CHAPTER 10

UPPER-AIR ASCENTS

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Contents

LIST OF DIAGRAMS

10.1 INTRODUCTION

The foregoing chapters of this Handbook have emphasized the three-dimensional nature of the processes which determine the weather. Information on the vertical structure of the atmosphere is, therefore, vitally necessary if we are to understand these processes and to use them to predict the weather. For this reason observations of pressure, temperature, humidity and wind in the free atmosphere are made regularly at a number of land and ocean stations, although, largely because of the cost, the number of upper-air stations is small compared with the number of surface stations, particularly over the oceans. Fortunately, irregularities in the spatial distribution of temperature, humidity, and wind are not as great in the upper air as at the surface, although increasingly accurate and sensitive instruments are revealing small-scale variabilities, particularly in the vertical. Knowledge of these small-scale features may be important for some purposes, as in the study of radio and radar propagation, but not for others, such as aviation route-planning and navigation. This difference highlights the two main uses of upper-air data:

(a) To determine the vertical structure of the atmosphere at or near a given location: important in the assessment of local weather features, etc.

(b) To determine the atmospheric flow patterns at various heights or pressure levels: important in the assessment of the development of synoptic systems and in aviation forecasting.

In this chapter the analysis of individual upper-air ascents, or aerological soundings, will be described, together with some indication of the uses to which the data may be put. More detailed information on the use of upper-air data in local forecasting is contained in the chapters on individual elements (Chapters 16 to 23).

The objects of any analysis of meteorological data are:

(i) To scrutinize the data critically, so as to reject all obvious errors and estimate the significance and plausibility of any variations in the observations which may appear abnormal.

(ii) To gain as complete an understanding as possible of the physical state of the atmosphere at the time, and of the physical processes going on.

(iii) To estimate the nature of the changes from the present state which would occur if the atmosphere were subject to various dynamic or thermodynamic processes.

Enlarging on these three points, with particular reference to upper-air soundings:

The first, (i), demands that the analyst shall be familiar with the 'normal' patterns of upper-air temperatures and winds, and the variations that may occur in different types of situation. Some variations from the expected pattern may be real, while others may immediately strike the experienced analyst as being incorrect. For instance, an error of 10 degrees in the temperature at one level is easily recognized: such errors may arise in transmission or, infrequently, by mis-plotting.

The analyst must also have a knowledge of the instrumental errors, and of the degree to which the basic measurements are processed and smoothed at the observing station before transmission. This is particularly essential with upper-air observations. The accuracy of the British Mark 2B radiosonde is discussed in 10.3 (page 3), and compared with the errors of the Mark 3 sonde derived from a limited number of flights. More general remarks on the error of radiosonde and radar-wind measurements are made in the next chapter, which deals with the analysis of upper-air charts.

With this background knowledge, a scrutiny of one particular sounding and comparison with its predecessors from the same station and with neighbouring contemporary soundings will bring to light most of the errors of a temperature sounding. The detection of errors in an upper-wind sounding is generally much more difficult.

It may be noted here that the scale or degree of accuracy to which an analysis may be carried out is determined by the size of the observational mesh and by the time interval between ascents. In the United Kingdom the average distance between upper-air stations is about 250 kilometres, and over Europe the network is almost as dense, but over the oceans, and over many land areas outside the temperate zones of the northern hemisphere, the network is very sparse. The interval between observations is normally 12 hours, although a few stations report at six-hourly intervals and some only once a day. The space and time intervals between observations make it possible for some mesoscale or small synoptic-scale systems to pass unnoticed through the mesh or to affect the measurements at only one station to a greater or lesser extent.

The second, (ii), involves a detailed inspection of the sounding and the carrying out of various preliminary constructions. By this means considerable progress can be made in answering the question, 'What is the atmosphere like now?'. Much information can be deduced from the changes in lapse rate, hydrolapse, wet-bulb potential temperature, winds and thermal winds, and about air masses, the presence and activity of fronts, cloud types, and the advection of warm or cold air.

The third, (iii), involves estimating what the result would be of heating or cooling of the air represented on the sounding, either by dynamical ascent or subsidence, or by radiation, conduction and convection, or by advection. The possibilities of convective overturning, formation and dissolution of cloud need to be considered. Whether such heating or cooling will take place is something which can only be decided from wider considerations on a synoptic scale.

10.2 TYPES OF UPPER-AIR OBSERVATION

10.2.1 Pressure, temperature, and humidity

The most common means of 'sounding' the atmosphere is by radiosonde. The instrument carries sensors which respond to the ambient pressure, temperature, and humidity, each of which, through transducers, alters the frequency of an oscillatory circuit. The signals are broadcast and are usually received and decoded at the ground station. The height of the sonde at any time is calculated from the observed pressure, temperature and humidity values, using the hydrostatic relationship. The errors of radiosonde observations are discussed in 10.3 (page 3).

Other forms of sounding instruments include dropsondes, dropped from aircraft and used mainly for research purposes, and rocketsondes for the measurement of temperature profiles at very high levels. Temperature profiles throughout a substantial depth of the atmosphere can be deduced from satellite observations of infra-red radiation. Although the soundings lack detail, and the uncertainties and errors are greater than for 'conventional' soundings, the observations are proving valuable, particularly in data-sparse areas.

Of the above, the forecaster will receive almost all his upper-air information from radiosondes, supplemented in the British Isles for local forecasting purposes by the tethered-balloon temperature and humidity profiles (BALTHUM) from Cardington, Bedfordshire, and possibly by data from one or two high television masts.

10.2.2 Wind

The most frequent method of wind measurement is by tracking a radar reflector attached to a rising balloon. When a radiosonde ascent is being carried out, the radar reflector is attached to the same balloon, and the heights to which the winds are referred are usually those calculated from the sonde data. At other times, when only winds are being measured, the heights are calculated from the radar observations. The errors of radar-wind observations are discussed in some detail in Chapter 11 — Upper-air charts.

Wind observations from aircraft may often be of value, particularly when they are made by Doppler radar. Winds measured by this method are means over a short period, and may be taken as spot winds; they are more accurate than those derived from the vector difference between the estimated air track and the observed ground track of the aircraft. Difficulties sometimes arise in using the observations, since they are not usually made at standard levels or times. Aircraft winds are of greater value in the analysis of upper-level flow (see Chapter 11) than in the study of the vertical structure at a given location.

Winds may also be measured by visual observations of a free balloon by theodolite. These pilot-balloon observations are not regularly made in the British Isles or Europe where the radar-wind network is considered adequate for synoptic purposes, but in many other parts of the world they are still a regular and essential part of the daily observational routine. Winds may also be measured by means of radar observations of falling 'chaff', (this technique is normally used at very high levels, the chaff being released from a rocket), or by observations of constant-level balloons.

10.3 ACCURACY OF RADIOSONDE OBSERVATIONS

10.3.1 Sources and types of error

The error of a given pressure, temperature or humidity observation depends upon many factors. Differences between any two sondes can exist because of design differences, slight changes from one production batch to another, variations of solar altitude and the albedo of the underlying surface, and differences in handling from station to station. For this reason, when considering sonde performance it is necessary to avoid combining data from what are really different populations. For example, by confining attention to sondes of a single design flown in darkness from one station, only the possible effects of different production batches and of altitude above the earth have to be considered. Such sondes comprise a single population. If other sondes made to the same design are flown in daylight, they will exhibit additional errors which will depend upon the altitude of the sun and the surface albedo. The presence of these 'radiation errors' forms a different sonde population. For a given population of sondes the difference of their mean from the truth is termed the 'systematic error'.

The values observed by an individual instrument will differ from the population mean: Harrison[1] termed this the 'sonde error'. From some aspects the sonde error may be looked upon as a systematic error, as it changes only gradually during a flight, but from the synoptician's point of view it is a 'random error', changing in a random way from sonde to sonde of a given population.

In addition to the sonde error, which is highly correlated from level to level, there are, on any given ascent, random fluctuations from level to level which are a result of small-scale inhomogeneities in the atmosphere and of random errors of the radiosonde itself. The individual values show a scatter which is regarded as superimposed on a smoother atmospheric structure which the sonde is trying to measure; this random error contributes to the errors of the turning-point data selected to represent the atmospheric profile but largely cancels out in the calculation of geopotential.

10.3.2 The assessment and magnitude of the errors

The systematic error of a sonde population can be assessed only by comparison with an independent and accurate source of data. No such source has yet been found, but special 'reference' sondes have been designed (for temperature only) which are of greater accuracy than routine radiosondes. The next step logically would be to compare reference sondes with operational sondes flown together in single assemblies. Unfortunately the effort and cost of mounting special comparisons preclude enough work of this kind being done.

It would be sufficient for operational purposes to ascertain the systematic differences between given sonde populations, rather than their departures from truth. The direct way would entail flying together sondes drawn from the populations concerned. The effort and cost of such work are prohibitive. Moreover, the results of such trials are of use only in finding population differences that are not time-dependent. For the systematic differences between, say, sondes flown at night and by day, or at two different stations, other techniques must be sought.

Hawson and Caton [2] developed a method whereby sondes flown simultaneously at different places can be compared, using chart analysis to remove the atmospheric differences. They also used a method whereby a time-series of soundings at one station can be used to estimate the effect of direct solar radiation and to observe in part the effect of broad changes in albedo. These methods are generally applied to geopotential data and are discussed at greater length in Chapter 11 – Upper-air charts.

