CHAPTER 18

HUMIDITY

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CHAPTER 18

HUMIDITY

18.1 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 17 - Temperature, 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, although computer analyses and prognoses are now capable of providing useful indications of the broad-scale humidity field in the free atmosphere. 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 19 - Clouds and precipitation, Chapter 20 - 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.

18.2 MEASURES OF HUMIDITY AND THEIR CONSERVATISM

18.2.1 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.

The formal definitions of the meteorological measures of humidity are contained in WMO 'Technical Regulations',[1] and 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.

18.2.1.1 Vapour pressure. The vapour pressure, e, is the partial pressure exerted by the water vapour present in the sample of air under consideration. The saturation vapour pressure, e_s, at temperature T is the partial pressure exerted by water vapour in equilibrium with a plane water (or ice) surface at the temperature T.

18.2.1.2 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, U, is defined as $U = 100e/e_s$; it is always expressed as a percentage.

18.2.1.3 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 r and expressed in grams per kilogram of dry air. If p is the pressure of the moist air and e is the vapour pressure then $r = 622e/(p-e)$ (in grams per kilogram). If r_s is the humidity mixing ratio at saturation, then $100r/r_s$ represents the relative humidity to a high degree of accuracy.

18.2.1.4 Dew-point temperature. The dew-point temperature, T_d, 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.

18.2.1.5 Frost-point temperature. The frost-point temperature, T_f, 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.

18.2.1.6 Dry- and wet-bulb temperatures. The dry-bulb temperature, T, of the air is the temperature recorded by a well-ventilated dry-bulb thermometer. The wet-bulb temperature, T_w, is the temperature to which air is cooled when pure water is allowed to evaporate into the air, adiabatically and at constant pressure, until saturation is reached. Note that the process is adiabatic, so that the heat required to evaporate the water must come from the air, and the water must be at the temperature T_w throughout. If T and T_w are the dry- and wet-bulb temperatures, r the humidity mixing ratio and r_s 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,

$$T_w = T - L(r_s - r)/ C_p$$

A derivation of this equation is contained in textbooks (see, for example, Petterssen [2]).

18.2.1.7 Dry- and wet-bulb potential temperatures. The dry-bulb potential temperature, θ, of a sample of air which does not at any stage contain condensed moisture is the temperature which the air would have if brought to a standard pressure - normally 1000 millibars - by an adiabatic process. On a tephigram the process is represented by a line of constant potential temperature, a straight line perpendicular to the isotherms.

In a similar way, the wet-bulb potential temperature, θ_w, is the temperature which a hypothetical sample of air would have if, initially saturated and at a temperature equal to the wet-bulb temperature of the original sample, it were brought to a standard pressure adiabatically and with saturation maintained throughout the process. On a tephigram the process is represented by the saturated-adiabatic lines.

18.2.1.8 Specific humidity. The specific humidity, q, (also termed the 'mass concentration' or 'moisture content') of moist air is the ratio of the mass m_v of water vapour to the mass $(m_v + m_a)$ of moist air in which m_v is contained, m_a being the mass of dry air, i.e. $q = m_v/(m_v + m_a)$. Since m_v is much smaller than m_a, specific humidity for a given sample is almost identical with humidity mixing ratio.

18.2.1.9 Vapour concentration. The vapour concentration, d_v, is defined as the ratio of the mass of water vapour, m_v, to the volume, V, occupied by the mixture, i.e. $d_v = m_v/V$. The alternative terms 'absolute humidity' and 'vapour density' applied to this quantity are not now favoured.

18.2.2 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.

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

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

18.2.2.3 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.

18.2.2.4 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.

18.3 ERRORS OF HUMIDITY MEASUREMENTS

18.3.1 Surface observations

Routine measurements of the humidity near the surface are normally made by means of dry-bulb and wet-bulb thermometers exposed in a standard screen, while a continuous record of relative humidity changes is usually made by a hair hygrograph. In the determination of the wet-bulb temperature, the flow of air past the thermometer must be such that the air near the bulb becomes saturated (see 18.2.1.6) but does not remain there long enough to inhibit further evaporation. In practice, lack of adequate ventilation (1 to 1.5 m/s) in the screen is more often a cause of inaccuracies than is too rapid a flow of air, and in stagnant conditions the wet-bulb reading may be a degree or more too high. Zobel[3] has quoted an instance where the derived relative humidity was 9 per cent too great and the dew-point 3.8 degrees too high.

Less frequently, errors in the dry-bulb temperature, and hence in the wet-bulb depression, may arise when moisture has been deposited on the dry-bulb, as sometimes happens during fog, and the subsequent evaporation of the moisture causes the dry-bulb thermometer to read too low. George[4] has carried out a number of tests to determine the magnitude of this effect; the maximum error he found was just under 1 degree, leading to an over-estimate of relative humidity of about 10 per cent. Occasionally, melting snowflakes may be blown on to the dry-bulb thermometer, leading to a similar error during evaporation.

The hair hygrometer is basically less accurate than the wet-bulb and dry-bulb psychrometer; the 'Handbook of meteorological instruments'[5] quotes an accuracy of 5 per cent. However, it is less dependent upon adequate ventilation than is the wet-bulb and dry-bulb psychrometer and, provided that adequate checks are made, can be used to detect gross errors of the type described by Zobel.[3]

18.3.2 Upper-air observations

The gold-beater's skin which forms the humidity element of the British radiosonde suffers from both lag and hysteresis effects which lead to errors in the reported humidities, particularly at the higher levels. The speed of response of the skin to humidity changes depends upon the ambient pressure and temperature, and upon the relative humidity indicated by the element; it represents the rate at which the state of the gold-beater's skin reacts to atmospheric humidity changes. Hysteresis, on the other hand, is a semi-permanent change which occurs when the skin has reached a fairly dry state through exposure to low humidity; subsequent increase in ambient humidity is not indicated fully by the element, which remains slightly shorter than it should be, until fairly high humidities are reached when the skin appears to recover its original properties. Both aspects have been studied fully by Glückauf[6] in the laboratory, and McIlveen and Ludlam[7] have made detailed comments upon the meteorological implications of Glückauf's results.

For a given temperature and pressure, the variation of the response time, τ, of gold-beater's skin with relative humidity is shown in Figure 1; τ is defined as the time taken for the difference between the indicated relative humidity and a constant ambient relative humidity to decrease by a factor $1/e$, where e is the base of Napierian logarithms. The obvious important feature is the large increase in τ/τ_{min} at very low and very high humidities.

