Approx. range in dew-point
	mg.m. ⁻³	°F.	mg.m. ⁻³	°F.
January	610	3	560	2½
February	610	3	470	2½
March	1,190	6	520	3
April	1,200	5½	710	3
May	1,390	6	690	2½-3
June	1,700	6	720	2-2½
July	1,940	5	790	2-2½
August	2,900	8	1,120	3
September	1,580	4	650	2
October	1,440	6½	480	1½
November	890	3½	380	1½
December	660	3	210	1
SUMMER				
Clear days	3,310	8-8½	2,070	6
Overcast days	1,350	3½	1,300	4
WINTER				
Clear days	1,000	6	440	2-2½
Overcast days	370	2	280	1-1½
Snow cover	410	3	300	3

Table 2 shows for each month the approximate times of the day at which primary (and secondary, if present) maxima and minima of humidity at heights up to 100 metres occurred in the mean at Rye. There is a distinct minimum in the dawn period when the temperature curves also show a minimum. After dawn the humidity rises sharply and the primary maximum occurs about midday in mid-winter. As the season advances the time of maximum becomes progressively earlier and in midsummer is about 0700-0800 G.M.T. except for the month of July. In July the secondary maximum which is noticeable shortly after midday in

*Handbook of Weather Forecasting*TABLE 2 *Times of occurrence of primary and secondary maxima and minima of humidity at heights up to 100 metres at Rye*

	Time of			
	Primary max.	Secondary max.	Primary min.	Secondary min.
	G.M.T.			
January	1300-1400	2300-2400	0800-0900	1700
February	1200-1300	not pronounced	0001	0700-0800
March	1000-1100	2200-2300	0300-0600	1900
April	0800-0900	2200	0400-0500	1800-2000
May	0700-0800	1200-1300	1800-2000	{ 0300-0500 }
June	0700-0800	1200-1300	0200-0400	{ 1000-1100 }
July	1400-1500	0800	0100-0400	1100
August	0800	not pronounced	1500-1600	not pronounced
September	0800	not pronounced	1700-1900	0400
October	0900-1000	not pronounced	0600	1800
November	1000-1100	not pronounced	2100	{ 1700 }
December	1200-1300	0300-0500	0700-0900	{ 0100-0700 }
				2100

May and June becomes the primary but there is still a fairly well marked secondary maximum in July at 0800 G.M.T. Best, Knighting, Pedlow and Stormonth consider that:

"The explanation of the double maxima lies in the role that turbulence plays in redistributing the moisture content of the air. After dawn there is evaporation at the surface owing to the temperature rise caused by insolation, and the humidity rises at all levels, the diffusing agency being slight turbulence. As the morning proceeds turbulence increases and carries away moisture from the lower atmosphere rather more quickly than it is supplied by evaporation, so that the net effect is a decrease in humidity. At midday the evaporation is a maximum at the surface, counter-balancing the loss of water vapour carried upward by turbulence, and shortly after midday there is a second maximum. In the afternoon both evaporation and turbulence decrease, but the turbulence acts through a deeper layer than that in which the readings were taken and the humidity decreases."

The amount of detail in Section 15.5.1 is much greater than the detail with which a forecaster can make a routine analysis. Further, some of the changes observed at Rye must be due to the characteristics of the particular site. It is clear, therefore, that straightforward practical application of the values at Rye to other sites is neither possible nor justifiable. Nevertheless an understanding of the values observed at Rye should contribute towards routine analyses and forecasts of humidity in the lower atmosphere which are physically sound.

15.5.3. Humidity in air masses

Belasco⁴⁴ has computed some characteristics of various air masses which affect the British Isles and an account of part of this work was given in Section 14.8.1. Table 14 in that section contains mean vapour pressures at the 700-millibar level of some sixteen air masses and Table 15 contains mean vapour pressures at 50-millibar intervals from 950 to 450 millibars in summer (July and August) and winter (January and February), also for sixteen air masses. Tephigrams of average temperature and humidities in winter and summer of some eight air masses of polar and tropical origin in the neighbourhood of the British Isles have been included in Figures 15, 16 and 17 of Section 14.8.1.

Humidity

These tables and figures also indicate the change in humidity which occurs in the mean at various pressure levels in air masses as they are moved quasi-horizontally from their source region towards the British Isles by the large-scale flow patterns of major pressure systems. The modifications are in general accord with what could be expected from a consideration of the physics of the processes taking place. For example, as tropical air moves away from its source, cooling, by contact with the underlying cooler sea surface, reinforces an already stable lapse rate so that the vertical transfer of humidity from the lower layer by mechanical turbulence is effectively damped by this stable lapse rate and there is only a slight change in humidity. In contrast, as polar air moves from its source region it generally passes over seas which become progressively warmer as the air approaches the British Isles. Convection is widespread and vigorous and as the air moves over still warmer seas convection is maintained. This convection takes place through a considerable vertical extent and, in spite of the removal of some water from the atmosphere by showers, it is a powerful mechanism for distributing water vapour upwards through the troposphere. The changes in humidities (and temperatures) of various air masses can be seen from Table 14 and Figures 14 and 15 of Section 14.8.1. The pronounced increase in vapour pressures in polar air masses P_1 and P_7 , compared with those in P_{IC} in its source region contrast strongly with the minor modifications in tropical air T_A as it moves to the British Isles as tropical air T_2 (that is, on an anticyclonic path from south-west of the Azores). Belasco found that in polar air at all heights water-vapour concentration increases with the age of the polar air and that the increase in water-vapour content is much greater at low than at high levels. Thus in summer between polar air P_1 (northerly airstream direct from polar regions) and P_7 (long sea track over the Atlantic; approaching the British Isles on a track of about south-west) there is an increase in mixing ratios from 6.4 to 9.4 grammes per kilogram at 950 millibars and from 0.95 to 1.4 grammes per kilogram at 500 millibars.

Average and extreme vapour pressures at screen level at Kew for all air masses defined by Belasco (see Section 14.8.1 for details) are given in Table 3.

Mean values quoted by Belasco should not be applied directly to forecasting without consideration of the latest observations. Modifications can be introduced to allow for the physical processes taking place, for example, advection, convection, frontal up-slope and precipitation.

Additional statistical information on humidities in the upper air over the United Kingdom may be obtained from Meteorological Office upper air publications.¹⁵⁻¹⁶

15.5.4. *Humidity in frontal regions*

The idealized concept of abrupt changes in air-mass characteristics through frontal surfaces is useful for analysis and forecasting, but detailed investigation of fronts shows that the changes which actually occur are often much more complicated. That this is particularly true for humidity is very clearly shown from a study made by Sawyer.⁵ Between November 1950 and July 1952 the Meteorological Research Flight made 23 flights to study the detailed structure of the atmosphere in the neighbourhood of fronts in the lower or middle troposphere. The outward and inward leg of each flight was normally flown at a different level and the levels varied from flight to flight. In this series of flights the lowest leg flown was at 940 millibars and the highest at 450 millibars. Although the aircraft observations of humidity were not sufficient to establish the complete horizontal and vertical distribution of humidity which accompanied fronts, it was found

Handbook of Weather Forecasting

TABLE 3 Average daily mean and extreme daily maximum and minimum vapour pressure* near the surface at Kew in summer and winter

Daily vapour pressure	Air mass													All air masses combined, 1886-1915						
	T ₁₁	T ₂	T ₃	T ₄	T _Q	H _O	H _{NE}	H _{SE}	H _{SW}	H _{NW}	A ₁	A ₂	P ₁	P ₂	P ₃	P ₄	P ₅	P ₆	P ₇	
millibars																				
Summer																				
Average	17.1	17.4	18.0	18.1	14.7	14.1	11.8	15.1	16.9	13.6	11.4	11.2	13.4	13.2	14.9	15.0	15.6	13.7
Extreme maximum	20.1	19.8	21.7	22.2	16.2	17.6	14.8	17.3	19.4	15.5	13.1	13.1	15.3	14.6	17.5	16.8	17.7	
Extreme minimum	15.4	15.5	14.2	14.9	12.4	10.7	10.9	12.6	13.8	11.6	9.5	9.4	11.4	11.9	12.8	13.3	13.7	
Winter																				
Average	10.9	10.3	9.6	8.6	8.9	7.0	5.1	6.2	8.4	6.4	4.1	3.9	4.9	4.7	6.3	6.3	7.6	7.7	9.2	6.8
Extreme maximum	12.4	12.0	11.1	10.4	10.5	9.1	6.0	7.8	9.8	7.8	5.3	5.0	5.8	6.0	7.5	7.4	8.9	9.4	10.6	
Extreme minimum	9.4	8.3	7.6	6.7	7.3	5.5	4.0	4.8	6.9	5.1	2.7	3.0	3.9	3.0	5.4	5.3	6.3	6.2	7.9	

*Computed from the daily mean temperature and mean relative humidity.

Humidity

possible to use the dew-points reported by the routine radio-sonde ascents to prepare a vertical dew-point cross-section through each front. Such sections did not show the complex and irregular distribution of water vapour which the aircraft observations demonstrate, but comparison between the cross-sections and the aircraft observations of humidity showed that the cross-sections were sufficiently accurate to indicate the broad features of the humidity pattern.

