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HANDBOOK

OF

WEATHER FORECASTING

METEOROLOGICAL OFFICE

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PREFACE

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

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

CHAPTER 3

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ANALYSIS OF UPPER AIR SOUNDINGS

3.1. INTRODUCTION

Chapters 1 and 2 have dealt with the analysis of observations which have been represented on two-dimensional or topographical surfaces, mean sea level for surface analysis and various isobaric surfaces for upper air analysis. Of course isobaric surfaces are not strictly two-dimensional in space, but they may be considered so in practice because the slopes are usually very small.

This chapter will cover the analysis of observations from individual ascents in the vertical. Such a set of observations above a fixed station is sometimes called an aerological sounding. Important changes in the temperature and wind can take place over periods of six hours or less, but complete upper air soundings are normally available only at intervals of 12 hours. Vertical cross-sections embodying a graphical representation of any of the elements of a number of soundings in space or in time will not be discussed in this chapter (see Chapter 4).

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, which particular reference to upper air soundings:

(i) Demands that the analyst shall be familiar with the "normal" pattern of upper air temperature and wind soundings. After very little experience he will soon recognize that most temperature soundings show a lapse of temperature in the troposphere slightly less than the saturated-adiabatic lapse rate and approximate isothermal conditions in the stratosphere. Variations from this normal pattern are frequent and often quite considerable; many of these variations are immediately recognizable and are perfectly valid. Others would immediately strike the competent analyst as being incorrect. For example, an error of 10° in the temperature at one level is easily recognized; such errors not infrequently arise among inexperienced plotters and in transmission.

In addition to having a knowledge of the kind of values to be expected on a given occasion the analyst must have a fundamental appreciation of the accuracy to which instruments measure their values and the degree to which these fundamental measurements are processed and smoothed at the observing station before transmission. This is particularly essential with upper air observations.

With all this background knowledge, a scrutiny of one particular sounding and comparison with its predecessors from the same station and with neighbouring contemporary soundings (which forms the first step in the analysis) will readily bring to light obvious errors in temperature soundings. The detection of errors in upper wind soundings is generally much more difficult, as also is an accurate evaluation of the significance of variations from an otherwise smooth pattern of observations.

One may note 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. In the United Kingdom the average distance between neighbouring upper air stations is about 150 miles, over the Atlantic it is at least three times this figure. It is therefore quite possible for some meso-scale or even synoptic-scale feature with dimensions 50–100 miles or more to pass unnoticed through the mesh or to affect one station's observations to a greater or lesser extent.

(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, about air masses, the presence and activity of fronts, cloud types, and the advection of warm or cold air.

(iii) Involves estimating what would be the result 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 the province of the forecaster, rather than the analyst, and is something which can only be decided from wider considerations on a synoptic scale.

3.2. UPPER AIR OBSERVATIONS

3.2.1. *Methods used to obtain upper air information*

The most common means of "sounding" the atmosphere is by a radio-sonde. These instruments which measure temperature, pressure and humidity are released twice a day as a regular routine from special stations equipped with radar and radio receivers on land, and also from the Ocean Weather Ships. By attaching a radar reflector to the sonde, it can be tracked as it rises through the atmosphere and from its changing position the upper winds may be computed.

A similar type of instrument is sometimes used in association with aircraft; but instead of the sonde rising upwards attached to a balloon, it can be dropped with a parachute from a high flying aircraft. This method, the "drop-sonde" is rarely used now.

Winds may also be measured by visual observations of a free balloon with a theodolite. These pilot-balloon observations are not regularly done in the United Kingdom or Europe where the radar wind-finding network is commonly thought to be sufficient, but in other parts of the world they are still a regular and essential part of the daily observational routine. Even in the United Kingdom there are occasions when the detailed structure of the wind in the lowest few thousand feet has to be measured and the rather coarse measurements of the radar balloons are inadequate. Aircraft ascents and captive-balloon soundings (BALTHUM) also provide important supplementary information.