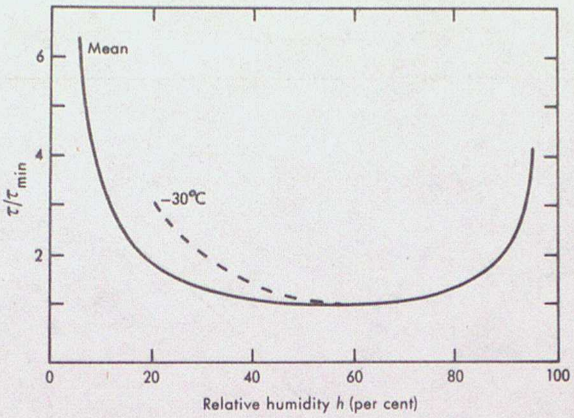

FIGURE 1. Variation with indicated relative humidity h of the Glückauf response time τ divided by the minimum value at the same temperature

——— represents the mean of Glückauf's observations over a range of temperatures from −30°C to 18°C; – – – represents his observations at −30°C.

The response time, τ, for non-extreme values of relative humidity increases with decreasing temperature rather more rapidly than logarithmically, and in much the same way as the saturated water vapour concentration, d_{vs}, increases with temperature, suggesting that the response time is primarily dependent upon d_{vs}. For the ICAO standard atmosphere, the variations with height of temperature, T, and the response time are shown in Figure 2. It will be noted that τ increases almost logarithmically with height from a few seconds at temperatures above freezing-point to about 10 minutes at -55°C. For very high or very low humidities, τ from Figure 2 must be multiplied by the ratio τ/τ_{min} obtained from Figure 1.

The effects of hysteresis are illustrated in Figure 3. It can be seen, for example, that when the skin has been dried to such an extent that the indicated relative humidity is just under 2 per cent, subsequent indicated values in the range 10-40 per cent are 10-11 per cent too low even though the skin has reached equilibrium with the surrounding atmosphere; above 40 per cent indicated relative humidity the correction decreases until at 70 per cent it becomes zero, that is to say the skin has recovered its original properties.

The above facts lead to the conclusions that the most serious errors will occur when the air being sampled, or that which has recently been sampled, is either very dry or very moist. It also seems likely that the errors will be greater at higher than at lower levels, but this may be partly

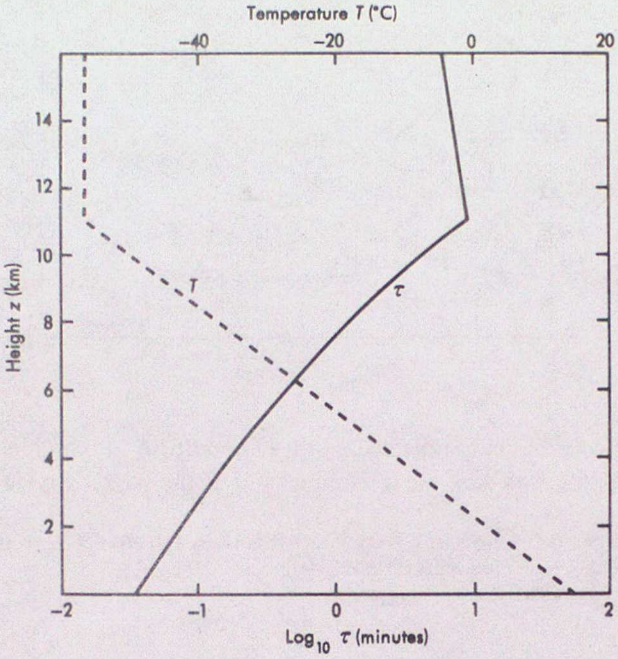

FIGURE 2. The Glückauf response time τ and air temperature T as a function of height z in the ICAO standard atmosphere

——— τ - - - - - T

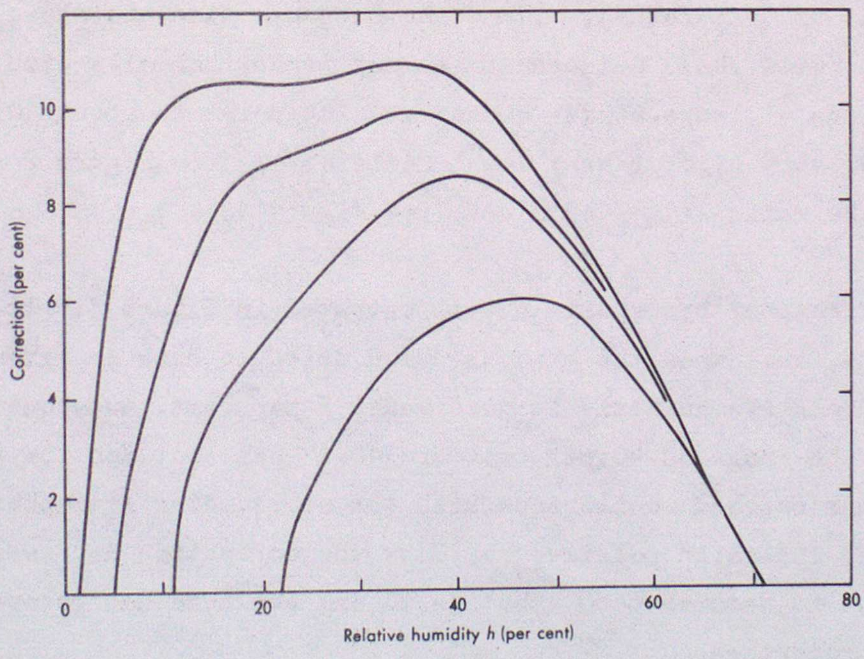

FIGURE 3. The effects of hysteresis on the relative humidity indicated by gold-beater's skin in the laboratory

Each curve shows the correction to be applied to any indicated relative humidity after a certain minimum value has been indicated. Corrections needed following the indication of minimum relative humidities other than 1.7, 5, 11 and 22 per cent can be found by interpolation between the curves.

compensated for by the greater variability of humidity at low levels; (this leads us to the question of the representativeness of humidity soundings, which will be discussed in the next section).

An example of the behaviour of the gold-beater's skin on passing through a cloud layer into very dry air above is shown in Figure 4, where the sonde observations are compared with aircraft measurements.