A feature, more or less well developed on all the dew-point cross-sections, was a tongue of relatively dry air extending downwards in the vicinity of the front and tilted in the same direction as the front (see Figure 8). The dry tongue was

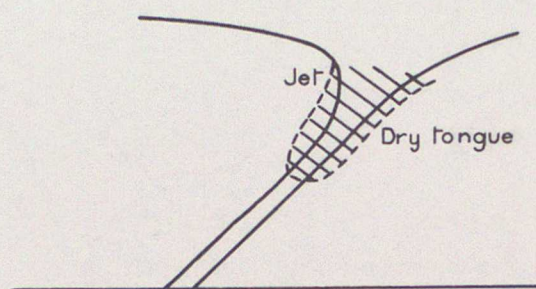


FIGURE 8 *Schematic illustration of a dry tongue associated with a front*

less well developed in association with the cold fronts than with the warm fronts; it extended down to an average pressure level of 700 mb. in cold fronts and 800 mb. in warm fronts. In about half the fronts the driest air was found within the frontal transition zone itself, but there were also occasions when it was found on either the cold or warm sides of the transition zone. Some striking changes of frost point occurred as the aircraft entered or left the dry tongue; a good example, that of 4 January 1952, is reproduced in Figure 9. About half the flights showed

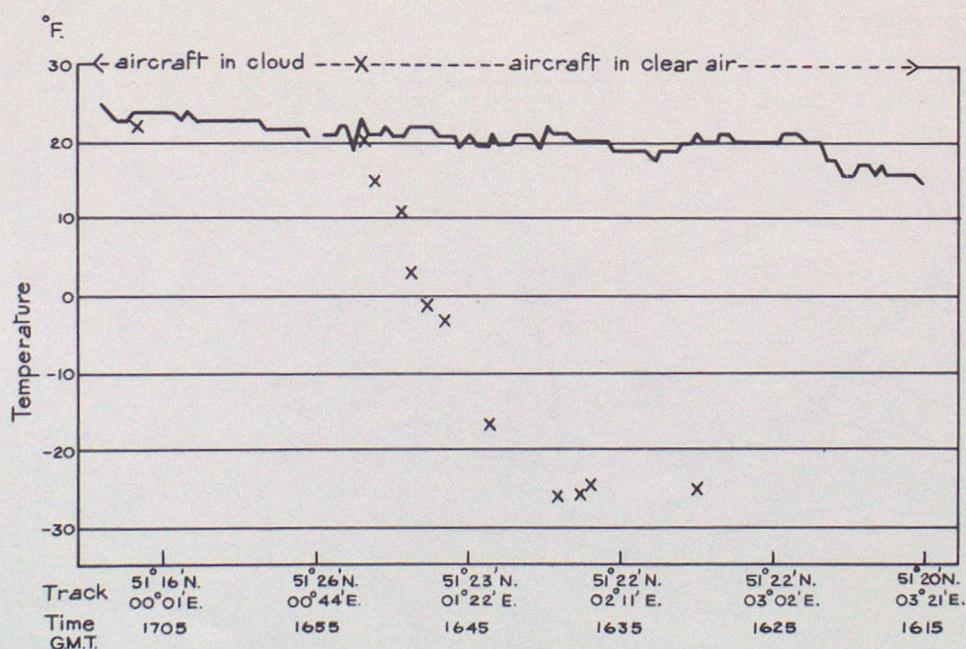
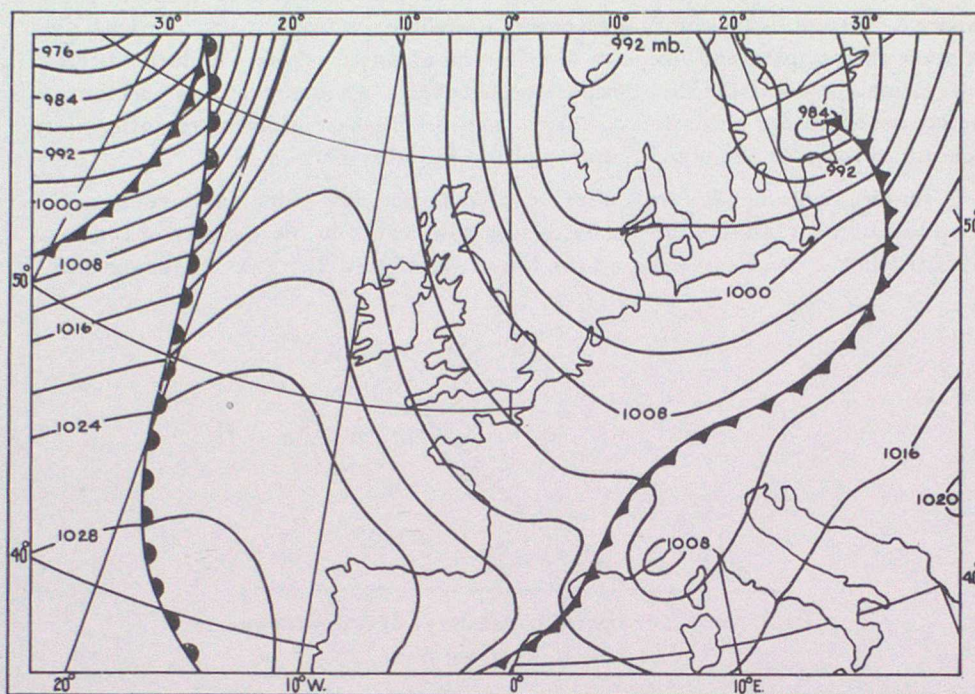
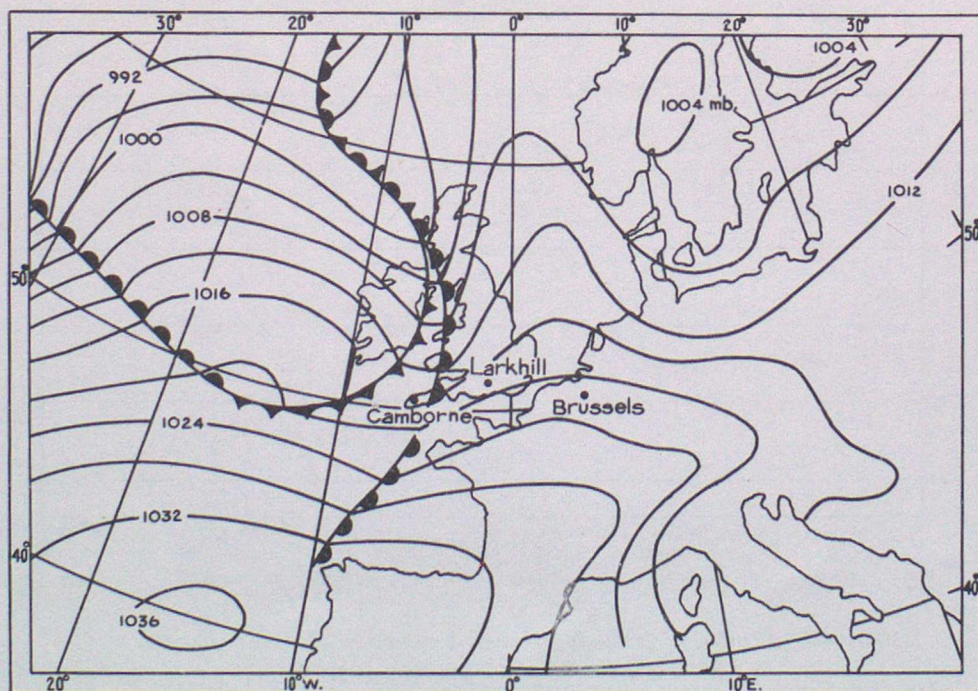


FIGURE 9 *Profile of temperature and frost-point at 800 millibars, 1615-1710 G.M.T., 4 January 1952*

Track: 51°20' N. 3°21' E. to Farnborough
Temperature is indicated by a continuous curve, frost-point by crosses.

Handbook of Weather ForecastingFIGURE 10 *Surface chart, 1500 G.M.T., 3 January 1952*FIGURE 11 *Surface chart, 1500 G.M.T., 4 January 1952*

Humidity

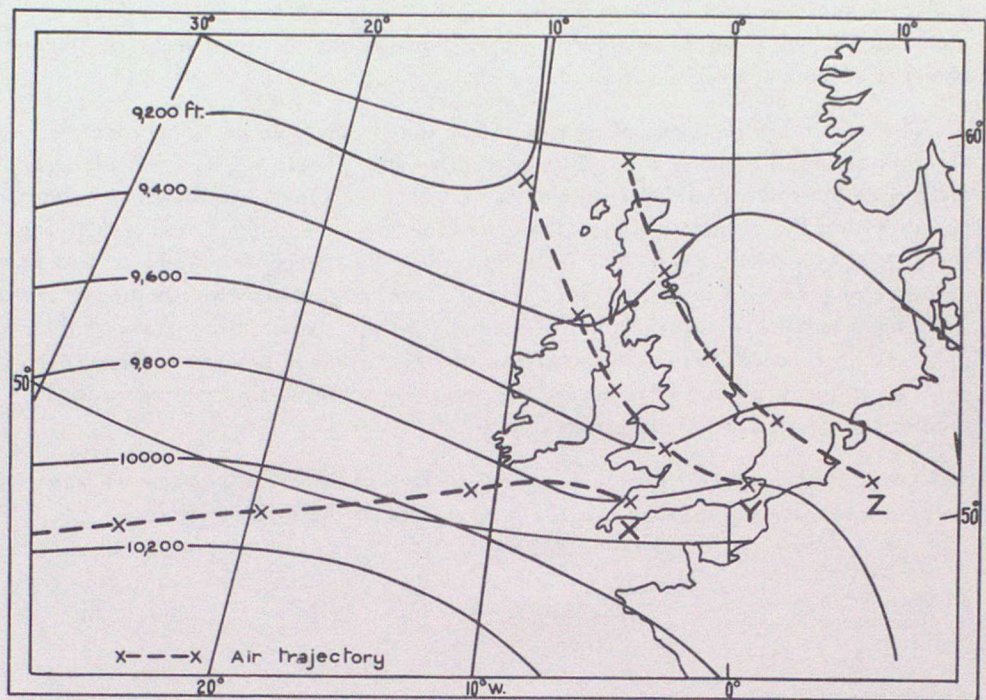


FIGURE 12 700-millibar contour chart, 1500 G.M.T., 4 January 1952 and 24-hour trajectories at 700 millibars terminating at that time