Systematic differences between sonde populations vary considerably, but the largest systematic difference between routine sonde designs flown at night over Europe is equivalent to an error in mean temperature between the surface and the 100-millibar level of about one degree Celsius, with the value for most sondes being less than 0.7 degree. For the present (1976) sondes of France and the Federal Republic of Germany the mean temperature difference is probably less than 0.15 degree, with the U.K. Mark 3 sonde showing differences of similar magnitude. The data given above refer to night-time soundings: in day-time the sondes are subjected to heating by solar radiation and the differences between them are unlikely to be the same.

Knowledge of the 'sonde error' is needed when comparing results obtained from individual instruments of a given population. The sonde error of the U.K. Mark 2B sonde was determined by Harrison [1] from a series of twin flights; Table 10.1 shows the results.

TABLE 10.1 U.K. radiosonde Mark 2B

	Sonde error (standard deviation of a single value)					
Pressure level (*mb*)	700	500	300	200	100	80
Pressure (*mb*)	4	5	7	8.5	9	9.5
Temperature (*tenths degC*)	4	5	8	8.5	8.5	8.5
Relative humidity (*per cent*)	4	4	5	-	-	-
Geopotential (*geopotential metres*)	4	8.5	19	29	40	45

More recent work by C.L.Hawson (unpublished) suggests that the sonde error of geopotential is now about 30 metres at the 100-millibar pressure level.

Similar data, derived by A.H. Hooper (unpublished) for a limited series of twin flights, for the U.K. Mark 3 sonde (scheduled for service from 1976) are given in Table 10.2.

TABLE 10.2 U.K. radiosonde Mark 3

	Sonde error (standard deviation of a single value)				
Pressure level (*mb*)	500	300	100	50	30
Pressure (*mb*)	3	1	1	0.5	0.5
Temperature (*tenths degC*)	1	2	2	2.5	3.5
Geopotential (*geopotential metres*)	3	6	9	13	21

The humidity errors of the Mark 3 sondes are expected to be similar to those of the Mark 2B.

As a guide to the performance of other sondes it is noted that sonde errors of geopotential at the 100-millibar pressure level over Europe at night, determined by Hawson for May 1973, ranged from 19 to 39 geopotential metres (standard deviations).

It was mentioned in 10.3.1 (page 3) that the 'random error' contributes to the errors of the turning points selected to represent the atmospheric profile, and hence to errors of the lapse rate within the layers bounded by those turning points. The random error from one data point to another of a single ascent of the U.K. Mark 2B, determined by Harrison,[1] is given in Table 10.3.

TABLE 10.3 U.K. radiosonde Mark 2B

	Random error (root-mean-square value of single differences)							
Pressure level (*mb*)	850	700	500	300	200	150	100	80
Pressure (*mb*)	2	2	2	2	1.5	1.5	1	1
Temperature (*tenths degC*)	3	3	3.5	3.5	3.5	3	3.5	3.5
Relative humidity (*per cent*)	5	3.5	2.5	1	-	-	-	-

10.4 THE TEPHIGRAM

10.4.1 General remarks

The simplest way to represent the variation of temperature with height is in the form of a temperature/height curve. This is not a thermodynamic diagram but it is nevertheless a convenient tool to use at times: for example, the fine detail given on a tethered-balloon ascent (BALTHUM) is more usefully displayed for some purposes on a rapidly-constructed temperature/height graph than on any other available printed diagram.

For most purposes, however, it is much more advantageous to use a diagram based upon the thermodynamic properties of air and water vapour. Such a diagram is known as a thermodynamic diagram; the one most commonly used in the United Kingdom is the tephigram (Metform 2810). Another, used mainly at radiosonde stations in the computation of the heights of isobaric surfaces, is the Vaisala diagram, which uses temperature and the logarithm of pressure as co-ordinates.

On all the thermodynamic diagrams used in meteorology, the temperature and dew-point (as a measure of humidity) may be plotted at given heights or pressure levels. The lapse rates of temperature and dew-point are clearly displayed.

10.4.2 Theoretical considerations

Before discussing the function and uses of the tephigram it will be helpful to review the fundamental theory underlying its construction.

We may start with the first law of thermodynamics which may be expressed in the form:

$$dQ = c_v dT + dw = c_v dT + p d\alpha \qquad . \ . \ . \ . \ (10.1)$$

and the equation of state of a perfect gas:

$$p\alpha = RT \quad . \quad . \quad . \quad . \quad (10.2)$$

where dw is the work done by unit mass of air on its environment, dQ denotes the heat communicated to unit mass, c_v the specific heat of dry air at constant volume, T thermodynamic temperature, p pressure, α specific volume and R the specific gas constant for air. We may assume that the constants refer to dry air at this stage. If the air contains water vapour certain modifications are required. These will be discussed in a later paragraph.

If equation 10.1 is divided through by T it follows that:

$$\frac{dQ}{T} = c_v \frac{dT}{T} + p \frac{d\alpha}{T} \quad . \quad . \quad . \quad . \quad (10.3)$$

The term dQ/T which appears on the left-hand side of equation 10.3 is important in relation to the tephigram, and is defined as the change of entropy of the air. The integral $\int dQ/T$ round a closed cycle of reversible changes is zero and so $\int dQ/T$ from an arbitrary initial state can be used to define the specific entropy of the system. It is seen that the specific entropy increases or decreases according to whether heat is absorbed by or removed from the parcel considered. Entropy is often denoted symbolically by ϕ.

One of the assumptions that may be made in considering atmospheric behaviour is that vertical motion is adiabatic. This means that $dQ = 0$, so that equation 10.3 becomes:

$$c_v \frac{dT}{T} + p \frac{d\alpha}{T} = 0 \quad . \quad . \quad . \quad . \quad (10.4)$$

or, putting $T = p\alpha/R$,

$$c_v \frac{dT}{T} + R \frac{d\alpha}{\alpha} = 0 \quad . \quad . \quad . \quad . \quad (10.4a)$$

Differentiating equation 10.2,

$$\frac{dp}{p} + \frac{d\alpha}{\alpha} = \frac{dT}{T} \quad . \quad . \quad . \quad . \quad (10.4b)$$

and substituting for $d\alpha/\alpha$ in equation 10.4a yields:

$$c_v \frac{dT}{T} + R \frac{dT}{T} - R \frac{dp}{p} = 0,$$

whence, since $c_v + R = c_p$,

$$\frac{dQ}{T} = c_p \frac{dT}{T} - R \frac{dp}{p} = 0 \quad . \quad . \quad . \quad . \quad (10.5)$$

where c_p is the specific heat of dry air at constant pressure.

Integrating

$$\int \frac{dQ}{T} = \int \left(c_p \frac{dT}{T} - \frac{R dp}{p} \right) = \int d(\ln T^{c_p} p^{-R}) = \text{constant}$$

or

$$T^{c_p} p^{-R} = \text{constant}$$

or

$$T \propto p^{R/c_p} \quad . \quad . \quad . \quad . \quad (10.6)$$

which provides the basic relation between T and p during an adiabatic process. Thus

$$\theta = T \left(\frac{1000}{p} \right)^{R/c_p} \quad . \quad . \quad . \quad . \quad (10.7)$$

where θ denotes the potential temperature, which is defined as the temperature assumed by a parcel of air when that parcel is expanded or compressed adiabatically to a pressure of 1000 millibars. A value of the potential temperature therefore defines a given adiabatic process. The potential temperature must remain constant during an adiabatic process.

If equation 10.7 is differentiated logarithmically, one obtains from equation 10.5:

$$\frac{dQ}{T} = d(c_p \ln\theta) \qquad . \quad . \quad . \quad (10.8)$$

10.4.2.1 The axes of the tephigram. We may let one co-ordinate of the diagram represent temperature. This is logical since temperature is one of the measured quantities which it is required to analyse. The other co-ordinate may be chosen so that it represents $c_p \ln\theta$. Since $c_p \ln\theta$ is constant during an adiabatic process, the straight lines $c_p \ln\theta$ = constant parallel to one axis are also dry adiabats and intersect the isotherms. It is now seen why the tephigram is so called. One co-ordinate is temperature and the other entropy. Since the latter quantity is often denoted by the Greek letter ϕ (phi) the name of tephigram developed into popular usage.

Equation 10.7 can be used to compute the pressure as a function of the temperature T and potential temperature θ, and the corresponding pressures can be displayed on the tephigram by a system of pressure lines (isobars) which are slightly curved and run obliquely across the axes of θ and T.

The thickness values of the consecutive 100-millibar layers bounded by the standard pressure surfaces 1000, 900, 800 mb etc. may be calculated from the thickness equation (see section 2.2.6.1 of Chapter 2 – Dynamical ideas in weather forecasting):

$$z_2 - z_1 = R\overline{T} \ln {}^{p_1}/{}_{p_2} \qquad . \quad . \quad . \quad (10.9)$$

where z_1, z_2 are the lower and upper heights of the corresponding pressure surfaces p_1, p_2, and $\overline{T}$ is the mean virtual temperature of the layer which may be determined from the isotherm which cuts the actual temperature sounding in such a way as to form two equal areas, the lower of which is bounded by the isotherm and the isobar p_1, and the upper by the isotherm and the isobar p_2. In Figure 1 the dashed line T_1 T_2 represents the actual temperature sounding. $\overline{T}$ is the appropriate isotherm, the value of which, when substituted in equation 10.9, gives the thickness $z_2 - z_1$ of the layer bounded by p_1 and p_2.