The relative-humidity profile which would be indicated by a gold-beater's-skin humidity element, ascending at a known rate, can readily be derived for simple real profiles, and the forecaster may find it instructive to carry out a few calculations based on the data given in this section. One or two examples have been discussed by McIlveen and Ludlam.[7] Of more value in operational meteorology would be the reverse process - to derive the actual profile from that indicated by the element. This, however, is a complex and laborious process and cannot be carried out manually in the time available for computing and assembling a synoptic report. However, there is every hope that, when station computers are available, lag and hysteresis effects will be corrected, leading to a valuable increase in the accuracy of the sounding and extending its usefulness to greater heights. In such circumstances the corrected humidity data represent running-mean values over larger and larger layers as the response time, τ, increases with altitude.

18.4 DISTRIBUTION OF MOISTURE IN THE ATMOSPHERE

Nearly 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 atmosphere in contact with the surface; secondly by mechanical turbulence, mainly in the lowest 300 metres 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

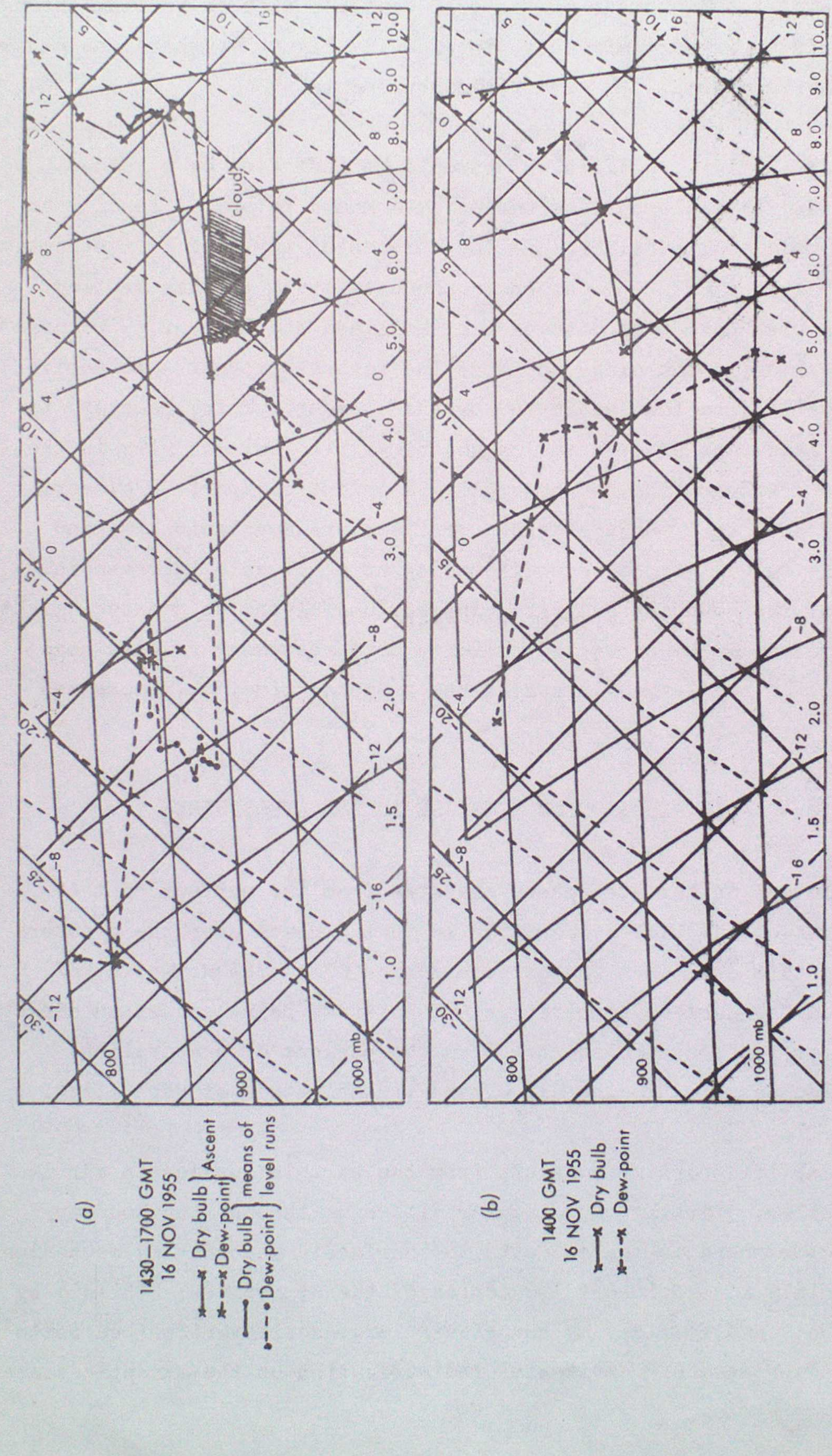

FIGURE 4. Temperatures and humidities as measured (a) from an aircraft and (b) by radiosonde when a layer of stratocumulus was present

(Diagrams adapted from MRP 1055, 1967)

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 paper by Sutcliffe,[8] 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.

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 6 - Depressions and related features. Superimposed on these variations is a variability of humidity on a small scale (of the order of a few kilometres). 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 hundred metres or so 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 [9-11] 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 degrees in 24 kilometres are common and changes of 15 degrees or more are sometimes observed.

Such variations in humidity are often brought about by convection or by mixing, on a range of scales, of air from different origins. Some examples will be discussed in detail in 18.4.2 and 18.4.3.

18.4.1 Humidity in the lower layers of the atmosphere

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. 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. 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 are 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 Stormonth [12] 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. Vapour concentration, d_v, 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. Vapour concentration may be converted to vapour pressure in millibars by use of the equation

$$e = d_v T/216\,700$$

where d_v is the vapour concentration in milligrams per cubic metre and T is the temperature in kelvins. Values of corresponding dew-points may then be obtained from hygrometric tables or from the humidity slide rule.

The mean diurnal variation of the vapour concentration at two heights (1.1 and 106.7 metres) is shown in Figure 5 for all days in selected months. In Figure 6 are shown similar curves for clear and overcast days in winter and in summer; based on a much smaller sample of data than the mean curves, they show some variations which may not be significant, but it is felt that these data are more likely to be of direct value to the forecaster than are the means for all days. The corresponding diurnal variations in dew-point temperature are given in Table 18.1.