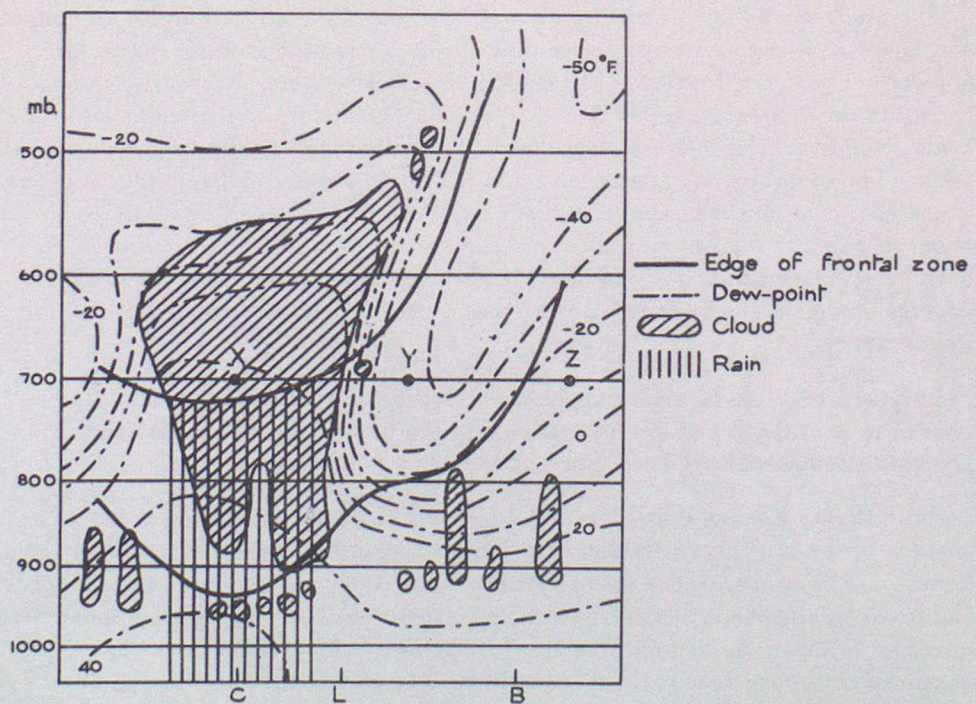


FIGURE 13 Cross-section, 1500 G.M.T., 4 January 1952
C = Camborne, L = Larkhill, B = Brussels

Handbook of Weather Forecasting

a sharp transition from moist to dry air and the change of frost-point averaged about 20.5°C. in 35 miles on these flights. Some of the sharpest transitions gave changes of more than 22°C. in 20 miles.

The very rapid increase of frost-point as the (frontal) cloud is approached should be noted in Figure 9; this is not uncommon. Table 4 gives the average difference between the temperature at the edge of the cloud and the frost-point at points within the clear air distant from the cloud by $\frac{1}{2}$, 1, $1\frac{1}{2}$. . . minutes flying time. It shows that the humidity falls off rapidly away from the cloud so that at a distance of less than ten miles the average depression of the frost-point is 5.5°C.; it changes only slowly with further distance from the cloud. From these results it seems improbable that, in the analysis of free-air humidity observations, extrapolation of humidity and temperature observations to saturation can accurately predict the position of cloud masses.

TABLE 4 *Average difference between frost-point at various distances from the edge of a cloud and the temperature at the cloud boundary*

Flying time from cloud edge	minutes									
	$\frac{1}{2}$	1	$1\frac{1}{2}$	2	$2\frac{1}{2}$	3	$3\frac{1}{2}$	4	$4\frac{1}{2}$	5
Approximate distance from cloud edge	miles									
	2	4	5	7	9	11	13	15	16	18
Temperature at cloud edge minus frost- point in clear air	degrees Celsius									
	3.3	4.3	4.6	5.0	5.9	7.2	7.2	6.7	6.4	7.4
Temperature at cloud edge minus temper- ature in clear air	-0.16	-0.33	-0.33	-0.39	-0.44	-0.44	-0.44	-0.39	-0.39	-0.44

To study the source of the dry air, trajectories of the air within the dry tongue and of the adjacent moister air were drawn from the routine contour charts for appropriate pressure levels. Although the trajectories were necessarily rather uncertain the analysis showed that the proximity of the dry to the moist air in the vicinity of the front can be reasonably attributed primarily to large-scale pseudo horizontal air movements and that the air in the dry tongue is dry because it has subsided, but the subsidence may have taken place some 24 hours or so before it reached the frontal region. In some warm fronts the dry air seemed to have been derived from the subsided air of the preceding ridge and not to have undergone any recent large vertical movement when found in or just ahead of the frontal zone.

Figures 10-13 show one example discussed by Sawyer which clearly illustrates the dry tongue and the fact that air in the frontal zone may, over the preceding 24 hours, have been drawn from widely separated areas.

The flights examined by Sawyer did not extend above 450 millibars but in another series of flights over southern England by aircraft of the Meteorological Research Flight, measurements of humidity were made in the upper troposphere and lower stratosphere. On seventeen occasions a frontal surface was penetrated at or above the 450-millibar level. Bannon, Frith and Shellard¹⁹ have examined these and found that there is often a large decrease in frost-point depression below air temperature in passing upwards through a frontal surface. They found a mean decrease in depression through the front (or through about 50 millibars) of some 8.2°C. On the seventeen flights the minimum change was zero and the maximum was 17°C.

Humidity

They also drew attention to the existence of very dry air just beneath the frontal surfaces in six of the seventeen cases. The frost-point appeared to be abnormally low just below the frontal surface, compared with the general trend of frost-point distribution with height in the lower air mass. In each of these cases the ascent was made in the vicinity of a jet stream. A typical example is shown in Figure 14. It appears that the dry tongue at lower levels may on occasions extend upwards above the 450-millibar level as indicated in Figure 8.

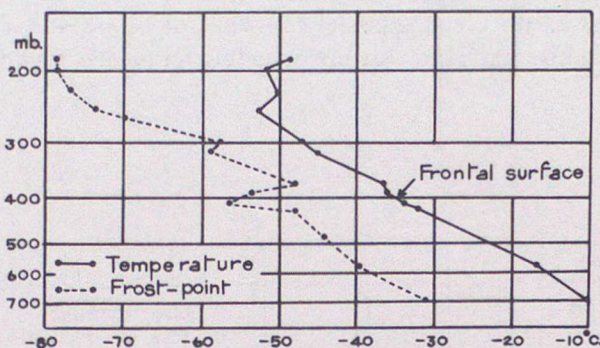


FIGURE 14 Aircraft sounding, 1030 G.M.T., 26 April 1949

15.5.5. Humidity in the upper troposphere and lower stratosphere

It was seen in Section 15.5.1 that the humidity elements commonly in use in radio-sondes are unreliable at the temperatures and humidities which normally occur at upper tropospheric and lower stratospheric levels. Consequently the practical forecaster normally has no reliable direct measurement of humidity on which to make an assessment of the distribution of humidity at these levels. The following sections should assist him in making reasonable estimates.

15.5.5.1. Humidities in and around high (cirrus) cloud. According to Ludlam²⁰ a humidity of 100 per cent with respect to water (that is, considerable supersaturation with respect to ice) is necessary before cirrus clouds begin to form. Once ice crystals are present they will grow in the ice-supersaturated region and the humidity in that region will fall to something like 100 per cent with respect to ice. (The physics of clouds are more fully discussed in Chapter 16). Observations by the Meteorological Research Flight which have been analysed by Murgatroyd and Goldsmith²¹ showed, however, that there was, on average, a frost-point depression of some 1.7°C. in cirrus cloud. This indicates a relative humidity of about 80 per cent with respect to ice so that there is an apparent discrepancy between theory and observation. However, the formation of cirrus clouds is often restricted to limited regions which may not be adequately represented by averages. Also many cirrus clouds are in process of dissipation because the ice crystals, once formed, tend to fall into drier layers where they slowly evaporate. Further, the Meteorological Research Flight observations were made with hygrometers which did not respond to short-period fluctuations so that each observation was a mean value over several miles. In view of these factors the discrepancy is probably more apparent than real but it is evident that close estimates of the humidity in cirrus clouds cannot be readily deduced.

15.5.5.2. Humidity distribution when high cloud is present - the "moist layer". Murgatroyd and Goldsmith²¹ drew attention to the type of humidity distribution in the upper troposphere when high cloud was present and to the occurrence of what they called the "moist layer". When high cloud was present upwind of the area of observation the authors state that the distribution of frost-point with height

Handbook of Weather Forecasting

usually exhibited a characteristic form. This form is illustrated in Figure 15(a) which shows the details actually observed at 1430 G.M.T., 12 November 1951. The ascent shows a rapid increase of humidity at 28,000 feet with a moist layer persisting to 32,000 feet - the height of the tropopause on that day. This "moist layer" became more pronounced the nearer the sounding area was to the cirrus cloud until, when the cloud was present near the aircraft, the sounding often took the form illustrated in Figure 15(b). The authors stated that the moist layer was not always easy to trace and, within it, the depression of the frost-point relative to the dry-bulb temperature might be as much as 8° to 11°C. , but the depression quickly decreased as the cloud approached. When no cirrus was observed Murgatroyd and Goldsmith found that the sounding frequently took the form shown in Figure 15(c).