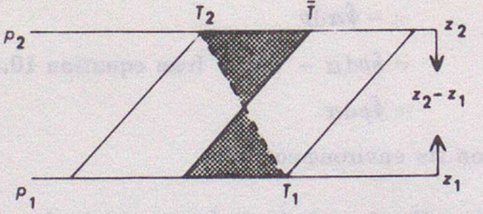

FIGURE 1. Calculation of the thickness of a layer of the atmosphere

The values of the thickness of layers between consecutive 100-mb isobaric surfaces are indicated on Metform 2810A (1956 edition) up to 300 mb by numbers on the 50-mb line between them. Each number gives the thickness in metres of an isothermal layer which is of the same thickness as the layer drawn on the tephigram. Above 300 mb the thicknesses of 50-mb layers are indicated on the 275, 225, 175 and 125-mb isobars.

The tephigram also possesses a nomogram for finding the distance of the 1000-mb surface above or below the point where the pressure and temperature are known. If the latter are at mean sea level then, if the mean-sea-level pressure is greater than 1000 mb, the thickness of the layer referred to represents the height of the 1000-mb surface above mean sea level. If the mean-sea-level barometric pressure is less than 1000 mb the thickness of the imaginary layer concerned must be subtracted from the thickness of the upper layer, say 1000 – 900 mb to obtain the height above mean sea level of the upper isobaric surface, that is, the 900-mb surface. Thus for a surface temperature of, say, 20 °C a line may be drawn connecting 20°C on the upper horizontal temperature scale with, say, 1020 mb on the vertical inside ordinate scale on the left. The line may be drawn through this point so that it intersects the outside vertical ordinate height scale. The height read off the latter scale is then 560 feet or 170 metres. If the pressure on the inner pressure scale is taken to read 980 mb instead of 1020 mb the height value must be subtracted from, instead of added to, the upper thickness layer to obtain the height of the upper isobaric surface.

The dry-adiabatic lapse rate represented on the tephigram by lines of constant potential temperature can be used to calculate the thicknesses also. The dry-adiabatic lapse rate is 9.8 degC per kilometre. Thus, if 9.8 degC is measured along a dry-adiabat, the isobars intersecting the chosen segment of the dry-adiabat will include a layer of 1 kilometre. This device can be used on any part of the diagram to deduce thickness values.

The figures on the left of Metform 2810 represent the height of the particular pressure level in the International Standard Atmosphere above 1013.25 mb. This atmosphere is constructed on the basis of assuming specified values for the pressure and temperature at each kilometre of height above the reference level. The heights are given in feet and in metres. These provide a useful approximate scale of height.

10.4.2.2 Representation of work on a tephigram. One of the most significant properties of most diagrams used to represent the thermodynamic properties of the atmosphere is that the heat absorbed by or work done upon a volume of air during a closed cycle of operations can be represented by an area on the diagram. Thus, it is readily seen that if the integral $\int T \mathrm{d}(\ln p)$ is evaluated round a closed curve,

$$\oint T \mathrm{d}(\ln \theta) = c_p \oint \mathrm{d}T - \oint RT\,\mathrm{d}p/p \qquad \text{by substitution from equation 10.5.}$$

The first integral on the right-hand side vanishes round a closed curve and the latter may be evaluated by substitution from equation 10.2. Thus,

$$\begin{aligned} \oint T \mathrm{d}(\ln \theta) &= -\oint RT\,\mathrm{d}p/p \\ &= -\oint \alpha \mathrm{d}p \\ &= \oint p \mathrm{d}\alpha - \oint R \mathrm{d}T \quad \text{from equation 10.4b} \\ &= \oint p \mathrm{d}\alpha \qquad \qquad \ldots . \quad (10.10) \end{aligned}$$

and this is the work done by the air on its environment.

10.4.2.3 The dry-adiabatic lapse rate. The dry-adiabatic lapse rate is the rate at which dry air cools with height as it ascends to levels where the pressure is lower. It is assumed that the air undergoes an adiabatic process. The values can be derived from equation 10.5, together with the hydrostatic equation and the equation of state.

Thus
$$\Gamma_d = g/c_p = 9.8 \text{ degC per kilometre} \quad \ldots . \quad (10.11)$$

where Γ_d is the dry-adiabatic lapse rate.

The dry-adiabatic lapse rate is an important parameter in the analysis of upper-air soundings. The evaluation of upper-air soundings on the tephigram relies among other things on a comparison of the environment curve and the dry-adiabatic curve.

10.4.2.4 Representation of water vapour. In the previous section some theoretical considerations upon which the construction of the tephigram depends have been outlined. However, the discussion so far is related only to dry air; no provision has yet been made to include the measurements of water vapour on the diagram. But the air in the atmosphere is never completely dry at the levels with which the tephigram is concerned. It always contains some water vapour.

Water vapour behaves in the same way as other gases. The main difference is that water substance liquefies and solidifies within a range of temperatures which occur commonly.

There are several ways in which the amount of moisture in the air can be expressed. The basic measurement is the mixing ratio which is the mass of water vapour contained in unit mass of dry air. Thus if r is the mixing ratio, $r = M_v / M_d$ where M_v is the mass of water associated with M_d, the mass of dry air considered. The mixing ratio is usually quoted in units of g/kg for convenience.

Another measurement of the humidity is the specific humidity. This is very similar to the mixing ratio and is the mass of water vapour contained in unit mass of moist air. Thus, $q = M_v / (M_v + M_d)$ where q is the specific humidity. The specific humidity is not used as much as the mixing ratio, principally because, when changes occur, the denominator changes as the amount of water vapour varies.

The mixing ratio and specific humidity are measures of the absolute humidity, that is, of the actual amount of water vapour contained in the air.

Another important measure of humidity is the vapour pressure, usually denoted by e. The vapour pressure is the partial pressure of the water vapour in the air. Thus $p = p_d + e$ where p is the total pressure of the atmosphere and p_d is the partial pressure of dry air, that is, the sum of the partial pressures of oxygen, nitrogen and the rarer gases in the air.

Clear distinction should be made between the water-vapour parameters defined above, which measure the actual water content of the air and the water-vapour parameters which are measures of the water-vapour capacity of the air and are thus independent of the actual water-vapour content. The latter parameters are a function of the temperature and measure the maximum amount of water vapour contained in unit mass of air when the space is saturated. The saturation mixing ratio r_s is therefore the mass of water vapour contained in unit mass of dry air when the space is saturated. The saturation vapour pressure e_s is the vapour pressure when the space is saturated. Both r_s and e_s increase with temperature.

A new parameter may now be introduced. This is the relative humidity, which is the ratio of the actual vapour pressure at a certain temperature to the saturation vapour pressure at that temperature. Thus the relative humidity is expressed by e/e_w. Clearly the relative humidity increases if the temperature decreases, provided the total vapour content remains constant.

The saturation-mixing-ratio isopleths are printed on the tephigram. They are the pecked lines which slope upwards to the right, and are labelled in grams of water vapour per kilogram of dry air.

Further moisture parameters are described in terms of various defined temperatures. The most important of these definitions are:

The dew-point temperature is the temperature at which a parcel of air would become saturated if it were cooled at constant pressure without any change in the total water-vapour content. The values of humidity as measured by the radiosonde are converted into dew-points in the coded weather message. The dew-points are then plotted on the tephigram at the appropriate pressure for the level concerned. Each dew-point corresponds to a point on the sounding curve. The latter points are, of course, identified by a pair of observations of pressure and temperature. When all the corresponding dew-points are plotted a dew-point curve is drawn through them. This curve, which lies to the left of the temperature curve, indicates the vertical distribution of moisture in the air column.

The wet-bulb temperature is the lowest temperature to which a sample of air may be cooled by evaporating water into it. The difference between the dry-bulb and the wet-bulb temperatures gives a measure of the relative humidity.

The virtual temperature is a temperature which is a function of the amount of moisture in the air. It is defined as the temperature of dry air having the same pressure and density as the moist air. Moist air may then be treated as dry air of temperature T_v. The virtual temperature T_v is given approximately by the expression

$$T_v = T(1 + 0.61r) \qquad \ldots \ldots (10.12)$$

where r is the mixing ratio already defined.

10.4.2.5 Adiabatic processes for saturated air. Moist air, that is air which, although containing water vapour, is not saturated, is normally treated as dry air in so far as the effects of undergoing the adiabatic process are concerned. There is a small difference arising from the differing specific heats of air and water vapour, but this is usually neglected. However, as soon as the air becomes saturated the dry-adiabatic process no longer applies because of the latent heat released during condensation of water vapour.

If a sample of saturated air is expanded it will remain saturated but some of the vapour will condense in the form of water or ice. If the sample was in a thermally insulated container the products of condensation would remain in the system. If the air was now compressed the products of condensation would evaporate and the initial conditions of pressure and temperature would be reached. Such a process would therefore be reversible and one of this kind is therefore called a reversible saturated-adiabatic process.