TABLE 18.1 Diurnal range of dew-point temperature at 1.1 and 106.7 metres at Rye

	1.1 m	106.7 m
	deg	*deg*
January	1.5	1.5
February	1.5	1.5
March	3.5	1.5
April	3.0	1.5
May	3.5	1.5
June	3.5	1.0–1.5
July	3.0	1.0–1.5
August	4.5	1.5
September	2.0	1.0
October	3.5	1.0
November	2.0	1.0
December	1.5	0.5
Summer		
Clear days	4.5	3.5
Overcast days	2.0	2.0
Winter		
Clear days	3.5	1.0–1.5
Overcast days	1.0	0.5–1.0
Snow cover	1.5	1.5

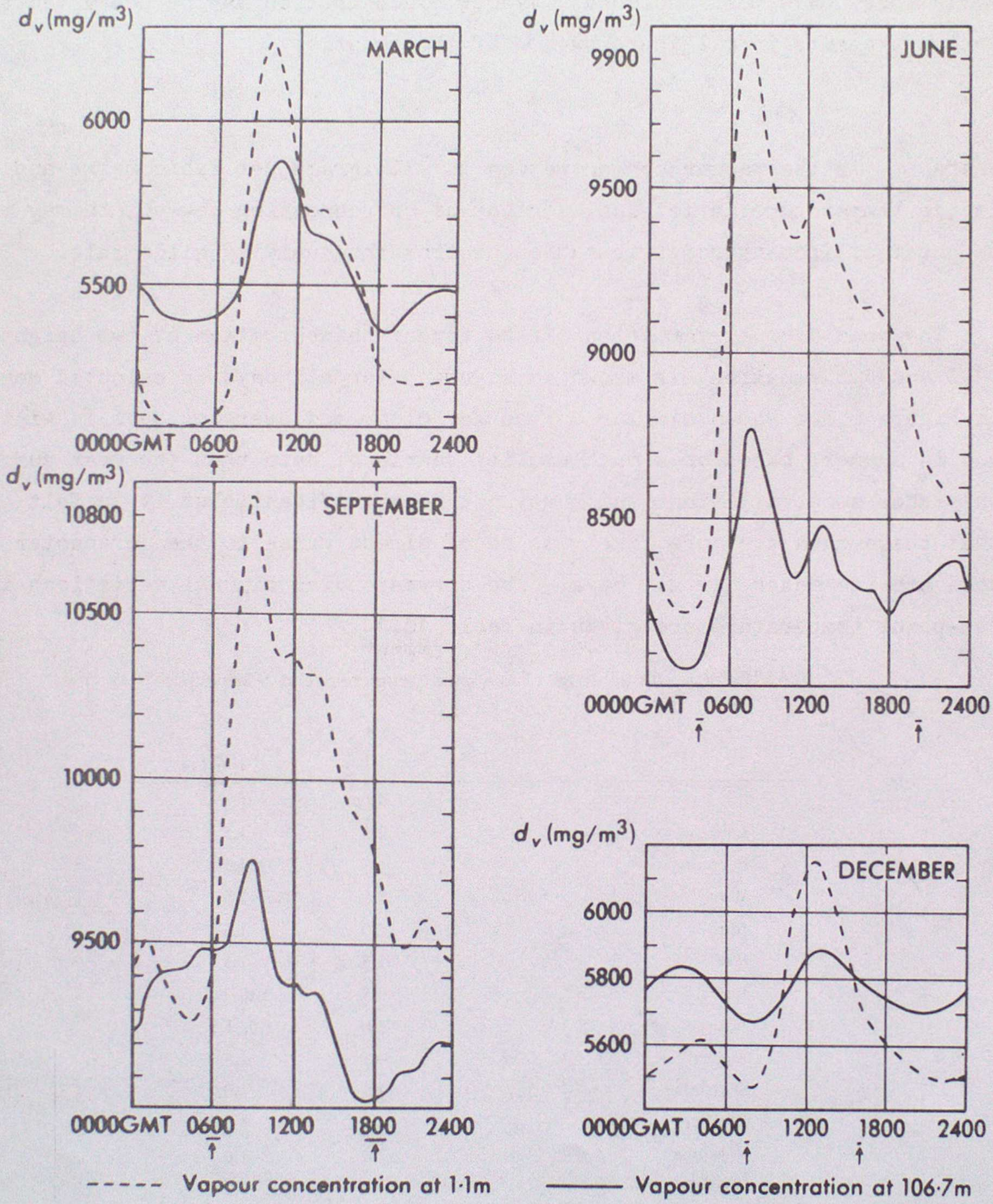

FIGURE 5. Diurnal variation of vapour concentration at 1.1 and 106.7 metres for March, June, September and December at Rye

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.