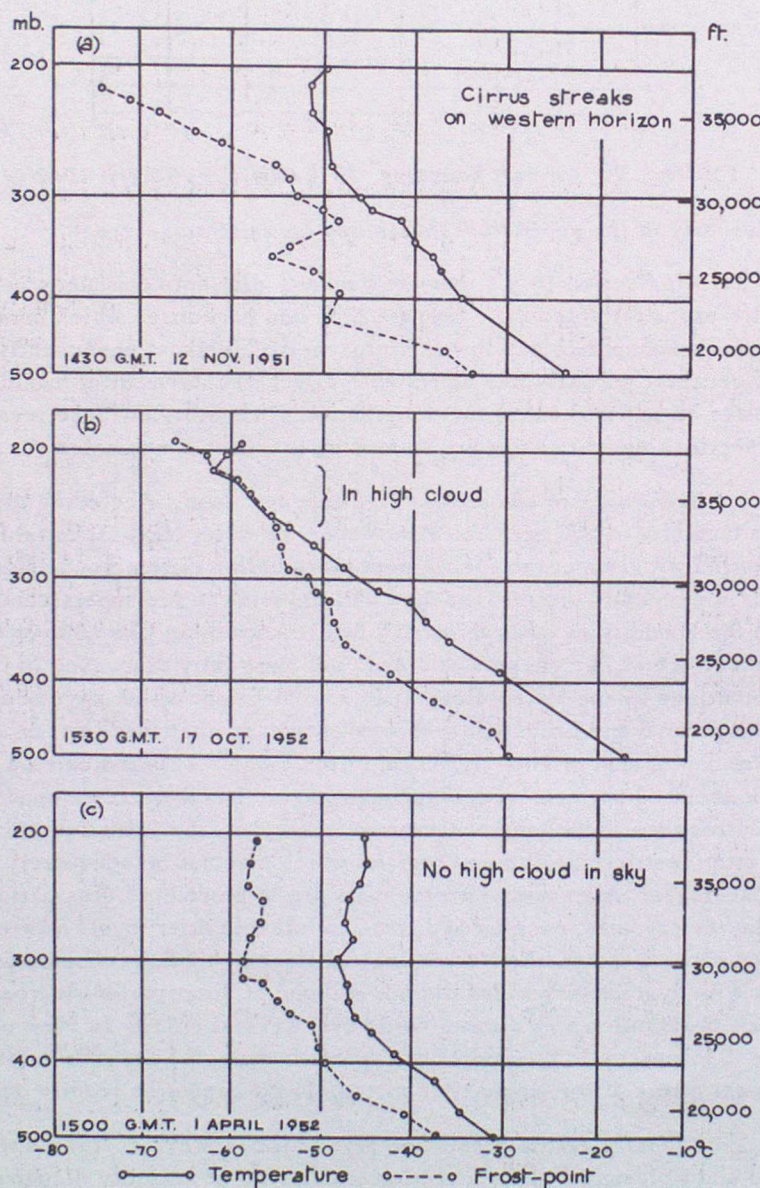


FIGURE 15 Aircraft soundings illustrating the "moist-layer" effect

Humidity

15.5.5.3. *Statistical information on humidity.* For upper tropospheric and lower stratospheric levels statistical information has not yet been published showing distributions of humidity in relation to synoptic types, upper air circulation patterns or other meteorological parameters. Two accounts of the general distribution over southern England have, however, been presented. The first was by Bannon, Frith and Shellard¹⁹ who examined humidity observations made from aircraft between 1943 and 1950. The operational ceiling of the aircraft available at the time imposed some bias towards low tropopauses. Further observations made from aircraft of improved performance have removed this bias and have been studied by Murgatroyd, Goldsmith and Hollings.²² The account which follows is a composite of both these papers. In the following treatment the frost-point is used to indicate the humidity and the ascents are shown with the pressure at the tropopause, P_c , as the datum level. Thus vertical distances from the tropopause level are expressed in pressures, that is, $(P_c - 25)$ millibars is at a height above the tropopause where the pressure is 25 millibars less than P_c and conversely $(P_c + 25)$ millibars is below the tropopause.

Examination of individual ascents showed that there was frequently a sharp fall of frost-point on entering the stratosphere. This rapid fall of frost-point does not always occur and there are many ascents in which the lapse rate of frost-point is not materially different above and below the tropopause. Typical examples of both types are shown in Figures 16 and 17.

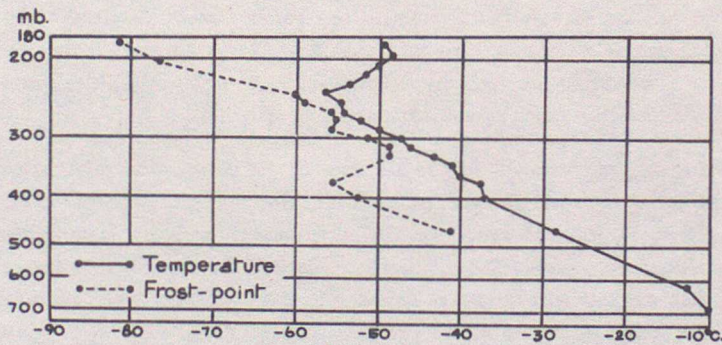


FIGURE 16 Aircraft sounding, 1500 G.M.T., 24 May 1948

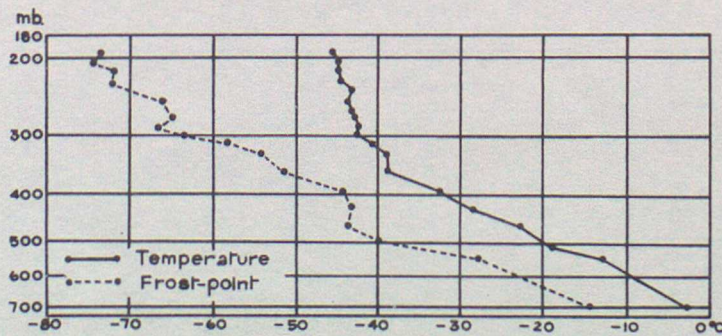


FIGURE 17 Aircraft sounding, 0930 G.M.T., 5 July 1949

If the aircraft ascents are separated broadly into these two types it is found that those showing the sharp discontinuity in the lapse of frost-point are more often than not associated with a tropopause showing an abrupt change from lapse to inversion of temperature. Conversely those ascents showing no marked discontinuity in the lapse of frost-point are more often than not associated with a tropopause which is not characterized by an inversion of temperature. On some of the occasions showing the marked frost-point discontinuity it was noted that the discontinuity did not always coincide with the level of the tropopause; sometimes it was higher and sometimes lower.

Within the vertical limit of the earlier ascents (about 40,000 feet) it was found that the frost-point in the stratosphere almost invariably decreased with increasing height. These frost-points were generally within the range -80° to -120°F . (-62° to -84°C .) and relative humidity in the stratosphere was almost always very low. It should be noted that these low frost-points at upper levels correspond to minute quantities of water vapour as indicated by the values given below.

Frost-point	Pressure	Humidity mixing ratio
$^{\circ}\text{C}$.	mb.	g.kg.^{-1}
-60	300	0.0224
-70	200	0.0081
-80	100	0.0033

Thus the low frost-points frequently observed at lower stratospheric levels represent only a few milligrams of water vapour per kilogram of dry air.

The bulk of published results of humidities in the lower stratosphere shows that the air is normally very dry. An indication of this extreme dryness and also of the variation of humidity across the tropopause is given in Table 5, compiled by Bannon, Frith and Shellard,¹⁹ who comment:

"The outstanding features of this table are (i) the rapid decrease in relative humidity as the stratosphere is entered, and (ii) the fact that the relative humidity continues to decrease, though less rapidly, the further into the stratosphere one goes. Mean values are given for each level. Ten of the ascents penetrated 100 mb. into the stratosphere, the mean relative humidity at that level being only 2.2 per cent, and the range 0.8 to 3.9 per cent."

TABLE 5 Frequency of relative humidity (with respect to ice)

Relative humidity																	Mean	Total
	0-	1-	3-	5-	10-	15-	20-	25-	30-	40-	50-	60-	70-	80-	90-	100	relative	no. of
	0.9	2.9	4.9	9.9	14	19	24	29	39	49	59	69	79	89	99		humidity	cases
Number of occasions																		
mb.																	%	
$P_c - 100$	1	7	2	2.2	10
$P_c - 75$	2	17	5	4	2.7	28
$P_c - 50$	0	25	11	14	3	3	5.0	56
$P_c - 25$	0	6	10	26	16	16	10	5	2	0	0	1	12.9	92
Tropopause	0	2	1	8	4	2	12	7	22	19	8	4	9	2	3	3	41.5	106
$P_c + 25$	0	1	0	3	5	4	7	5	25	15	13	9	5	4	3	2	44.3	101
$P_c + 50$	0	0	0	4	1	6	9	8	24	18	10	7	8	2	2	1	42.6	100

Humidity

It should be mentioned, however, that there are a few observations which indicate that relatively moist layers may exist in the stratosphere. Bannon, Frith and Shellard¹⁹ record two occasions (on successive days) when the comparatively high relative humidities of 26 and 75 per cent (with respect to ice) were observed at the 200-millibar level which, in each case, was well within the stratosphere. These results were completely at variance with all others obtained in the series of flights. Although the two observations might be correct Bannon *et alii* considered that there was a possibility that moister air from the cabin of the aircraft had leaked into the hygrometer and so vitiated the readings. Barrett, Herndon and Carter²³ also have described observations of moist layers in the stratosphere obtained on three high-altitude ascents of an experimental radio-sonde incorporating an electronic frost-point hygrometer. The apparatus was carried aloft by a large plastic balloon which reached heights of about 30 kilometres (pressures 09 to 16 millibars). On the three individual ascents a shallow layer, saturated with respect to ice, was reported (namely: at 106 millibars, between 130 and 115 millibars and between 135 and 125 millibars). No published account of additional ascents using the American apparatus has been traced. For operational reasons the three ascents described by Barrett *et alii* were all made in anticyclonic conditions of light surface winds and small cloudiness some one to three days after the passage of a cold front. The two flights on which moist air was observed in the series of flights analysed by Bannon *et alii* were made on successive days (19 and 20 January) also behind a cold front which had moved rapidly south-east and cleared the British Isles early on 18 January 1948. The broad similarity of the post cold-frontal synoptic conditions when the moist stratospheric layers were observed may be significant or coincidental.

These few reports of moist stratospheric layers are in such direct conflict with the increasing number of observations by the Meteorological Research Flight of very dry air in the stratosphere that it seems almost impossible to reconcile the two sets of results. At the time of writing the matter has not been finally resolved.