But supposing a sample of saturated air is expanded and the condensation products are allowed to fall out of the system. If the sample is now compressed to its initial pressure the temperature will be much higher than in the previous case since the return to the initial pressure will be a dry-adiabatic process. Such a process in which all the condensation products fall out is called a pseudo-adiabatic process.

The curve for the saturated-adiabatic process represents the temperature and pressure of a parcel of saturated air undergoing a reversible saturation process. The change in temperature with height or pressure is made up of two parts. The first part may be regarded as the change in temperature which would occur if the air were dry and undergoing a dry-adiabatic process. The second part may be regarded as the change in temperature brought about by the latent heat of condensation or evaporation which is released or taken up by the water content of the air as the water substance changes phase, either from the vapour state to the liquid or solid state in the case of condensation or from the liquid or solid state to the vapour state in the case of evaporation or sublimation. The saturated-adiabatic curves may be constructed from a series of points which may be computed from the exact mathematical equation which describes the saturated-adiabatic process. The development of this equation is somewhat complex and it will not be given in this handbook.*

The saturated adiabatics are printed on the tephigram by lines that curve upwards to the left, eventually becoming parallel to the dry adiabatics as the water vapour content at upper levels of the atmosphere becomes very small.

The saturated adiabatics may be used to find the wet-bulb potential temperature of the air at a given pressure and temperature. The wet-bulb potential temperature bears the same relation to the wet-bulb temperature as the potential temperature does to the dry-bulb temperature and is obtained by following the saturated adiabatic to 1000 millibars.

10.4.2.6 Normand's theorem. Normand[3] derived a very useful theorem relating to a property of the tephigram. This is based on the fact that the wet-bulb temperature is conservative with respect to evaporation or condensation when the latent heat is supplied by, or used to heat, the air. Normand showed that the dry-adiabatic line through the dry-bulb temperature, the saturated adiabatic through the wet-bulb temperature, and the dew-point line through the dew-point all meet in a point.

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* See 'International Meteorological Tables', Wld Met Org, Geneva, No. 188, TP 94, 1966, Introduction to Tables 4.4 and 4.13.

The application of this theorem to give wet-bulb temperature at various levels so that variation of wet-bulb potential temperature can be seen is very important and should be an integral part of the analysis of all temperature soundings. The wet-bulb potential temperature remains invariant during adiabatic or pseudo-adiabatic changes and is a useful parameter to indicate different air masses.

10.4.2.7 Plotting representative environment curves. The first operation in using the tephigram as a tool for the analysis of the vertical structure of the atmosphere is to plot the set of points defined by corresponding pressures and temperatures and draw a line through these points, smoothing any minor irregularities which show evidence of observational errors or coding mistakes. The shape of the curve may then be studied and information obtained about the lapse rate or decrease of temperature with height (decreasing pressure). Normally, the plotted curve shows a steady decrease of temperature with height. Occasionally, however, an increase of temperature with height occurs throughout a relatively shallow layer. Such reversals of the usual lapse rate are called inversions.

The temperature sounding or environment curve may then be compared with the dry-adiabatic or saturated-adiabatic curve. In order to make the latter comparison it is first necessary to plot the humidity values. These are defined by sets of dew-points corresponding to the pressures and temperatures already plotted to define the temperature curve. Temperature curves are usually drawn as solid lines. Dew-point curves are normally drawn as pecked lines.

Figure 2 illustrates an upper-air ascent on the tephigram. The curve shows a shallow inversion or increase of temperature with height near the surface, followed by a shallow isothermal layer to about

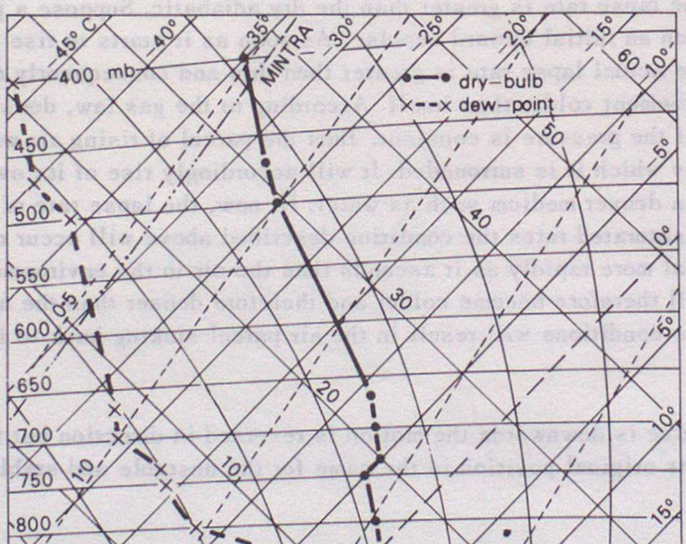

FIGURE 2. An ascent plotted on a tephigram section

960 mb. Above this, the temperature decreases quite markedly with height up to 750 mb where there is a further shallow isothermal layer. Above 750 mb the curve follows the saturated adiabatic very closely up to 500 mb, after which the temperature falls off less rapidly than before, whereas the saturated adiabatic tends to approach the dry adiabatic. The dew-point curve shows saturation near the surface with an inversion of the hydrolapse rate up to about 980 mb. Above this level the water vapour content decreases fairly rapidly as shown by the sharp dew-point lapse rate.

10.4.3 Stability and instability

10.4.3.1 Definition. The concept of stability and instability is one of the fundamental ideas which describe the processes at work in the atmosphere. A simple definition of this concept is as follows:

An element of mass is said to be in stable equilibrium if, on being given an initial impulse, the element returns towards its former position. It is said to be in unstable equilibrium if, on being given an initial impulse, it is accelerated along a path away from its initial position.

In the atmosphere the concept of stability or instability is usually referred to a layer of air or to an air mass. It is a measure of the behaviour of a parcel of air displaced upwards or downwards and depends on the lapse rate or decrease of temperature with height. If the actual lapse rate is greater than the dry adiabatic the air is unstable, whether dry or saturated. If the actual lapse rate is between the dry- and saturated-adiabatic lapse rates the air is stable if dry and unstable if saturated. If the actual lapse rate is less than the saturated-adiabatic lapse rate the air is stable whether dry or saturated. The reasons for these states of stability and instability are quite clear.

In the first case suppose the lapse rate is greater than the dry adiabatic. Suppose a parcel or bubble of air near the surface is given an initial upward impulse. As soon as it starts to rise it will cool at the dry-adiabatic rate. But the actual lapse rate is greater than this and consequently the parcel will almost at once be in an environment colder than itself. According to the gas law, density is inversely proportional to temperature, if the pressure is constant. Thus the parcel of rising air will be warmer and less dense than the air by which it is surrounded. It will accordingly rise of its own accord just as a bubble of air rises through a denser medium such as water. If, now, the lapse rate of the environment lies between the dry and the saturated rates the condition described above will occur only for saturated air, since dry air will now cool more rapidly as it ascends than the air in the environment. Such a parcel or bubble of dry air will therefore become colder and therefore denser than the air surrounding it. An initial impulse under these conditions will result in the air parcel sinking back to its original position. It will be stable.

Of course if the initial impulse is downwards the motion is reversed in direction but the tendency to move away from or return to its original position is the same for the unstable and stable cases, respectively.

The energy of instability is proportional to the area bounded by the environment curve and the process curve defining the adiabatic-lapse rate for dry or saturated air as the case may be (10.4.2.2, page 8).

10.4.3.2 Unconditional instability. If the actual lapse rate is greater than the dry-adiabatic lapse rate up to a height at which rising air will become saturated and is then not less than the saturated adiabatic, the atmosphere within the layers concerned is said to be unconditionally unstable.

An example is shown in Figure 3. The energy of instability will be released without any restrictive condition. It is, however, unlikely for such a pronounced lapse rate to occur, except perhaps at low levels over very warm surfaces.