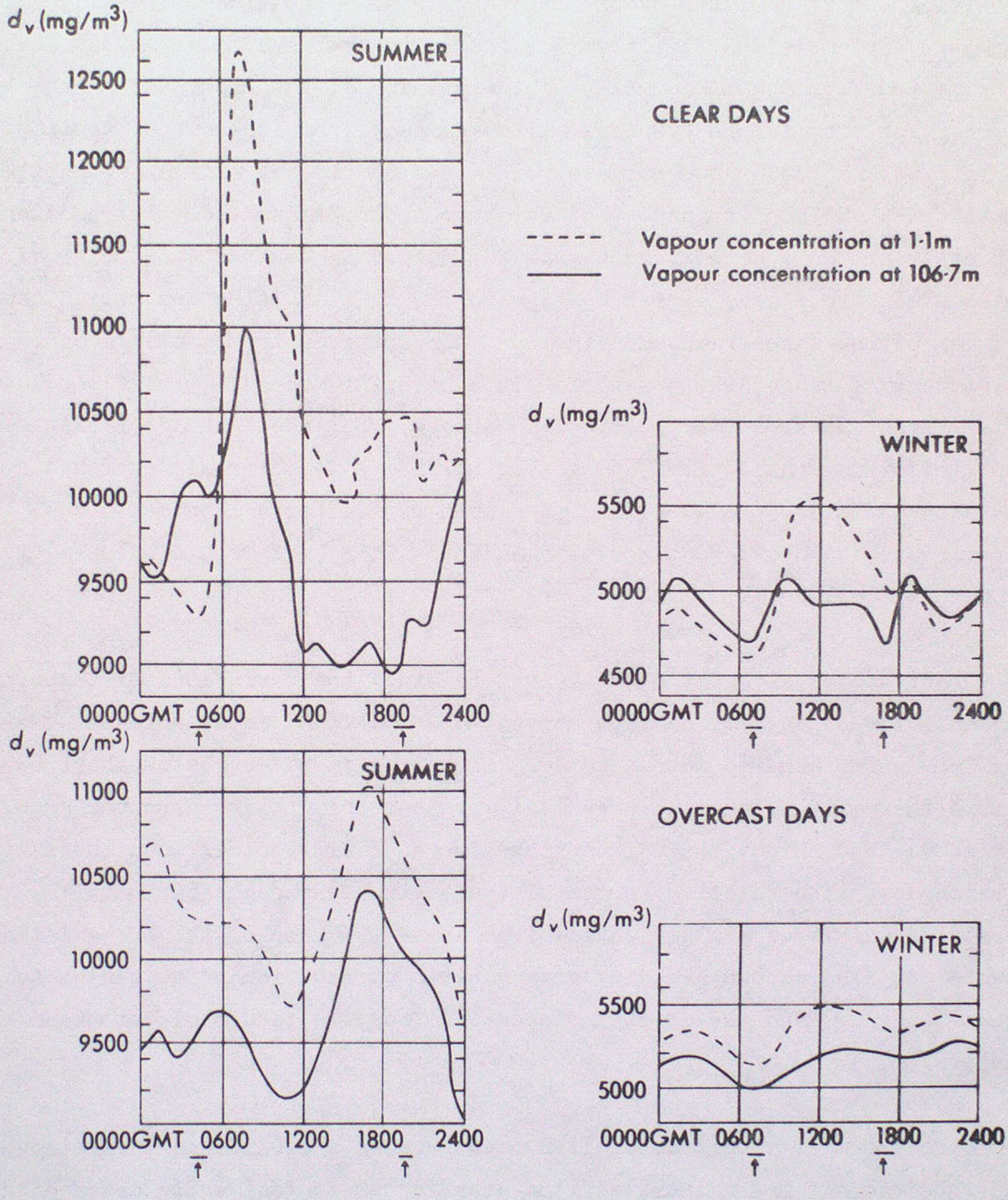

FIGURE 6. Diurnal variation of vapour concentration at 1.1 and 106.7 metres for clear and overcast days in winter and summer at Rye.

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

18.4.2 Humidity in air masses

Changes in humidity occur 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 the major pressure systems. The modifications are in general accord with what would 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 a cooler sea surface reinforces an already stable lapse rate, so that the vertical transport of moisture is suppressed to a large extent and there are only slight changes in humidity. In contrast, as polar air moves over warmer seas on its passage to the British Isles convection occurs and may become widespread and vigorous. In spite of the removal of some moisture by showers, the convection is a powerful mechanism for distributing water vapour upwards through the troposphere. Belasco[13] found, for example, that the mixing ratios from 950 to 500 millibars in polar air approaching the British Isles in summer from the south-west after a long sea track were greater by about 50 per cent than those in a northerly airstream direct from polar regions.

Grant[14] used aircraft observations to study the variations of temperature and humidity on days of shallow convection. He found moist patches, varying from a few hundred metres to 1-2 kilometres in size; below about 600 metres the moist patches were generally warmer than their surroundings, but above that height the moist patches were usually colder than their environment. The relative warmth of the environment suggests that it consisted mainly of air descending from above the base of the pre-existing inversion. The excess humidity mixing ratio varied from a few tenths to about 1.5 g/kg, and the moist air occupied from 0.4 to 0.8 of the space sampled.

Strong humidity contrasts often occur across a subsidence inversion. The subsided air has maintained its water-vapour content while descending from higher levels and is therefore relatively dry. An example of such a humidity gradient has been shown in Figure 4; a further example, from a study by Cornford[15] of the temperature and humidity distribution in a south-easterly airstream over the North Sea in winter, is given in Figure 7.

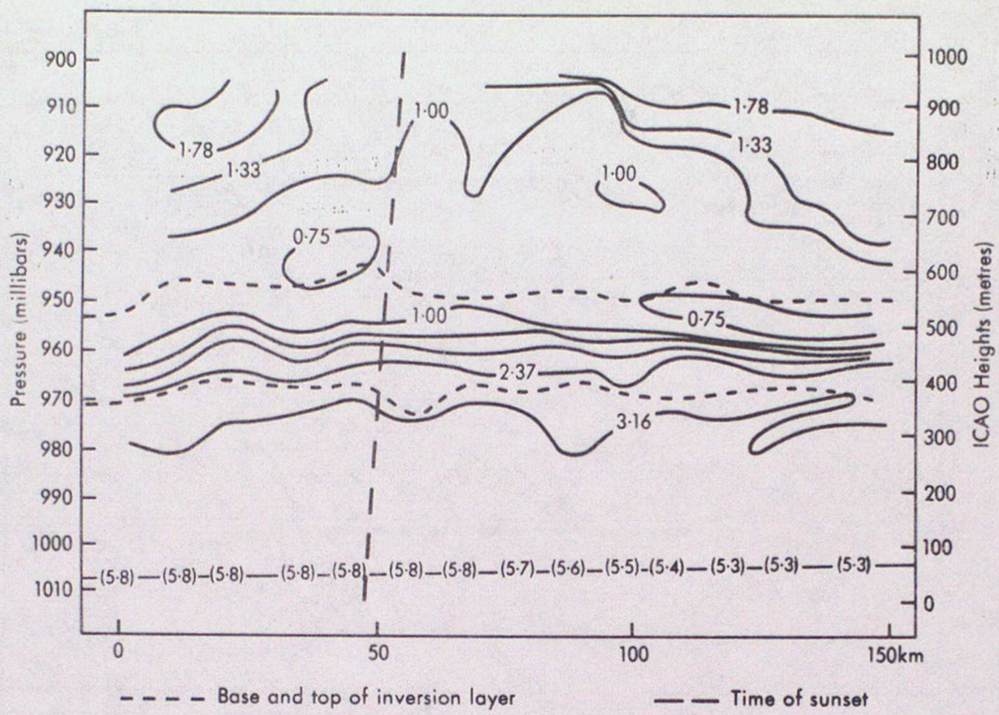

FIGURE 7. Variation of humidity mixing ratio, r, with pressure-height and horizontal distance travelled by the air

Isopleths are drawn at equal intervals of $\log_{10} r$, but are marked with the value of r in gram per kilogram.