Graphs showing mean ascent curves of temperature and frost-point from 100 millibars below to 50 millibars above the tropopause are shown in Figure 18, taken from *Geophysical Memoir No. 88*.¹⁹

Mean ascent curves for those occasions on which the tropopause was or was not characterized by an inversion of temperature are given in Figure 19, also taken from *Geophysical Memoir No. 88*. The number of observations at the various levels is as follows:

	P_c-50	P_c-25	Tropopause	P_c+25	P_c+50	P_c+75	P_c+100
Tropopause with temperature inversion	34	61	61	61	60	59	59
Tropopause without temperature inversion	17	24	24	24	24	24	24

The mean curve for the whole year in Figure 18 gives an average fall of frost-point of 4.1°C. in the 25-millibar layer below the tropopause and a fall of 6.3°C. in the 25-millibar layer above the tropopause, that is, there is on the average a steepening of the lapse rate of frost-point as the stratosphere is entered. In Figure 19 the corresponding values are 4.3°C. and 7.3°C. for tropopauses with inversion of temperature and 3.7°C. and 5.9°C. for other tropopauses. Thus a steep stratospheric lapse rate of frost-point tends to be associated with a temperature inversion.

Handbook of Weather Forecasting

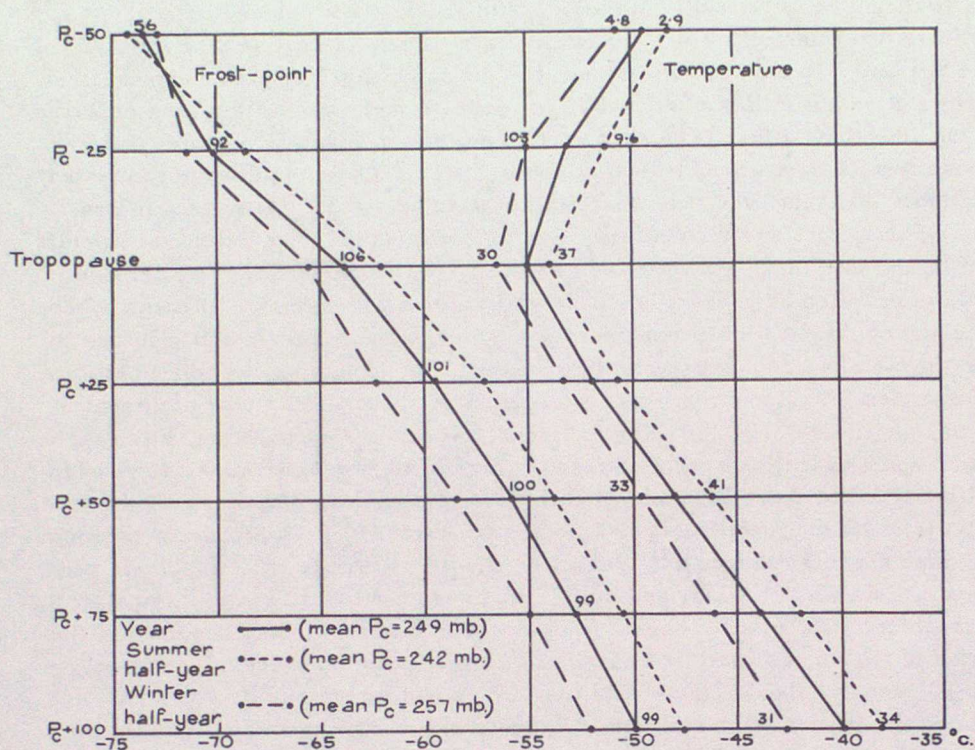


FIGURE 18 Mean ascent curves; summer, winter and year

The figures against the summer and winter temperature curves indicate the corresponding relative humidities with respect to ice. Those against the year frost-point curve are the number of observations available. Summer half-year = 16 May to 15 November; winter half-year = 16 November to 15 May.

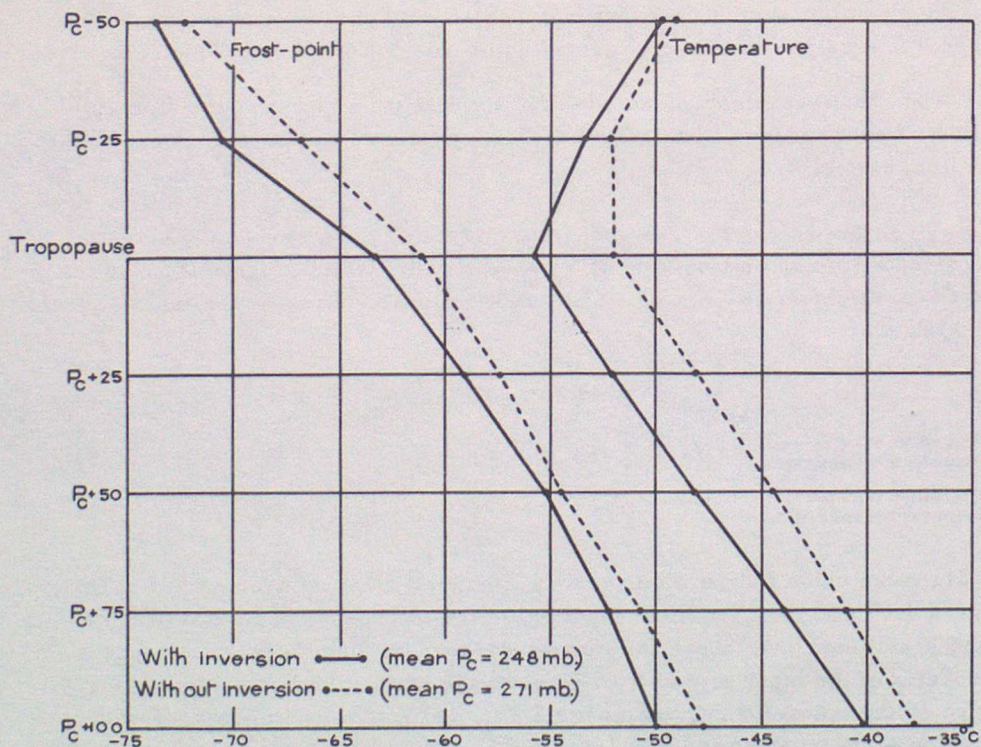


FIGURE 19 Mean ascent curves for tropopauses with and without inversion of temperature

Humidity

Graphs showing broad differences in humidities corresponding with air which had relatively high and low tropopauses were obtained by Bannon, Frith and Shellard¹⁹ and are reproduced in Figure 20. From the aircraft ascents they found 49 occasions for which they had full temperature and frost-point data from a pressure ($P_c + 100$) up to ($P_c - 50$) millibars. The mean tropopause pressure for these ascents was 271 millibars and this was the critical value which they used to separate the low and high tropopauses. From Figure 20 the authors observe:

"The higher mean tropopause is of course colder than the lower one; it also has a lower frost-point. The same applies to corresponding levels in the two stratospheres, and it will be noticed that in terms of relative humidity there is little between the two. At any fixed pressure level however the air is much drier in terms of frost-point if the tropopause is low, and this rule appears to be valid whether the level chosen is in the lower stratosphere or upper troposphere. There is a slight tendency for both temperature and frost-point discontinuities to be sharper at the higher tropopause than the lower one."

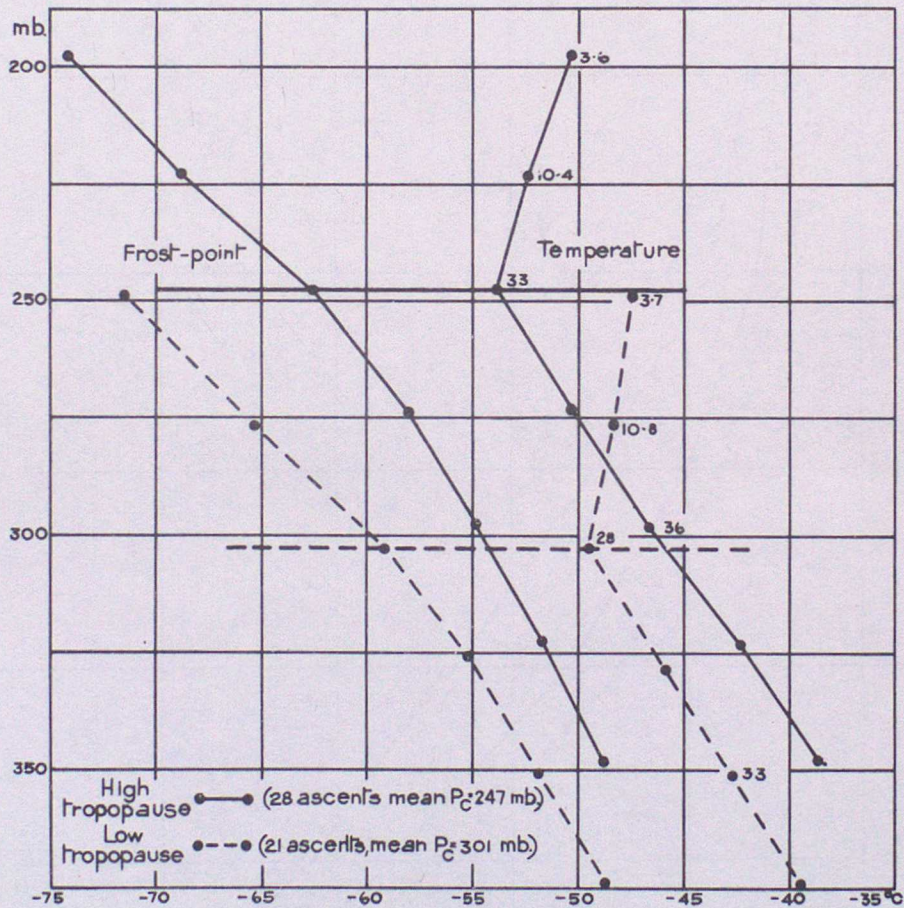


FIGURE 20 Mean ascent curves; selected ascents;
high tropopause and low tropopause.

The figures against the temperature curves indicate the corresponding relative humidity.