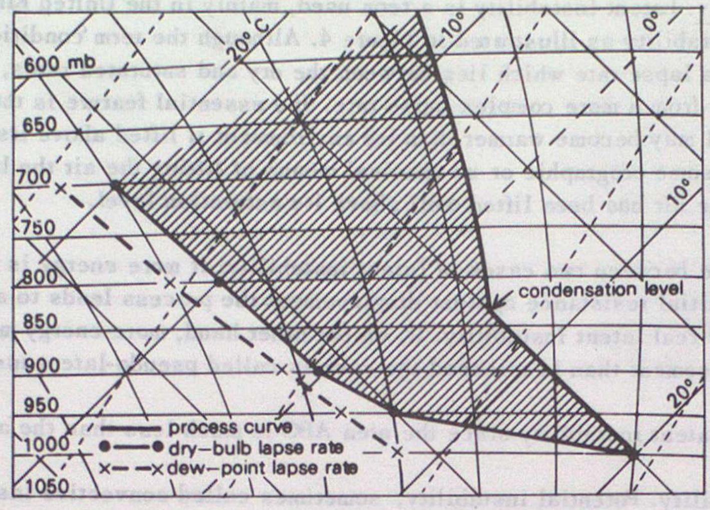

FIGURE 3. Unconditional instability

10.4.3.3 <u>Conditional instability</u>. If the lapse rate as shown by the environment curve lies between the values for dry air and saturated air the atmosphere within the layers considered is said to be conditionally unstable. An example is shown in Figure 4. An examination of the environment curve shows that, in

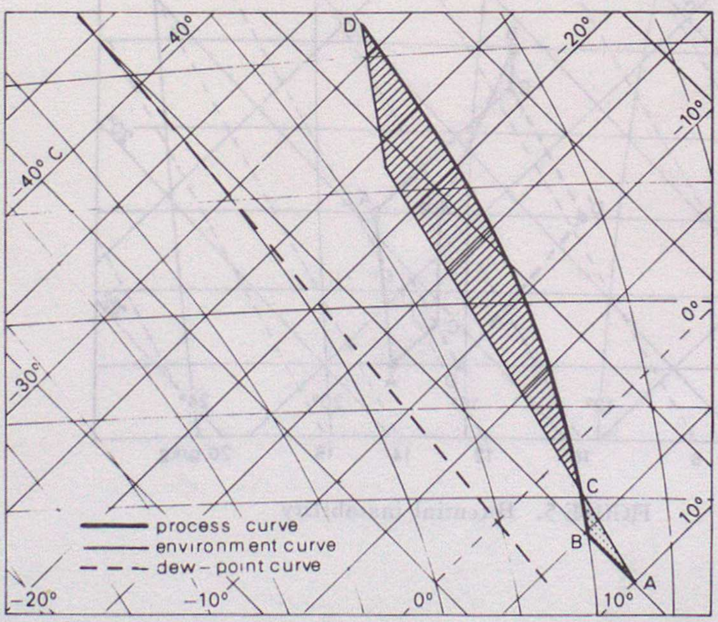

FIGURE 4. Conditional instability

the lower part of the sounding, the lapse rate lies between the dry and saturated values. Air forced to ascend from A will, it is assumed, become saturated at B, at which point the actual mixing ratio of the air at A is also the saturation mixing ratio; above B the process curve follows the saturated-adiabat marked by the thick line BC. The striped area represents the energy of conditional instability which would be released if air at A was lifted to C and then allowed to rise freely to D. The dotted area represents the energy required to lift air from A until it becomes warmer than the environment after passing C.

10.4.3.4 Latent instability. Latent instability is a term used, mainly in the United Kingdom, to define the state of conditional instability as illustrated in Figure 4. Although the term 'conditionally unstable' is normally used to define a lapse rate which lies between the dry and saturated rates, conditional or latent instability may arise from a more complex lapse rate. The essential feature is that the air is initially stable but a parcel may become warmer than its environment if lifted above its saturation level. In the case where there is some orographic or mechanical means of lifting the air the latent instability may not be realized until the air has been lifted well above its saturation level.

A distinction may be made between two cases of latent instability. If more energy is released than is required to overcome the initial resistance against displacement the process leads to a net gain of energy. This case is called real latent instability. If, on the other hand, more energy must be used to overcome the initial displacement than is released the case is called pseudo-latent instability.

Figure 4 represents real latent instability since the area ABC is much less than the area CDE.

10.4.3.5 Potential instability. Potential instability, sometimes called convective instability, refers to the realization of instability resulting from the lifting to saturation of a whole layer of air. The discussion so far has been restricted to the ascent of parcels or bubbles of air which rise through their environment. The usual treatment of potential instability is to consider the lifting of the air at the upper and lower boundaries of some chosen layer. For example in Figure 5 one may consider the 1000-mb and 900-mb pressure levels. Air at A (1000 mb) ascends along the dry-adiabatic lapse rate to C and then along the saturated adiabatic to B where the pressure is 900 mb. Similarly, air at A′ rises until it reaches

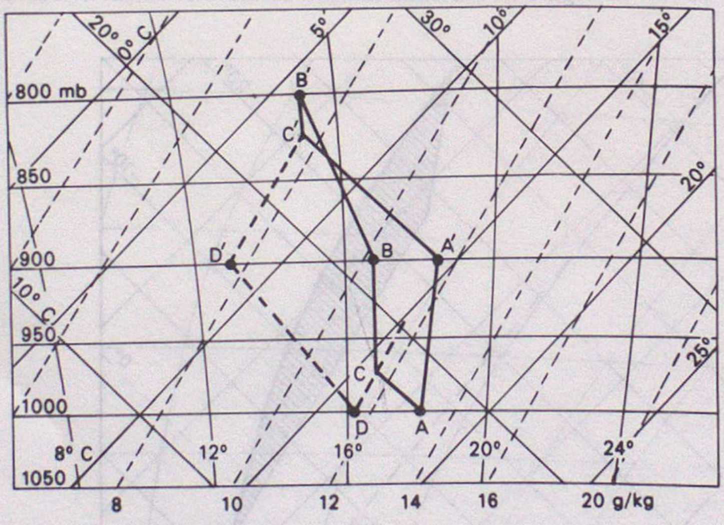

FIGURE 5. Potential instability

its condensation level at C′, and continues up to the saturated adiabatic until it reaches the 800-mb level at B′. The resulting lapse rate BB′ is steeper than the original stable lapse rate AA′. It is, in fact, now steeper than the saturated-adiabatic lapse rate. Now, since the whole 1000-900-mb layer has been lifted and so transformed to the 900-800-mb layer, it has become unstable. The effect of the lifting has been to develop instability from the original stable lapse rate. Since for instability the saturated adiabatic through B′ must be to the left of that through B it is clear that the wet-bulb potential temperature at A′ is less than at A. Thus a condition of potential instability is that the wet-bulb potential temperature should decrease with height.

Even if a layer of dry air is lifted it will change the lapse rate. This can be shown graphically on a tephigram, or mathematically. Thus, if dry air is undergoing upward motion:

(i) the layer becomes more stable if its lapse rate is greater than the dry adiabatic,

(ii) the stability of the layer is unchanged if the lapse rate is dry adiabatic,

(iii) the layer becomes less stable if the lapse rate is less than the dry adiabatic.

In the case of descent the relation is reversed thus:

(i) the layer becomes more unstable,

(ii) the stability of the layer remains unchanged,

(iii) the layer becomes more stable.

Upward motion of dry air tends to bring the lapse rate nearer to the dry adiabatic. It is only if the layer becomes saturated during the vertical motion that a previously stable lapse rate becomes unstable and potential instability is converted to true instability.

10.4.4 Condensation levels

Condensation in the atmosphere to form cloud is brought about in three main ways, by lifting, by mixing, or by convection. The tephigram is a useful tool for the assessment of when and where cloud is likely to form.

10.4.4.1 Lifting condensation level. If a mass of air is lifted bodily, for example by passing over a range of hills or mountains, or at a front, the air expands and cools adiabatically, or nearly so, until it becomes saturated. The height at which saturation occurs is called the lifting condensation level. It occurs at the pressure given by the intersection of the dry adiabatic through the dry-bulb temperature and the humidity mixing-ratio line drawn through the dew-point of the sample of air.

10.4.4.2 Mixing condensation level. In a layer of air which is thoroughly mixed, such as the surface boundary layer in moderate or strong winds, the temperature falls off with height at the dry-adiabatic lapse rate, while the humidity mixing ratio tends towards uniformity. In these conditions, the mixing condensation level is given by the intersection of the dry adiabatic through the surface temperature and the mixing-ratio line through the surface dew-point. If the layer is not mixed, but is likely to become so during the day as the boundary layer develops, the mixing condensation level may be estimated in a similar fashion, using the mean temperature and mixing ratio in the layer in which mixing is expected (see Chapter 16 – Wind).

10.4.4.3 Convective condensation level. The convective condensation level is the height at which condensation will begin when a sample of air is allowed to ascent through the environment adiabatically and without mixing, the temperature of the sample at any height being no less than that of the environment. The use of the tephigram to study convection will be discussed more fully in the next section and in Chapter 19 – Clouds and precipitation.

10.4.5 Assessment of convection

10.4.5.1 The 'parcel' method. In Figure 6 let ABCD represent the dry-bulb temperature curve at some time before convection begins. If T and Q are the expected surface temperature and dew-point at a later time, then the convective condensation level is found by drawing a dry adiabat from T and a constant mixing-ratio line from Q, to intersect at p_b. If p_b lies to the left of the environment curve, ABCD, convective cloud will not form; but if it lies to the right of ABCD, condensation will occur and the ascending parcel will thereafter follow a saturated adiabatic curve. In the parcel theory it is assumed that the 'parcel' rises without any mixing with the environment, and that it ceases to accelerate when it reaches a point, p_e, where its temperature is equal to that of the environment.

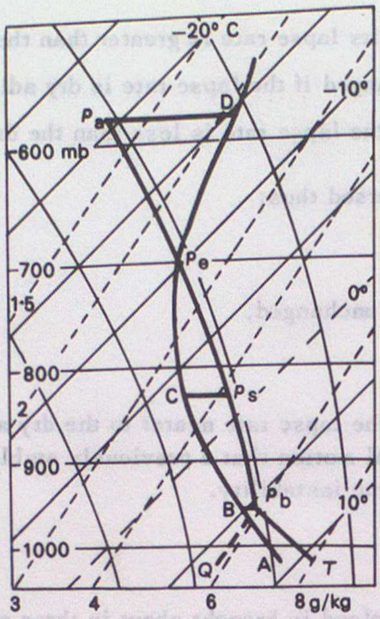

FIGURE 6. Construction on tephigram for parcel and slice methods for forecasting tops of convective cloud

ABCD: dry-bulb temperature before convection begins.
T: surface temperature expected as a result of day-time heating.
Q: surface dew-point forecast for the appropriate time.
$T p_b p_s p_e p_a$: temperature changes in a parcel of air rising from T to p_a ('process curve').