18.4.3 Humidity in frontal regions

The idealized concept of abrupt changes in air-mass characteristics through frontal surfaces is useful for many aspects of analysis and forecasting, but detailed investigations of fronts have shown that the changes which actually occur are often much more complicated, especially for humidity. In particular, studies by Sawyer,[10] Miles[16] and Freeman[17] have demonstrated the existence of a tongue of relatively dry air extending downwards from the upper troposphere or lower stratosphere, often within the frontal zone but at times above or below. Examples from Freeman's paper are given in Figures 8(a), (b) and (c):

(a) a warm front, showing very dry air at about 600 millibars within the frontal zone;

(b) a cold front, with very dry air at or slightly below 800 millibars within the frontal zone; and

(c) a complex cold front with very dry air both within and ahead of the frontal zone in the middle troposphere.

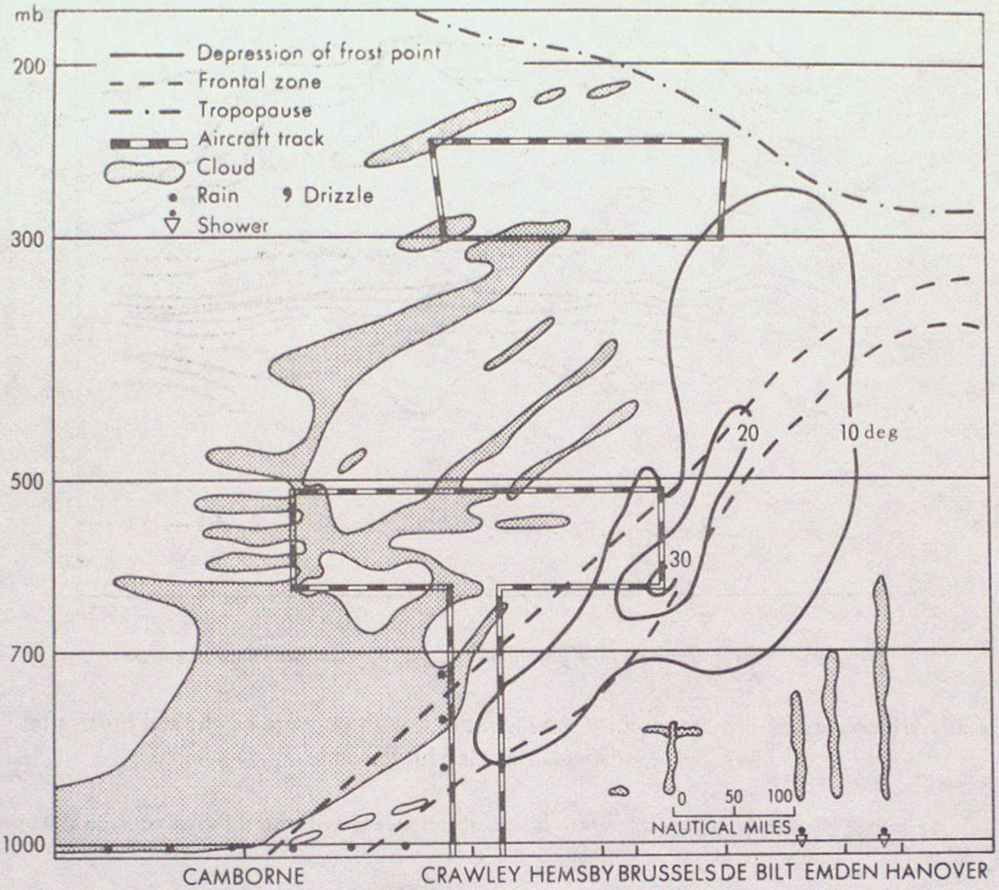

FIGURE 8(a). Warm front of 7 October 1955

Vertical cross-section showing humidity and cloud

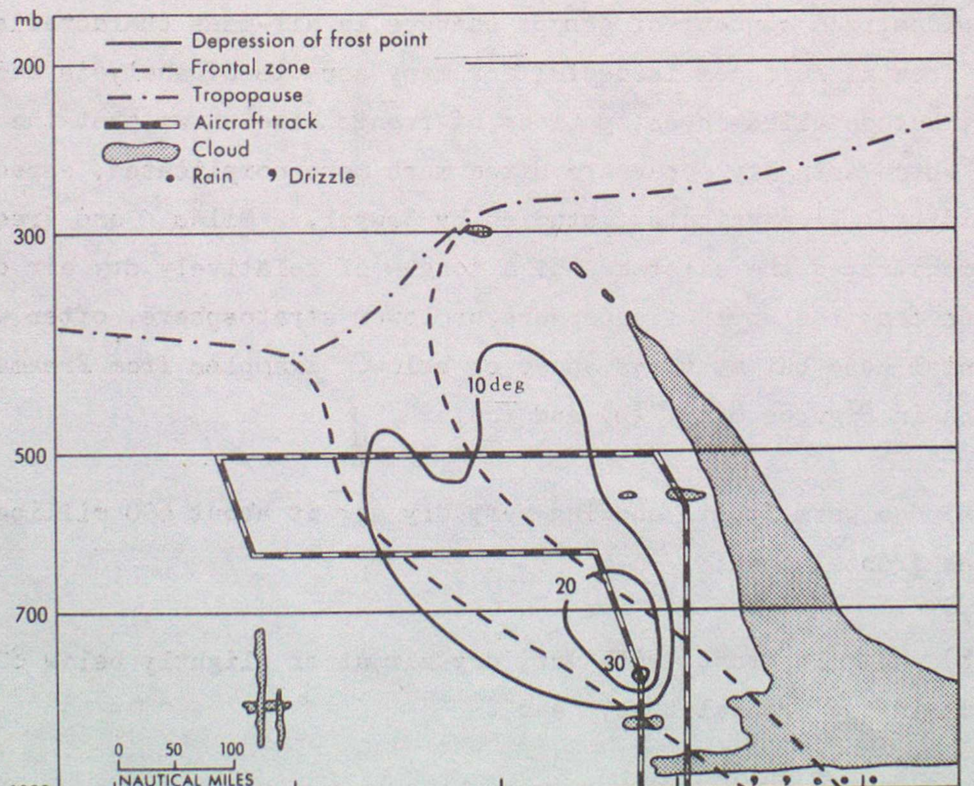

FIGURE 8(b). Cold front of 11 January 1955