From an examination of a few ascents which penetrated some 100 millibars beyond the tropopause (mainly confined to occasions of low tropopause) Bannon, Frith and Shellard concluded that "the frost-point in the stratosphere continues to decrease with height but that the rate of decrease gets less, and that if the frost-point does eventually become constant it probably does so at a value approaching

Handbook of Weather Forecasting

-120°F." The later work of Murgatroyd, Goldsmith and Hollings²² using a Canberra aircraft capable of ascents to 50,000 feet confirms this. The number of flights made for which data has currently been published is still small but the results indicate a pronounced tendency to constant frost-points of about -120°F. (-84°C.) at heights in excess of 100 millibars above the tropopause. Figure 21 shows the

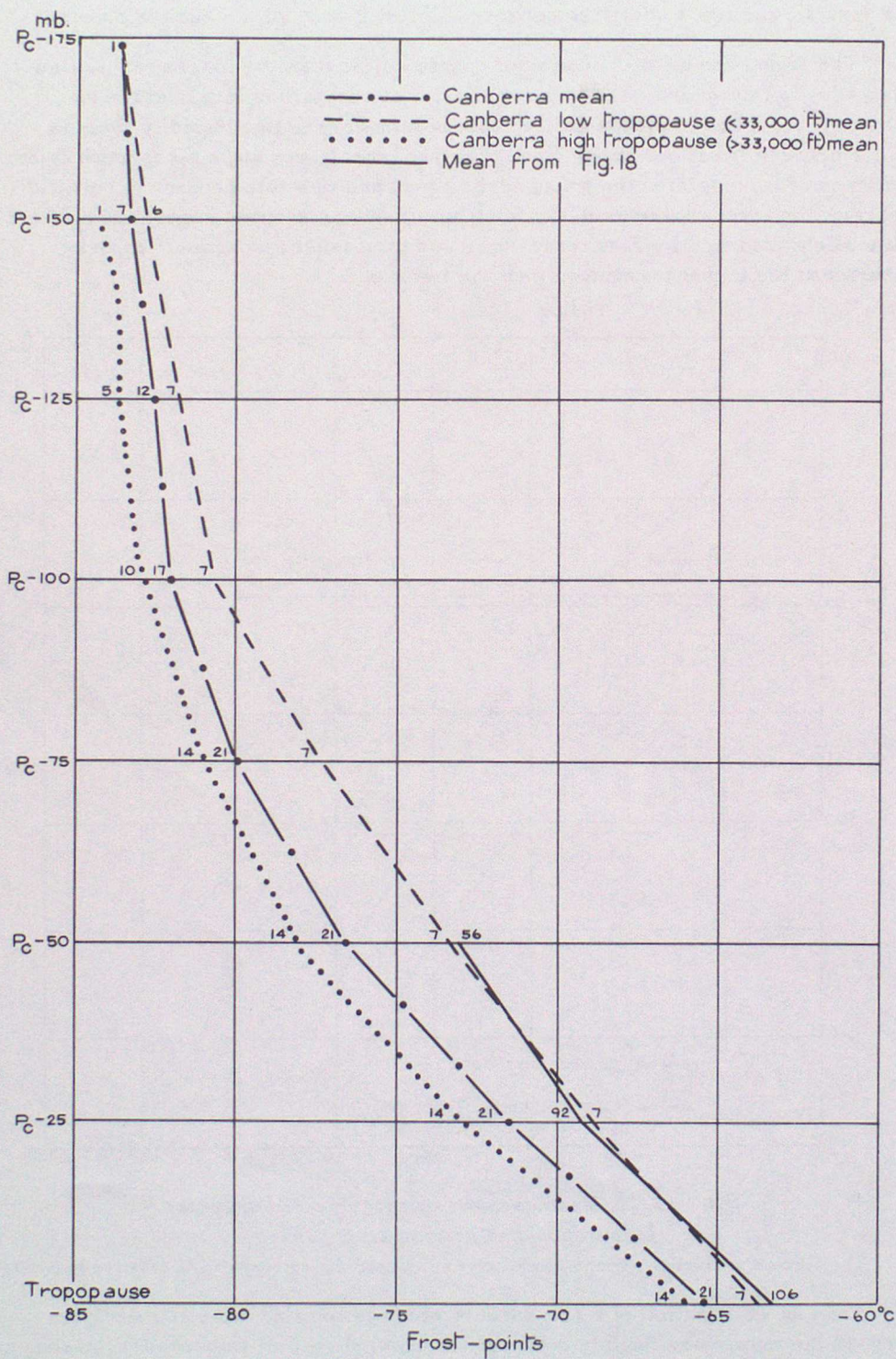


FIGURE 21 *Mean frost-points at levels above the tropopause*
 The figures against the curves give the number of observations available.

mean frost-points at levels above the tropopause as observed by the Canberra aircraft. The mean curve for the year from Figure 18 has been plotted for comparison and it will be seen that there is good agreement with the Canberra means on those occasions of low tropopause (below 33,000 feet). Figure 21 also confirms the conclusion of Bannon, Frith and Shellard that for high tropopauses frost-points in the lower stratosphere are generally lower than those for low tropopauses.

From Figure 21 it will be seen that, on average, there is a lapse of frost-point of about 6°C . in the first 25 millibars above the tropopause and this confirms the value given by Bannon, Frith and Shellard. This sharp lapse of frost-point is not maintained and at about 100 millibars or so above the tropopause the lapse of frost-point is only about 0.5°C . in 25 millibars. The change from a large lapse rate of frost-point near the tropopause to a smaller value at higher levels appeared to be discontinuous in some cases suggesting a "humidity tropopause" a few thousand feet above the temperature tropopause. The Canberra observations were not made at sufficiently small height intervals to enable Murgatroyd, Goldsmith and Hollings to establish whether the discontinuity was usually the case. The distance from the tropopause to the region of small lapse of frost-point varied between 2,000 and 10,000 feet and no relationship between it and other meteorological variables around the tropopause was discovered in the limited series of observations.

Most of the Canberra observations were made near the British Isles in February, March, June and July of 1954 and there appeared to be little variation from late winter to early summer in the frost-points above the tropopause. A few flights which have been made by the Meteorological Research Flight aircraft in the stratosphere from latitudes 35° to 65°N . show that the frost-point varies very little with latitude and appear to indicate the value of about -80°C . around 100 millibars above the tropopause as typical of a wide range of latitude. A limited number of flights made at about 45,000 feet in equatorial regions indicates that the value -80°C . is also typical of frost-points at these altitudes in the equatorial troposphere.

Bannon, Frith and Shellard¹⁹ obtained a number of correlation coefficients between various meteorological variables of the atmosphere. Correlations between frost-points at 200, 250 and 300 millibars (F_{200} , F_{250} , F_{300}) and the pressure at the tropopause (P_c) and also the temperature at 500 millibars (T_{500}) are given below:

	Correlation coefficient	No. of pairs of values
$F_{200}: P_c$	-0.66	87
$F_{250}: P_c$	-0.70	105
$F_{300}: P_c$	-0.66	105
$T_{500}: F_{200}$	+0.65	108
$T_{500}: F_{250}$	+0.83	127
$T_{500}: F_{300}$	+0.80	127

All these correlations are significant. The relationship between frost-points at 200, 250 and 300 millibars and the pressure at the tropopause are shown in Figure 22 which shows the lines of best fit between the observations of frost-points at the three levels and tropopause pressure. If the pressure at the tropopause is known it is possible from Figure 22 to estimate F_{200} , F_{250} and F_{300} to within 5.3° , 6.4° and 7.5°C . respectively on 80 per cent of occasions.

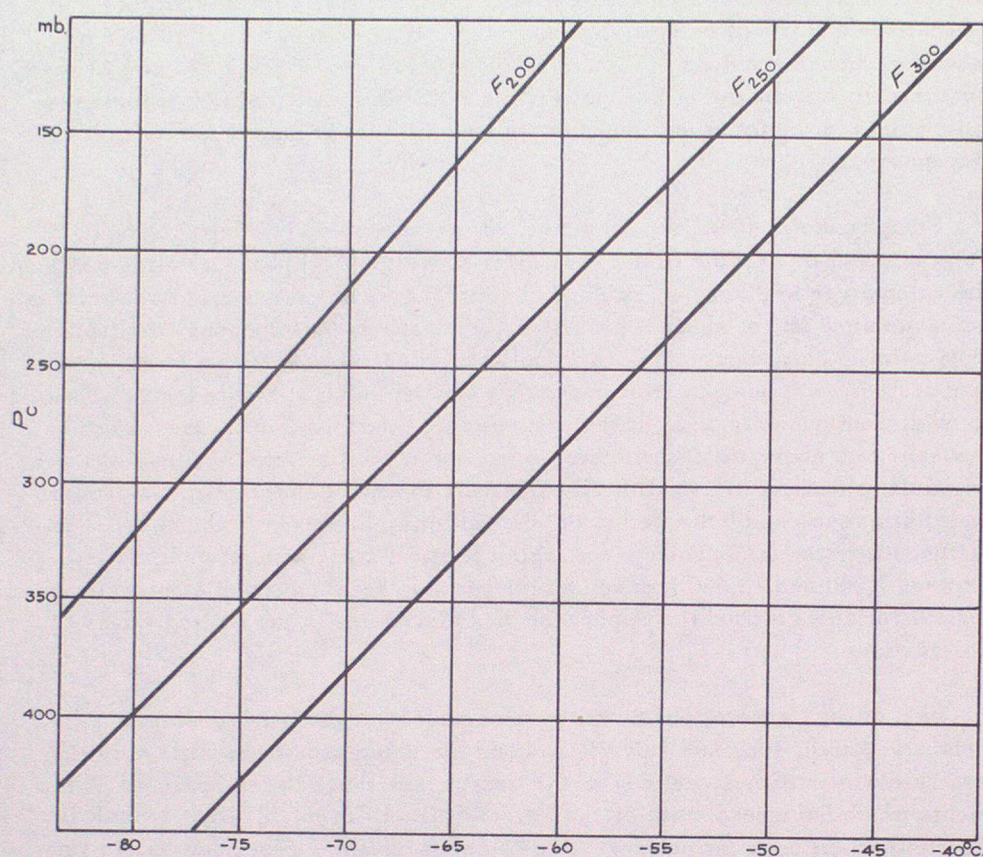


FIGURE 22 Relation between tropopause pressure and frost-points at the 300, 250 and 200-millibar levels

The correlations between F_{250} : T_{500} and F_{300} : T_{500} are particularly high. The authors believe that it may be significant that the correlation between frost-point and T_{500} is least at 200 millibars, a level which is predominantly stratospheric. This suggests that the frost-point in the stratosphere is less subject to control by tropospheric processes, or is less of an air-mass characteristic, than the frost-point at lower (predominantly tropospheric) levels. Figure 23 shows the relation between T_{500} and F_{250} (which pair of variables had the highest correlation). Given the temperature at 500 millibars, the line will indicate the frost-point at 250 millibars to within 5.3°C . on 80 per cent of occasions.