However, the parcel has an upward velocity when it arrives at p_e, the energy of motion being represented by the area between the environment and process curves. Theoretically, the parcel could rise as far as p_a, where area $p_e p_a D p_e$ (where p_aD is an isobar) is equal to area ABC$p_e p_b T$A.

10.4.5.2 The 'slice' method. When part of the atmosphere ascends in the form of convection currents there must be some compensatory downward movement of the rest of the air, the 'environment'. Adiabatic warming in the descending air leads to an increase in the temperature of the environment (except when the lapse therein is greater than the dry-adiabatic lapse rate). The excess temperature of an upward moving saturated parcel is therefore reduced, and equilibrium will be reached at a lower level than it would in an unmodified environment. The precise equilibrium point depends upon the environmental lapse rate, γ, and the ratio of the area over which convection is taking place to the area covered by the environment, but Petterssen[4] has shown that the environment is stable if

$$\gamma < \gamma_s$$

where γ_s is the saturated-adiabatic lapse rate. In Figure 6 this corresponds to p_s. In practice p_s may be difficult to determine, particularly if there are irregularities in the environment curve.

10.4.5.3 The effect of entrainment. In the parcel and slice methods for assessing the vertical extent of convection it is assumed that the bubble rises through the environment without any mixing at the boundaries. Mixing does occur, however, and acts to reduce the excess temperature and moisture content of the bubble. The effect is to reduce the vertical velocity and hence the height to which the bubble can rise; it also warms and moistens the environment, tending to stabilize it against further convection. The magnitude of the effect varies with the rate of entrainment and the size of the convective elements, and quantitative allowance is not possible in practical forecasting.

The problem of forecasting convection is dealt with in detail in Chapter 19 – Clouds and precipitation, but it can be said that on many convective days the general cloud-top height is often rather above p_s, with some tops rising as high as p_e. In severe local storms, where the updraught is 'protected' from the environment, tops occasionally reach levels near p_a (which at times may be well above the tropopause).

10.4.6 Further practical applications of the tephigram

10.4.6.1 Frontal zones. The upper-air ascent is valuable in identifying frontal zones where sudden changes in temperature, dew-point, and lapse rates may occur. It may also be used as a tool for assessing the slope of a front if the surface position is known. Figure 7 shows a characteristic lapse rate through a warm-front surface.

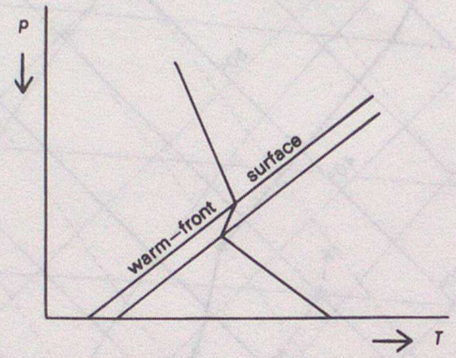

FIGURE 7. Temperature ascent through a warm front

The lapse rate below the frontal surface is saturated adiabatic or even greater. A sharp change to an inversion through the warm-front zone is typical, above which the lapse rate becomes positive in sign but less in magnitude than the saturated adiabatic; a similar diagram occurs when the ascent pierces a cold front although the lapse rate in the cold air in this case is usually steeper than in the warm-front case. An example of an ascent through a warm-front surface is shown in Figure 8 (overleaf).

10.4.6.2 Subsidence and ascent. The temperature curve often shows an inversion when subsidence is occurring. This is illustrated in Figure 9. Suppose the layer 500 mb to 600 mb is subsiding. The initial

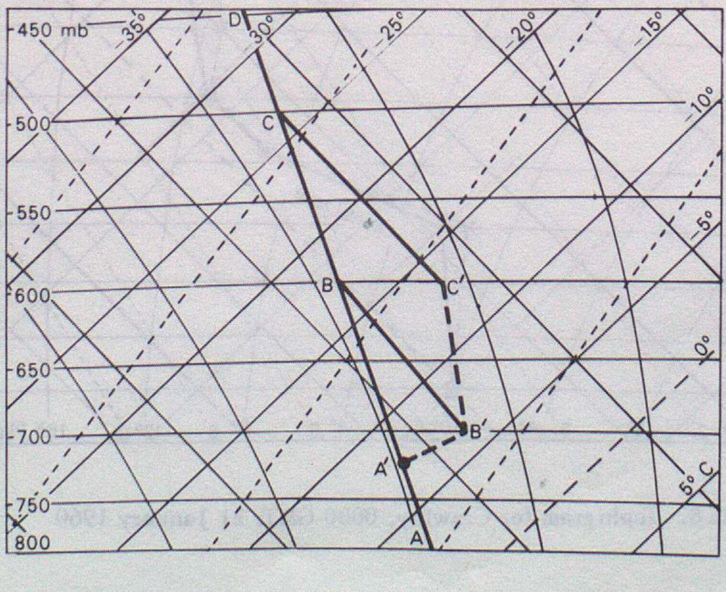

FIGURE 9. Subsidence inversion on the tephigram

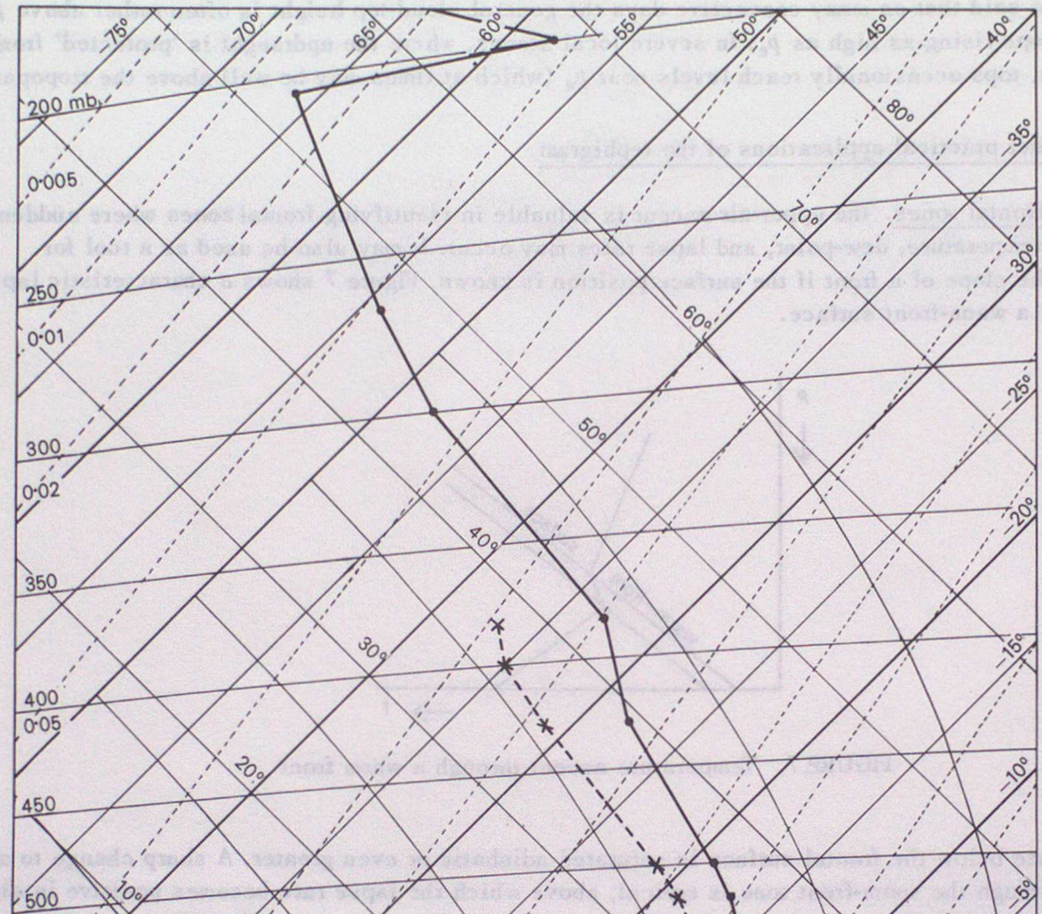

FIGURE 8. Tephigram for Crawley, 0000 GMT, 21 January 1960

lapse rate before subsidence is shown by the line ABCD. During subsidence, air at 500 mb will subside to 600 mb along the dry-adiabatic CC′. Similarly, air at 600 mb will subside to 700 mb along the dry-adiabatic BB′. If below 700 mb there was no subsidence, the initial lapse rate should exist there, and this discontinuity between subsiding and non-subsiding air would result in an inversion A′B′ and a final curve AA′B′C′CD. In practice, the rate of subsidence decreases gradually downward and a stable layer or less-marked inversion might be expected in place of the inversion A′B′. However, there is often vertical mixing in the layer AA′ and/or a stratocumulus cloud layer which loses heat by radiation. Both these factors tend to lower the temperature at A′ near the bottom of the subsiding layer and to sharpen the inversion, despite some subsidence through it. Figure 10 is a good example of subsidence.