Vertical cross-section showing humidity and cloud

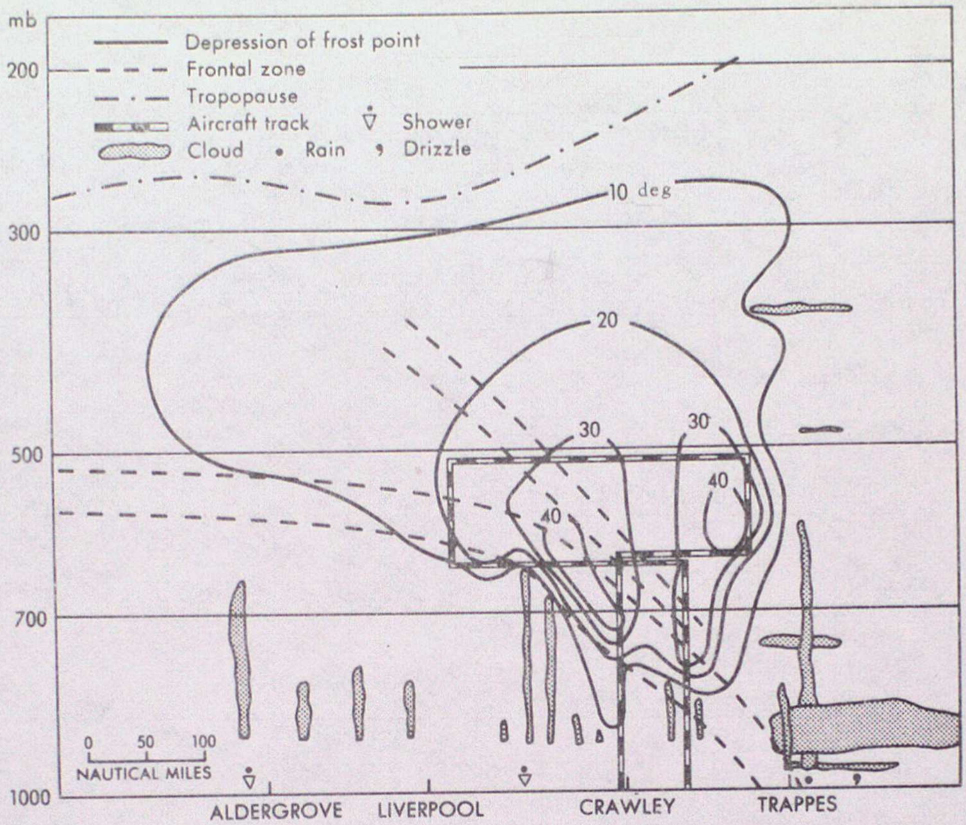

FIGURE 8(c). Cold front of 16 September 1954

Vertical cross-section showing humidity and cloud

The humidity elements in routine use in radiosondes suffer from lag and, at times, hysteresis effects (see 18.3, page 5) which make their indications unreliable, or at least difficult to interpret, in the upper troposphere. However, McIlveen and Ludlam[7] have shown how allowances can, in principle, be made for both lag and hysteresis. When this can be done in practice, more reliable indications of upper tropospheric cloud and frontal surfaces should become regularly available to the forecaster.

18.4.4 Humidity in the stratosphere

Routine measurements of the humidity in the stratosphere are not yet available, but an extensive **series** of observations was made, using frost-point hygrometers mounted in aircraft, during the 1950s. The results have been presented by Bannon, Frith and Shellard,[18] Murgatroyd, Goldsmith and Hollings[19] and Tucker,[20] and are summarized in Figure 9, in which the frost-point temperatures have been plotted against the pressure difference between the appropriate level and the tropopause.