A further method for obtaining a statistical estimate of frost-points to somewhat higher levels of the order of 50,000 feet over southern England was put forward by Helliwell and Mackenzie.²⁴ Figure 24 is a graph showing the plot of air temperatures against frost-points actually observed from 81 high-level ascents in Meteorological Research Flight aircraft. Observations in the stratosphere are represented by circles and those at the tropopause by crosses. Three broad areas are distinguishable on the diagram. One in the upper left is predominantly tropospheric, that in the lower right is stratospheric and there is an intermediate band running from lower left to upper right indicating observations at or around the tropopause.

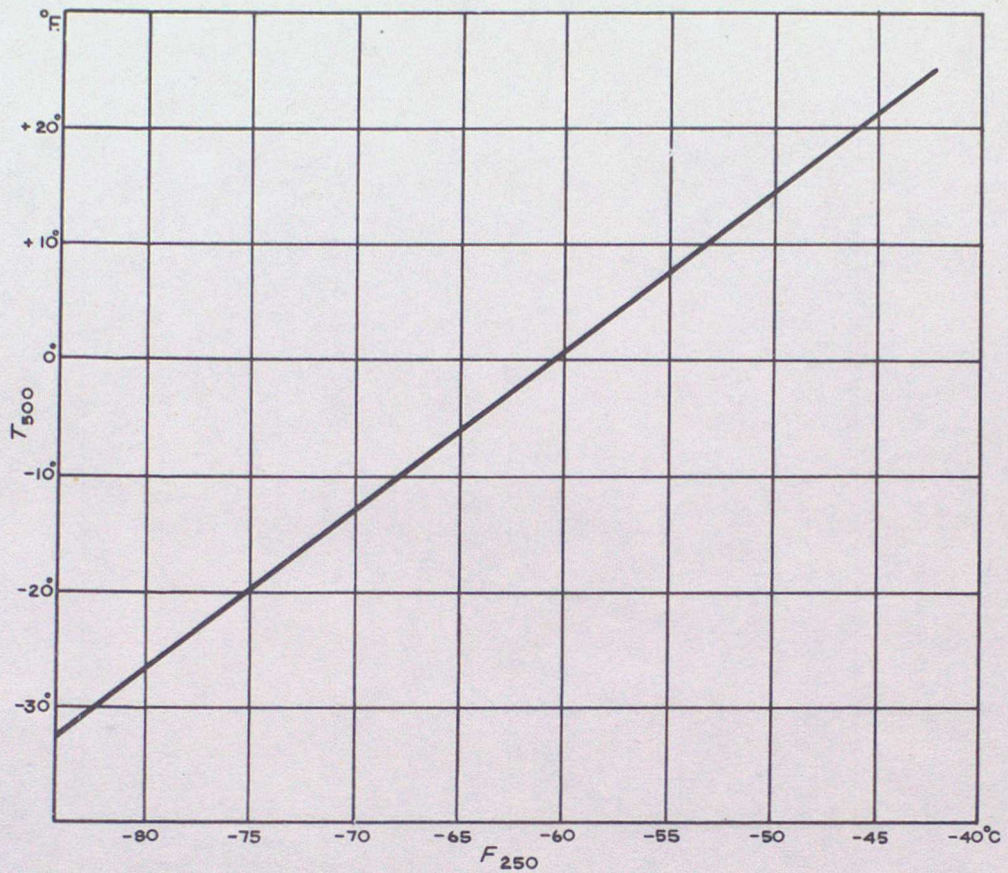


FIGURE 23 Relation between temperature at the 500-millibar level and frost-point at the 250-millibar level

Standard deviations of the depression of the frost-point at selected levels are shown in Table 6 and this table, combined with Figure 24 will give a first estimate of the frost-point depression over southern England at any level for a given ambient temperature.

TABLE 6 Mean temperature and mean depression of frost-point

Height	Mean temperature	Mean frost-point depression	Standard deviation of frost-point depression
ft.	°C.	°C.	
25,000	-37.3	10.4	8.9
30,000	-47.4	9.3	7.2
34,000	-54.9	9.5	6.8
36,000	-56.1	11.7	7.5
40,000	-55.2	20.3	9.4
44,000	-53.4	27.1	6.0
48,000	-53.1	29.2	6.3

RESTRICTED

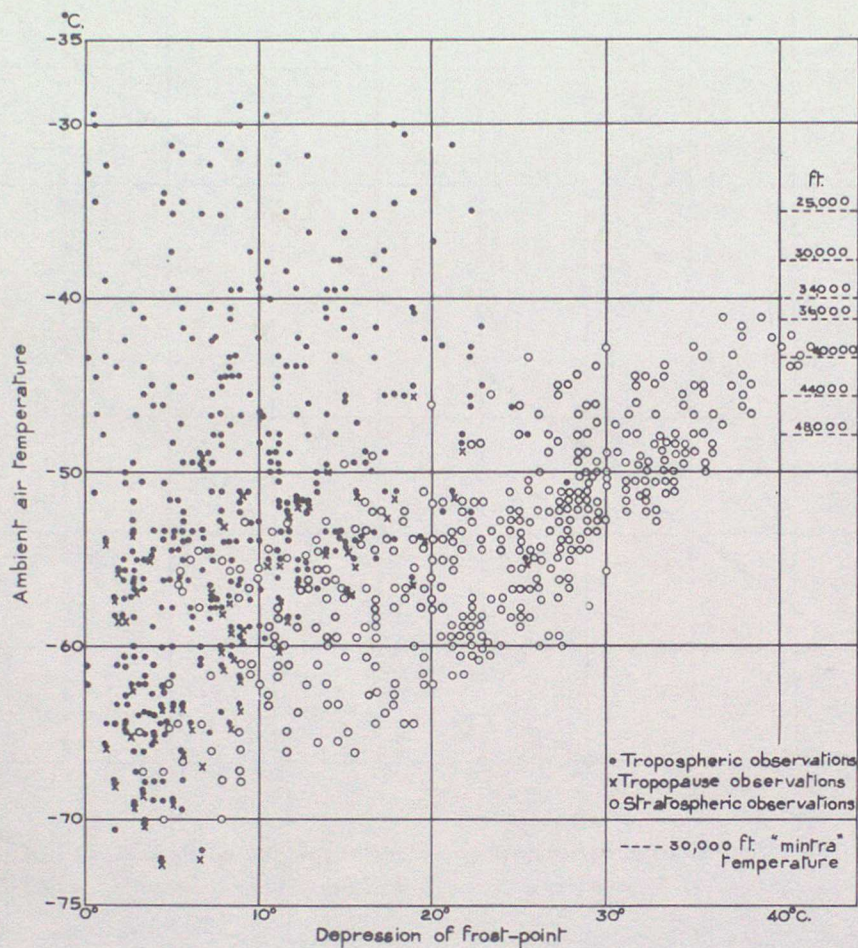


FIGURE 24 *Relation between ambient air temperature and frost-point depression observed on high-level ascents above 25,000 feet*

15.6. FORECASTING HUMIDITIES

It is convenient to consider the forecasting of humidities in three strata of the atmosphere: at or near screen level; in the lower and mid troposphere; in the upper troposphere and lower stratosphere.

15.6.1. At or near screen level

Over land areas in north-west Europe the number of dew-point reports is usually sufficient to determine quite accurately the current distribution of humidity on the synoptic scale, but minor variations due to topography, the nature of vegetation etc. which may be of importance to very local forecasting may not be detected by the network. Over adjacent seas the observing network is much more open and at times the observations may cause an indeterminacy in the analysis. Nevertheless on most occasions the current distribution of dew-point can be determined with fair accuracy from the surface analysis and the broad changes in dew-point can be estimated from a consideration of both the surface analyses and forecast surface charts together with considerations of air-mass modification and diurnal changes. The finer detail in the changes is often difficult to assess,

Humidity

particularly as many airstreams reach the British Isles after a sea track across areas from which observations may have been sparse. The following information should assist in making reasonable estimates.

15.6.1.1. *Advection of warm air across a cool sea.* In an investigation of haars or North Sea fogs on the coast of Great Britain Frost²⁵ has developed the theoretical aspects of the decrease of temperature and increase of humidity of a warm air mass as it moves from land over a cool sea. If u_1 is the initial specific humidity of the warm air as it leaves a land surface and u_2 is the specific humidity corresponding to air saturated at the temperature of the sea (assumed constant), then the specific humidity u_3 of the air after traversing various distances across the sea surface is given by

$$u_3 - u_2 = (u_1 - u_2) f(xz),$$

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

x (km.)	100	200	300	400	500	600	700	800	900	1000
$f(xz)$ for $z = 4$ feet	.175	.152	.141	.133	.127	.123	.119	.116	.113	.110

For the purpose of this calculation in practice humidity mixing ratios may be substituted for specific humidities. The table then means that the difference in humidity mixing ratio is reduced to a value less than one-fifth after 100 kilometres track across the sea and thereafter slowly but steadily to about one tenth. These figures were calculated for a wind speed of 4 metres per second (about 9 knots) but Frost considers they are valid for wind speeds of less than Beaufort force 4 to 5 (about 12 to 20 knots).

For those isobaric situations where the air leaves a land mass shortly before reaching our shores this formula enables a numerical estimate to be made of the humidity of the air on arrival at our coasts. Coupled with the rules for temperature and height of inversion given in Section 14.10.2 a reasonable estimate of the temperatures and humidities of the lower few hundred feet of the air can be made.

15.6.1.2. *Advection of cool air across a warmer sea.* Some rules for computing mixing ratios of cool air after advection across a warmer sea were given in Section 14.10.