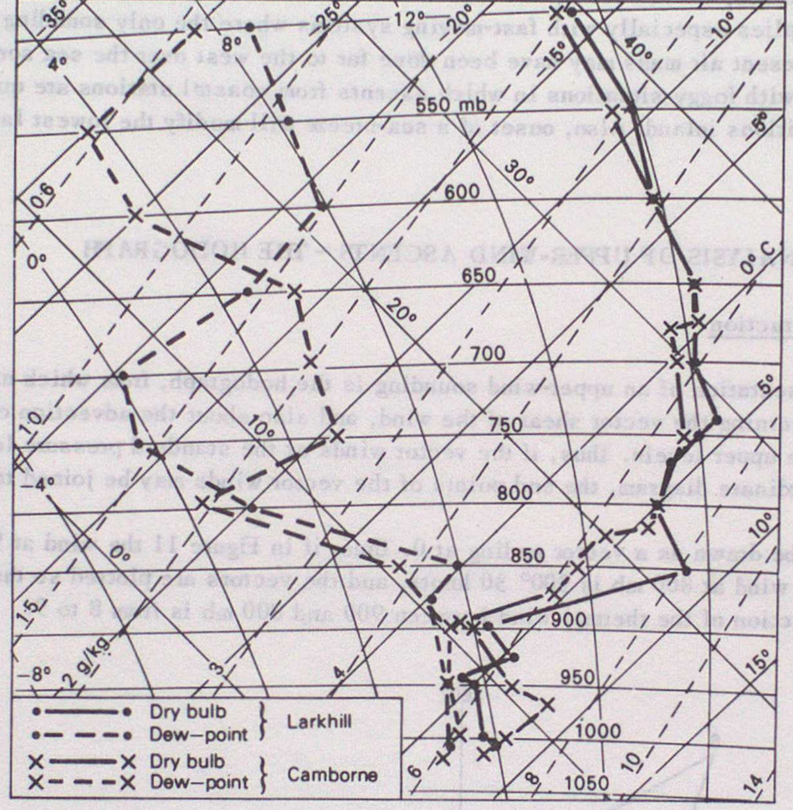

FIGURE 10. Tephigrams for Larkhill and Camborne, 0200 GMT, 29 January 1953

Air which is undergoing upward vertical motion is usually saturated throughout all or much of the layer. If, therefore, a temperature curve follows the saturated-adiabatic curve, and if this curve is overlapped or accompanied by a dew-point curve showing nearly 100 per cent relative humidities, then the layer in question is probably undergoing upward vertical motion (as in Figure 8 above 750 millibars).

10.4.6.3 Modification of tephigrams to take account of local conditions. An essential preliminary to using a tephigram for any forecasting purpose is to modify the most representative available sounding so that it is, as nearly as one can judge, completely representative of the air that is or will be over the station during the forecast period. This modification to bring the sounding into line with actual present conditions is part of the analyst's job, and simply involves taking account of present surface conditions and advection.

The procedure is:

(i) Draw in the surface isobar on the tephigram and mark in the present surface temperature and dew-point.

(ii) Connect the surface temperature and dew-point with the curves reported on the radiosonde in a manner which takes account of the prevailing conditions, that is if fog (see section 20.5 of Chapter 20 – Visibility), strong wind (see 10.4.4.2, page 15), heating or cooling (see Chapter 17 – Temperature) is present, assume an appropriate lapse rate and hydrolapse in the lowest layers.

This procedure must be adopted before any use can be made of the tephigram for forecasting temperatures, convection or fog. It applies especially with fast-moving systems where the only sounding available and representative of the present air mass may have been done far to the west over the sea some six to nine hours earlier, and also with foggy situations in which ascents from coastal stations are quite untypical of low-level foggy conditions inland. Also, onset of a sea-breeze will modify the lowest layers and affect fog forecasts.

10.5 ANALYSIS OF UPPER-WIND ASCENTS – THE HODOGRAPH

10.5.1 Method of construction

Another form of representation of an upper-wind sounding is the hodograph, from which useful deductions can be made concerning the vector shear of the wind, and also about the advection of air of different temperatures at the upper levels. Thus, if the vector winds at the standard pressure levels are plotted on a polar co-ordinate diagram, the end points of the vector winds may be joined together.

The wind vector may be drawn as a vector ending at 0. Thus, if in Figure 11 the wind at 900 mb is 240° 20 knots, and the wind at 800 mb is 300° 30 knots, and the vectors are plotted so that they terminate at 0, the direction of the thermal wind between 900 and 800 mb is from 8 to 9.

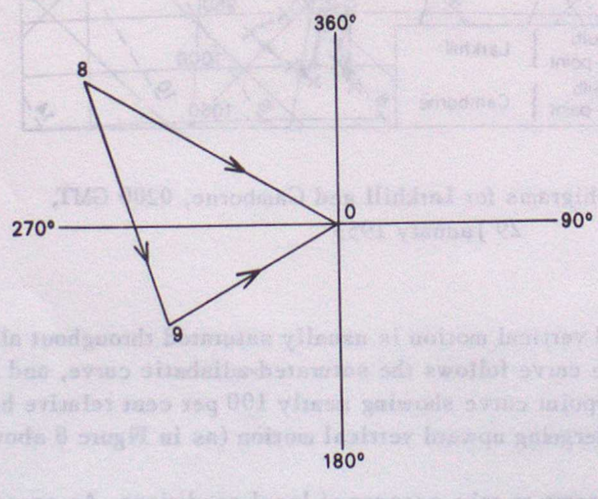

FIGURE 11. Construction of wind vectors

When plotting the complete hodograph of the upper winds at a station it is not usually necessary to draw all the lines 09, 08, etc. but simply to plot the points 9, 8, 7, ... and then join the points 8–9, 7–8, etc. to show the thermal winds between such levels. A typical hodograph, with the winds on which it is based, is shown in Figure 12.

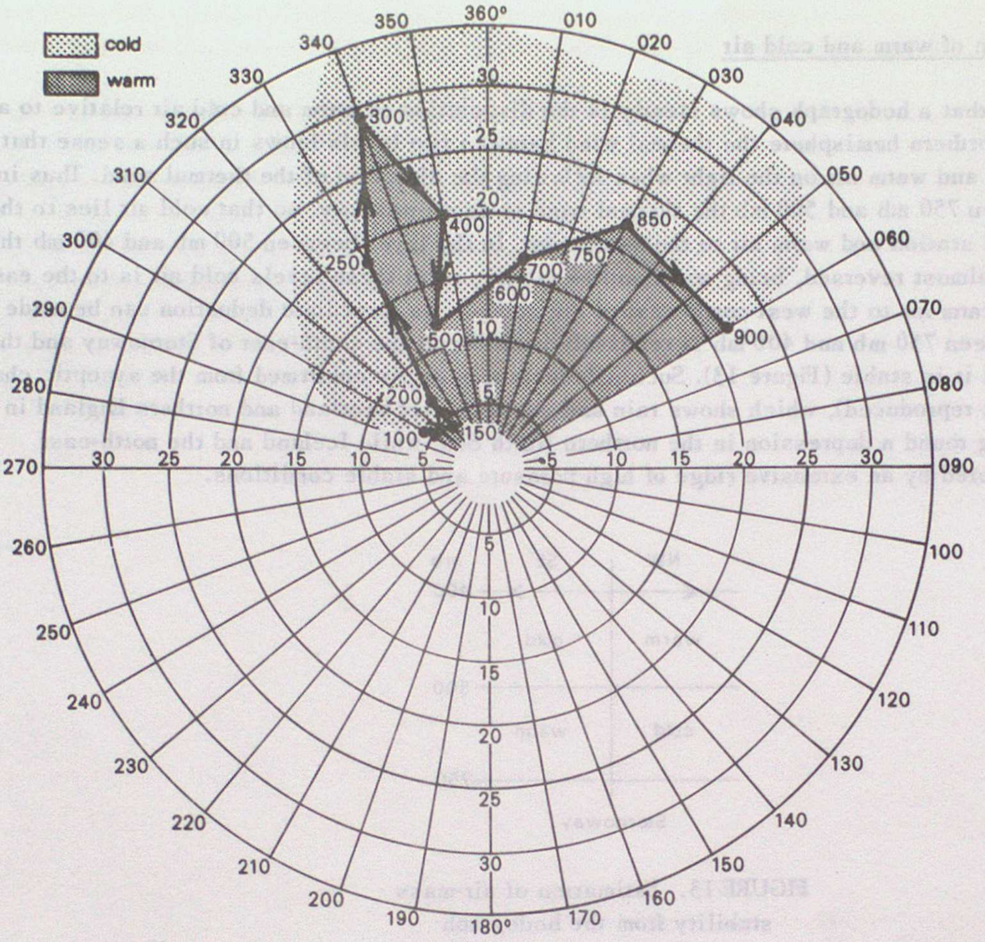

FIGURE 12. Hodograph of Stornoway ascent, 0600 GMT, 9 June 1961

mb	*degrees*	*kn*	*mb*	*degrees*	*kn*	*mb*	*degrees*	*kn*
900	060	22	600	360	14	250	330	18
850	030	22	500	340	12	200	320	7
750	020	19	400	350	20	150	320	4
700	010	17	300	340	29	100	290	6

The 900-mb wind, 060° 22 knots, is represented by the line from the point labelled 900 to the point 0; the 850-mb wind 030° 22 knots by the line from the point labelled 850 to the point 0 and so on. The thermal wind between 900 mb and 850 mb is the line drawn from the point 850 to the point 900 and in that direction. Thus the thermal winds between successive levels are clearly shown. It is also clear that the thermal wind between two widely separated levels, (such as 900 mb and 500 mb), which is frequently computed and used in the construction and use of upper-air charts, is in fact only an average or resultant thermal wind. In Figure 12 the direction of the resultant thermal wind from 900 mb to 500 mb (found by joining the points 500 and 900) is different from the direction of any of the thermal winds in the component layers. Incidentally, it may be remarked here that the difficulty of calculating a thermal wind between 1000 mb and 500 mb is considerable; even if the 1000-mb level actually exists, as it did on 9 June 1961 at Stornoway where the surface pressure was 1008 mb. On this occasion the surface wind was 360° 20 knots and the wind shear between the surface and 900 mb was considerable. Clearly the computation of a notional '1000-500-mb' thermal wind, taking the lower level as the surface, or 900 mb or 850 mb (all of which are done in practice), will lead to some big variations in possible computed thermal winds.