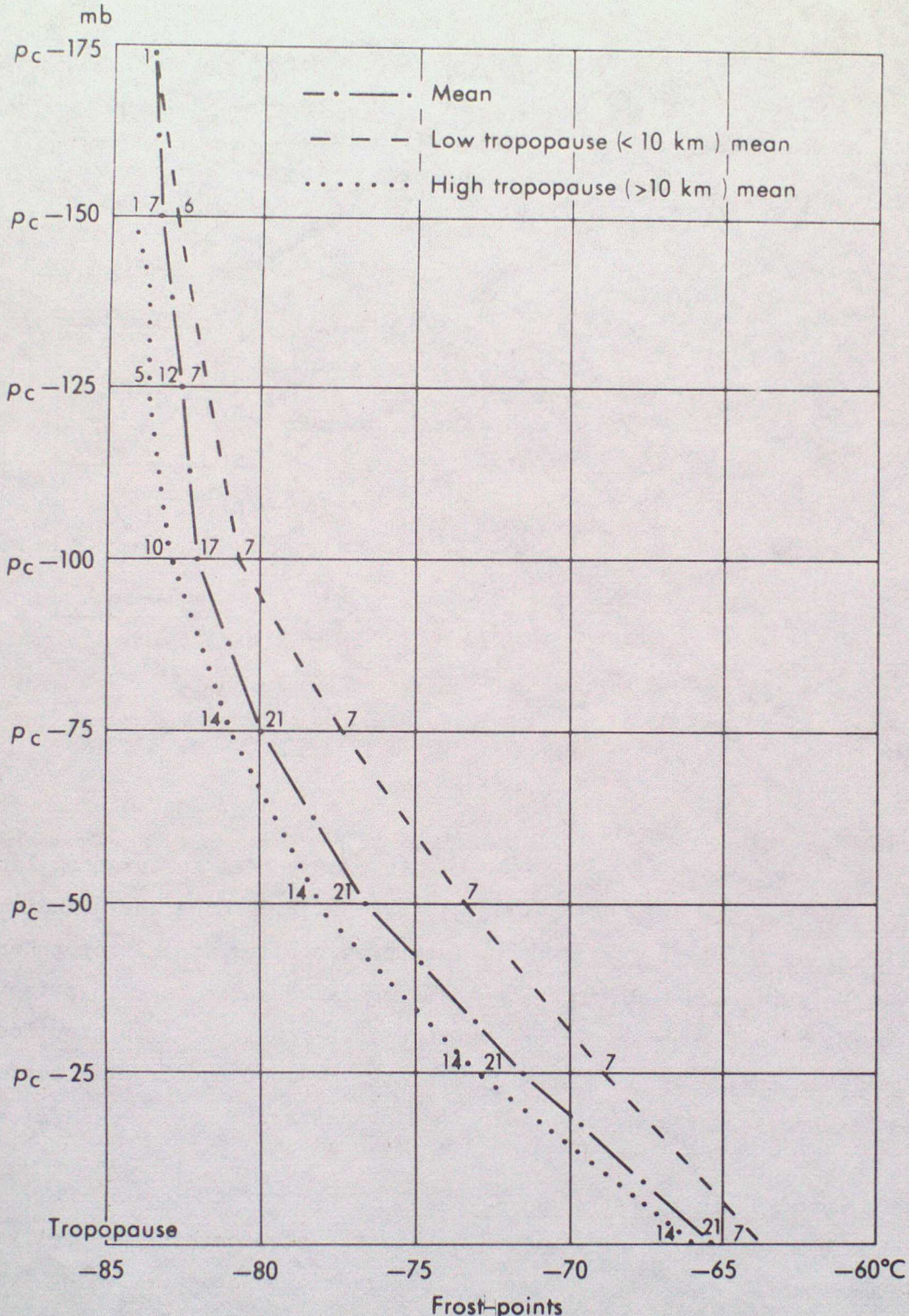

FIGURE 9. Mean frost-points at levels above the tropopause

The figures against the curves give the number of observations available

The main feature is that the lapse rate of frost-point usually does not show a discontinuity at the tropopause, in contrast to the behaviour of the temperature lapse rate. However, the lapse rate of frost-point temperature does decrease slowly with increasing height, until at levels 100-150 millibars above the tropopause (15-16 kilometres), the frost-point reaches a value of about -80°to -85°C. There is no significant seasonal variation in the frost-point at these heights, and there is little variation with latitude. However, frost-points in the lower stratosphere are generally higher when the tropopause is low than when it is high. Above about 16 kilometres, observations are few, but it appears (Williamson and Houghton[21]) that the mixing ratio remains substantially constant with height at about two or three milligrams of water vapour per kilogram of dry air.

18.5 FORECASTING HUMIDITY

18.5.1 <u>At or near screen level</u>

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, 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.

18.5.1.1 <u>Advection of warm air across a cool sea</u>. In an investigation of haars or North Sea fogs on the coast of Great Britain, Lamb[22] 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 q_0 is the initial specific humidity of the warm air as it leaves a land surface and q_s is the specific humidity corresponding to air saturated at the temperature of the sea (assumed constant), then the specific humidity q of the air after traversing various distances across the sea surface is given by

$$q = q_s + (q_0 - q_s)\,f(x,z)$$

where x is the distance traversed and z is the height above the surface. Taking z as 1.22 metres (that is, screen height) the following figures are the values of $f(x,z)$ for trajectories up to 1000 kilometres in steps of 100 kilometres.

x(km)	100	200	300	400	500	600	700	800	900	1000
$f(x,z)$ for z = 1·22 m	0.175	0.152	0.141	0.133	0.127	0.123	0.119	0.116	0.113	0.110

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 8 knots) but Lamb considers that they are valid for wind speeds of less than Beaufort force 4 to 5 (about 12 to 20 knots).

18.5.1.2 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 17.7.8 of Chapter 17 - Temperature.

18.5.1.3 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 480 to 900 kilometres, 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 1-3 degrees for west-south-west to west gradient winds. Differences for north-west to north gradient winds 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 kilometres of the windward coast.

18.5.1.4 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 this 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 temporarily reversed 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, Peters[23] 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 5 and 10 per cent were recorded, and increases of 25 or 30 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 8.6° and $10.7^{\circ}C$) respectively. The largest rise on an individual occasion was from 9.8 millibars (dew-point $6.7^{\circ}C$) at 16 GMT to 14.6 millibars (dew-point $12.6^{\circ}C$) at 18 GMT, the onset of the breeze having been at 1730 GMT. 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 20 - Visibility), and so increase the chance of the development of radiation fog during the ensuing night.

18.5.1.5 <u>The föhn effect</u>. It was noted in Chapter 17 - Temperature - 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 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 10, let T_1 represent the surface temperature and T_{d1} 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 VT_{d2} to the appropriate surface pressure, that is T_2 and T_{d2} would represent the new temperature and dew-point of the föhn air when it had descended to its original pressure level.

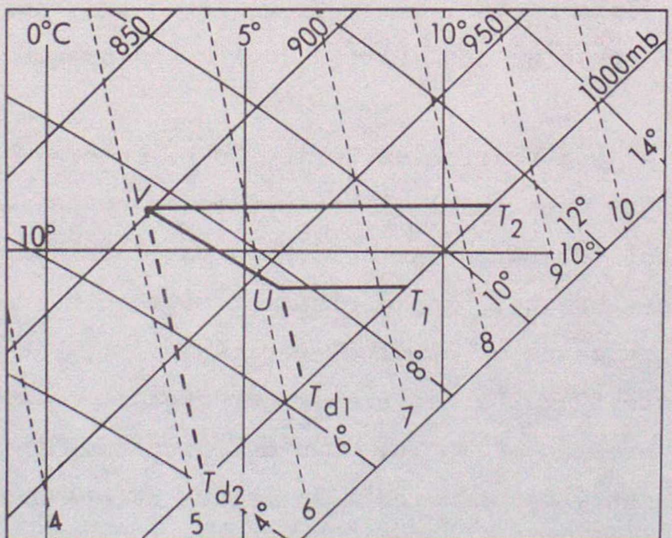

FIGURE 10. Calculation of decreased dew-point in föhn conditions

Girdwood,[24] Green et alii,[25] Green[26] and McCaffery[27] have discussed specific examples of the effect of the föhn wind on humidity.

18.5.1.6 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 18.4, page 9) 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 night.

18.5.2 In the free atmosphere

Within the United Kingdom the analysis of the distribution of humidity in the free atmosphere rests almost entirely upon humidities observed by radiosonde ascents. From the instrumental limitations described in section 18.3.2 (page 6) it is apparent that the radiosonde 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.

The analysis and prognosis of the large-scale humidity field, including changes brought about by vertical motion (broad-scale associated with frontal systems - subsidence and convection) can now be carried out by computer. The objective assessments of vertical motion are in general very much better than subjective ones, but there may be some areas, near jet streams and fronts for example, where subjective modifications to the objective products would improve the detail. As with temperature and wind forecasts, any such subjective modification must be attempted only by those experienced in the behaviour of the numerical models used and in the synoptic and dynamical aspects of the systems concerned.

The humidity at certain levels may also be changed if precipitation is occurring. Precipitation removes water or ice from the air and carries it to lower levels. If air at these 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 17.7.9 of Chapter 17 - Temperature.

Dolezel[28] has made a quantitative study of the changes, caused by evaporation from falling rain, in the humidity and temperature of the lowest 1200 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 3.6 millimetres 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:

	Relative humidities for		
	Lapse rate of 0.5 deg per 100 m	Isothermal	Inversion of −1 deg per 100 m
		per cent	
After 1 hour	78	79	83
After 2 hours	90	92	97
After 4 hours	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.

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