15.6.1.3. *Advection across a land surface.* As the United Kingdom is surrounded by sea there are very few airstreams whose humidities are primarily controlled by long land tracks and these are mainly confined to tracks from between about east-north-east and south-south-west. From all other directions air has a predominantly maritime track. Land tracks across the United Kingdom are confined, even with the most favourable trajectories, to lengths of about 300 to 500 miles, light winds of an anticyclonic or cyclonic nature being excluded. With particularly favourable wind directions (for example, west-south-west across southern England or north-north-west across the length of Scotland and England) dew-points on the windward shores may exceed those on the distant lee shore by about 1° to 3°C . for west-south-west to west gradients. Differences for north-west to north gradients are somewhat greater. This drop in the dew-point of air near the ground is due to the upward transport of moisture exceeding the evaporation from the ground. It is likely to be significant only when there is deep convection. A large part of the fall in dew-point probably occurs within a few miles of the windward coast.

15.6.1.4. *The effect of sea-breezes.* When a sea-breeze reinforces the existing wind there is usually little change in humidity. However, when a sea-breeze replaces a pre-existing land-breeze there is usually a noticeable change of humidity at coastal stations and this change generally decreases with increasing distance inland. If very humid air is present over adjacent coastal waters and there are slack pressure gradients the onset of the sea-breeze may in fact mean the advection of sea fog. In that case the dew-point of the air at coastal stations rises to a value very close to the sea temperature. Where the general wind is from the land but is reversed temporarily by a sea-breeze, the sea-breeze has often had quite a short track across the sea and its dew-point is not usually as high as the temperature of the sea surface. In these cases it is necessary to make an estimate of the dew-point at the coasts from the general synoptic situation, the probable length of sea track and the "local circulation" of air giving rise to the sea-breeze. As the sea-breeze extends inland the increase in dew-point tends to diminish and local physical features, type of vegetation and soils will exert a control on the variations at any one locality. These variations should be determined from an examination of a sufficient number of actual changes observed at the station during sea-breeze conditions.

In his analysis of sea-breezes at Worthy Down, Peters²⁶ states that "about one-third of the cases were characterized by no definite changes in relative humidity, but the remainder evidenced some striking increases with the arrival of the sea-breeze. Rapid rises of between five and ten per cent were recorded, and increases of twenty-five or thirty per cent within about half an hour following the onset of the sea-breeze occurred in several cases." From 38 occasions of sea-breezes at Worthy Down on which computations of vapour pressure could reasonably be made the mean values before and after the arrival of the sea-breezes were 11.2 and 12.9 millibars (dew-points 47.5° and 51.3°F.) respectively. The largest rise on an individual occasion was from 9.8 millibars (dew-point 44°F.) at 1600 G.M.T. to 14.6 millibars (dew-point 54.7°F.) at 1800 G.M.T., the onset of the breeze having been at 1730 G.M.T. In some localities the increase of humidity following the onset of a sea-breeze may materially increase a fog-point as calculated from a midday tephigram (see Chapter 17) and so increases the chance of the development of radiation fog during the ensuing night.

15.6.1.5. *The Föhn effect.* It was noted in Chapter 14 that the physical features of the British Isles are not sufficiently extensive or massive to produce particularly well marked Föhn effects. The Scottish Highlands, the Pennines, Welsh Hills and one or two rather more isolated areas of high ground do produce noticeable Föhn effects at times. In Föhn conditions air descends down the lee slopes, adiabatic warming occurs and the humidity of the air flowing down the lee slope is less than that of the undisturbed air at the same level to windward. An estimate of the decreased dew-point can be made by assuming that, as the air is lifted over the obstructing high ground, all moisture in excess of that required to produce saturation is removed as precipitation. This calculation is readily performed on a tephigram. In Figure 25 let T_1 represent the surface temperature and D_1 the dew-point in the free air to windward of the hills and assume that the air is lifted mechanically through 100 millibars. Then air would ascend via the path T_1UV corresponding to dry- and wet-adiabatic lapse rates. When the air descends on the lee side the dry-bulb temperature would follow the path VT_2 and the dew-point VD_2 to the appropriate surface pressure, that is, T_2 and D_2 would represent the new temperature and dew-point of the Föhn air flow when it had descended to its original pressure level.

Humidity

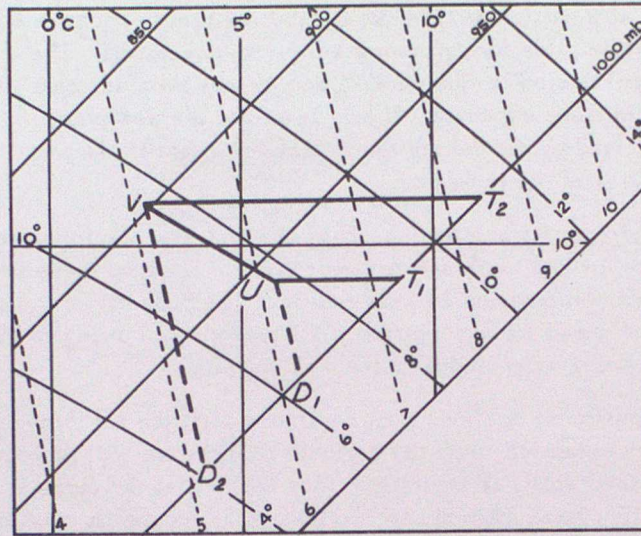


FIGURE 25 Calculation of decreased dew-point in Föhn conditions

McCaffery²⁷ has quoted an example of Föhn effect as air moved from Leuchars to Kinloss across the Cairngorms, the peaks of which rise above 4,000 feet. At Leuchars the air had a temperature of 33°F. and dew-point of 28°F. On the assumption that the air was lifted mechanically an average of 3,500 feet, McCaffery calculated that temperature and dew-point of the air at Kinloss should be 37° and 20°F. respectively. Observed values were 38° and 18°F. respectively.

15.6.1.6. *Diurnal and local variations.* Estimates of the diurnal variation in dew-points in air masses can be made from a consideration of the physical processes which are expected to be operative during the period. The descriptions of some observed variations given in Section 15.5 should assist forecasters to make reasonable estimates.

In showery conditions forecasters should note that an appreciable fall of rain in the late afternoon or evening may leave the ground wet and, on occasions of fairly light surface wind, materially increase the local dew-point (and fog-point) during the ensuing night.

15.6.2. *In the lower and mid troposphere*

Within the United Kingdom the analysis of the distribution of humidity in the free atmosphere rests almost entirely upon humidities observed by radio-sonde ascents. From the instrumental limitations described in Section 15.5.1 it is apparent that the radio-sonde does not show the true variations of humidity along its path. This is primarily due to the serious lag in the hygrometer which increases as the temperature falls. Nevertheless the reported values of relative humidity correspond within 10 to 20 per cent with the average relative humidity through a deep layer of the atmosphere.

From upper air ascents plotted on tephigrams together with an assessment of the horizontal and vertical distribution of humidity can be made over a 24- to 36-hour period. This assessment must then be modified according to the various physical processes which affect humidity and which are expected to take place. The more important of these processes are vertical motion and precipitation.

15.6.2.1. *The effect of vertical motion.* The effect of large-scale vertical motion on humidity at a particular level is difficult to assess because of the absence of quantitative rules for determining the vertical motion. The qualitative discussion of vertical motion in Chapters 14 and 16 may provide some guide to the vertical displacement to be expected. If this is known the dew-point (or frost-point) can be determined by displacing the initial dew-point (frost-point) along the water-content lines of the tephigram.

15.6.2.2. *The effect of precipitation.* It is obvious that precipitation removes water or ice from the air and carries it to lower levels. If air at those lower levels is unsaturated, evaporation will take place. The humidity of the lower layer will be increased and its temperature will be decreased owing to the latent heat of evaporation being extracted from the ambient air.

If liquid precipitation at or above 0°C. continues until the air through which it is falling becomes saturated, then the dry-bulb temperature will be reduced and the dew-point increased until, at saturation, they both equal the original wet-bulb temperature of the air. Some approximate working rules for the time taken for various rates of rainfall to reduce the dry-bulb temperature in the lower layers of the troposphere virtually to that of the wet-bulb (that is, also to increase the dew-point almost to the wet-bulb temperature) were given in Section 14.9.3.

Dolezel²⁸ has made a quantitative study of the changes, caused by evaporation from falling rain, in the humidity and temperature of the lowest 1,200 metres of the atmosphere. He found that the increase in humidity of this layer varied with the thermal stability of the air and that, for a given time after the commencement of rain, the increase in humidity was greater the greater the stability of the air. For a rate of continuous rainfall of 0.14 inches per hour and an assumed initial relative humidity of the layer of 50 per cent, he calculated that the relative humidity would increase according to the following table:

	Lapse rate of 0.5°C. per 100m.	Relative humidities for	
		Isothermal	Inversion of -1°C. per 100m.
		per cent	
After 1 hr.	78	79	83
After 2 hr.	90	92	97
After 4 hr.	96	99	106

In inversion conditions the falling rain comes from a warmer environment and hence the rate of evaporation is somewhat greater. The table shows that relative humidities increase more rapidly with increasing thermal stability of the air. Thus when estimating the increase of relative humidities and forecasting the possible formation of low stratus due to evaporation from falling rain, lapse rates should be considered.

15.6.2.3. *Humidity in air masses.* If recent observations are not available from an air mass which is expected to affect a forecast area near the British Isles a broad estimate of probable humidities can be obtained from Figures 15, 16 and 17 and Tables 14 and 15 of Chapter 14.

15.6.3. *In the upper troposphere and lower stratosphere*

At levels where the temperatures are below -40°F. (-40°C.) routine observations of humidity are not currently available to forecasters. Forecasts of humidity may be made on a statistical basis using any or all of the methods

Humidity

described in Section 15.5.5.3. Where other factors are believed to influence humidities, for example, high or low tropopauses, presence of high cloud, or the presence or absence of a temperature inversion at the tropopause, the statistical forecasts may be biased accordingly.

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