10.5.2 Advection of warm and cold air

The first thing that a hodograph shows clearly is the disposition of warm and cold air relative to a station. In the northern hemisphere the thermal wind between two levels blows in such a sense that cold air is on the left and warm air on the right when following the direction of the thermal wind. Thus in Figure 12 between 750 mb and 500 mb the thermal wind is south-westerly, so that cold air lies to the north-west of the station and warm air to the south-east. In the layer between 500 mb and 400 mb the thermal wind is almost reversed, being north-north-easterly, so at these levels cold air is to the east-south-east and warm air to the west-north-west of the station. An immediate deduction can be made that the air between 750 mb and 400 mb is relatively unstable to the south-east of Stornoway and that to the north-west it is stable (Figure 13). Such a deduction is amply confirmed from the synoptic chart for that date (not reproduced), which shows rain and showers over Scotland and northern England in the airstream flowing round a depression in the northern North Sea, while Iceland and the north-east Atlantic are covered by an extensive ridge of high pressure and stable conditions.

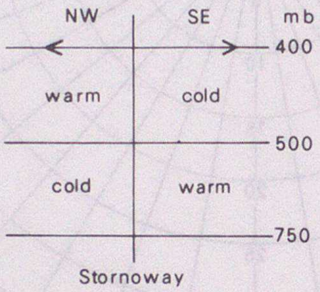

FIGURE 13. Estimation of air-mass stability from the hodograph

The hodograph also shows the direction in which the warmer or colder air is moving at any level. For example, in Figure 12, between 900 and 850 mb, the thermal wind is north-westerly and so, between these levels, colder air lies to the north-east of Stornoway and warmer air to the south-west but, since the winds in this layer are blowing from the north-east, it is clear that they will bring the colder air over the station. This is an example of 'cold advection' and always occurs when the wind backs with height, as in this example. In fact in Figure 12, cold advection is taking place at all levels up to 500 mb. Between 500 mb and 400 mb there is a slight temporary veer of wind with height and this implies the advection of warm air over the station, for the thermal wind direction is about 005°, and hence warm air lies to the west and cold air to the east of Stornoway. The actual winds through the layer are from 340° to 350°, that is, there is a small component from the west, and so a small amount of warm advection. At levels where there is no change of wind direction with height (as between 200 and 150 mb in Figure 12), the thermal wind is parallel to the actual winds so there is no warm or cold advection. At these levels the actual winds blow along the thickness lines and no advective change in temperature occurs.

10.5.3 Fronts

There are occasions when a frontal zone separates two air masses in each of which the wind may be remarkably uniform with height, so that the change from one wind regime to the other occurs almost entirely across a well-marked thin frontal zone. A hodograph of the wind structure through such a front shows very clearly the large thermal wind concentrated entirely in the frontal zone, and from such a hodograph it would be easy to identify the limits of the frontal zone. However, this is rarely found, and is probably largely confined to winter months on such occasions as when a very cold easterly continental stream flows beneath an upper south-westerly Atlantic stream. Usually the wind shear within each air mass is fully comparable with the wind shear across the frontal zone itself and even with well-marked and active fronts the winds change in the warm and cold air masses as well as across the frontal zones.

So, in general, the presence of fronts can be inferred much more reliably from the study of temperature and humidity changes than from the study of wind changes. Where the hodograph does become useful is in making deductions about the activity of a front whose presence is already known. From a study of the component of wind normal to the front, and the way in which the normal component changes with height, some inferences can be made about the vertical motion on the front.

Some practically useful deductions about the vertical motion at fronts may be made from elementary notions of the nature of frontal surfaces and the sort of vertical motion which occurs there, although Chapter 5 – Fronts and frontal weather – will show that frontal structure is more complicated. Let us imagine a sloping warm-front surface A_oB_o (in Figure 14) and suppose a vertical sounding, at time t_s, from X cuts the frontal surface at C_o. If the horizontal wind measured in the warm air at C_o exceeds the velocity of the front at this level, then the warm air must be moving up the frontal surface, otherwise the air would penetrate the front. (Similarly, an observed wind less than the speed of the front implies downward motion.) If the wind component perpendicular to the front increases with height, purely horizontal motion would be possible only if the front were moving faster aloft than near the ground, that is, its slope were decreasing. As, more usually, the slope of a front remains fairly constant and its speed does not exceed that of the winds near the ground, an increase with height of the wind perpendicular to the front can be taken as an indication that upslope motion is occurring in the warm air.

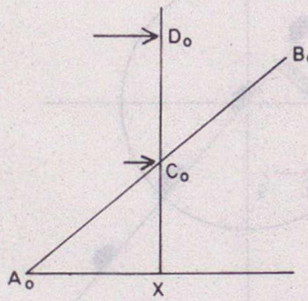

FIGURE 14. Increase with height of the component of the wind normal to a front

It is from such elementary considerations as these that we deduce the following rules:

At a warm front:

An *increase* of the wind component normal to the front with height above the frontal surface indicates that *upslope* motion of the warm air is occurring.

If the wind component normal to the front is constant then no vertical motion is occurring.

A *decrease* of wind component normal to the front indicates that *subsidence* of the warm air is occurring.

At a cold front:

A *decrease* of the wind component normal to the front in the warm air indicates that *upslope* motion is occurring.

No change in the normal component indicates no vertical motion occurring.

An *increase* in the normal component in the warm air above a cold front indicates *subsidence* of the warm air.

Fronts at which the warm air is ascending are known as 'ana fronts' and usually give much rain and thick cloud. Fronts at which the warm air is subsiding are known as 'kata fronts' and often give very little rain and only thin cloud. And despite the elementary nature of the frontal model on which it is based with its many unrealistic assumptions, the resulting ideas are very useful and easy to apply in practice and quickly supplement the temperature and humidity structure on a front as shown by tephigrams with information on the levels in which air is ascending or subsiding on a front.

The practical procedure is as follows:

(i) Determine the height of the frontal zone from inspection of tephigram.

(ii) Determine the orientation of the front from the surface chart and draw this on the hodograph (Figure 15). It is normally found that the orientation of the front is parallel to the thermal wind across the frontal zone.

(iii) Measure the wind components perpendicular to the front, in the warm air above the frontal zone - and assess the vertical motion on the front.

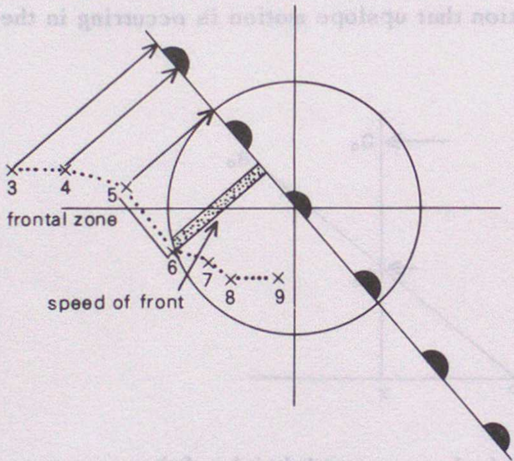

FIGURE 15. Orientation of a warm front shown on a hodograph

The speed of movement of the front, in a direction normal to its surface line of orientation, is the speed of the cold air just underneath the frontal zone. This can be read off from the same construction as that above.

These methods with a hodograph, applied to the weather ships' or west coast radiosonde ascents, can be very helpful in analysing fronts approaching the British Isles from the Atlantic in regions where ships' reports are few and the analysis of the position and speed of the front is in doubt.

It may happen that a comparatively accurate estimate of the past history of the frontal movement can be obtained from the surface charts. When this is so, and when the observed speed of the front is different (usually slower) from the computed speed from a hodograph, it usually means that the slope of the front is changing (decreasing) and the foregoing arguments about frontal ascent do not apply. As with all forecasting techniques, use of the hodograph requires some experience and skill to determine those occasions when it can be used with confidence and those on which it is inappropriate.

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