Other less frequently used means of "sounding" the atmosphere include constant-height balloons or "transo-sondes" which drift with the wind at fixed heights.

In the future, soundings of the high atmosphere may well be carried out by rockets, but it is difficult to say to what extent the data so obtained will be of use to the weather forecaster.

3.2.2. *Nature of the observations*

The observations of the upper atmosphere are

Observation	*Units in which observation is reported*
Wind direction	Tens of degrees
Wind speed	Knots
Pressure	Millibars
Temperature	Degrees Celsius
Dew-point	Degrees Celsius

As stated in Section 3.2.1. wind speed and direction are normally measured by radar. Pressure, temperature and humidity are measured by radio-sonde. The standard times of the observations are 0000 G.M.T. and 1200 G.M.T.

The accuracy of the observations is discussed in those chapters of this handbook which relate to the measurements concerned, namely Chapter 13, Wind, Chapter 14, Temperature, Chapter 15, Humidity, and in the Handbook of meteorological instruments, Part II.*

In order to evaluate the significance of the features of upper air soundings, and particularly of rapid changes, it is important to have some background knowledge of the known distribution of wind, temperature and moisture in the atmosphere and the gradients which occur. This has been covered in later chapters (for example Chapters 13-15 inclusive).

3.3. UPPER WIND SOUNDINGS

The most usual way of representing the upper winds over an area is simply to tabulate them on Form 2462 (reports of direction and speed of upper winds), usually with all the winds for one time on the same sheet. From such a tabulation the backing and veering of the wind with height (and its variations in speed) are easily seen, but thermal winds between any levels are not always easy to visualize or analyse.

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*London, Meteorological Office; Handbook of meteorological instruments, Part II, Instruments for upper air observations. London, H.M.S.O., 1961.

Table I illustrates wind data as shown on Form 2462.

TABLE 1 *Upper winds at Crawley, 0000 G.M.T., 21 January 1960*

Pressure level	*Direction*	*Speed*
millibars	*degrees*	*knots*
Surface	120	10
800	140	33
750	180	30
700	250	46
600	270	54
500	290	59
400	290	60
300	290	65
250	290	79
200	300	89

The analysis of winds at single stations by means of plotting the winds on a time cross-section is a technique which is not widely used in the United Kingdom and is perhaps more appropriate to those parts of the world where the density of upper air reports is very low and where significance is attached to rather small and temporary perturbations in a normally regular wind régime.

3.4. THE HODOGRAPH

3.4.1. *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 3.1 the wind at 900 mb. is 240° 20 knots and the wind at 800 mb. is 300° 30 knots, and the

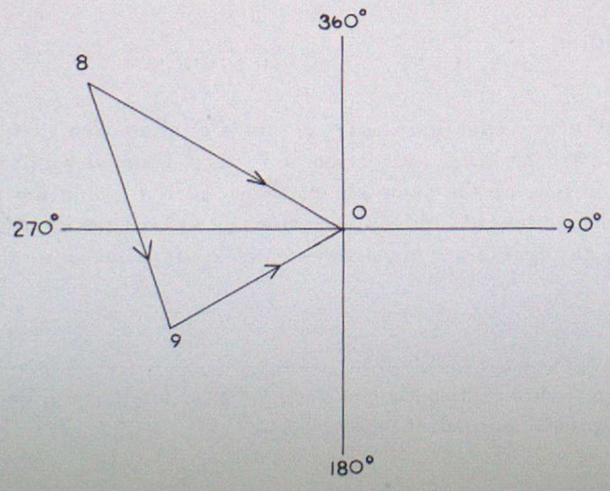

FIGURE 3.1 *Construction of wind vectors*

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. (See Chapter 2, Section 2.4.4. for definition of thermal wind.)

When plotting the complete hodograph of the upper winds at a station it is not always 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 each level. A typical hodograph with the winds on which it is based is shown in Figure 3.2.

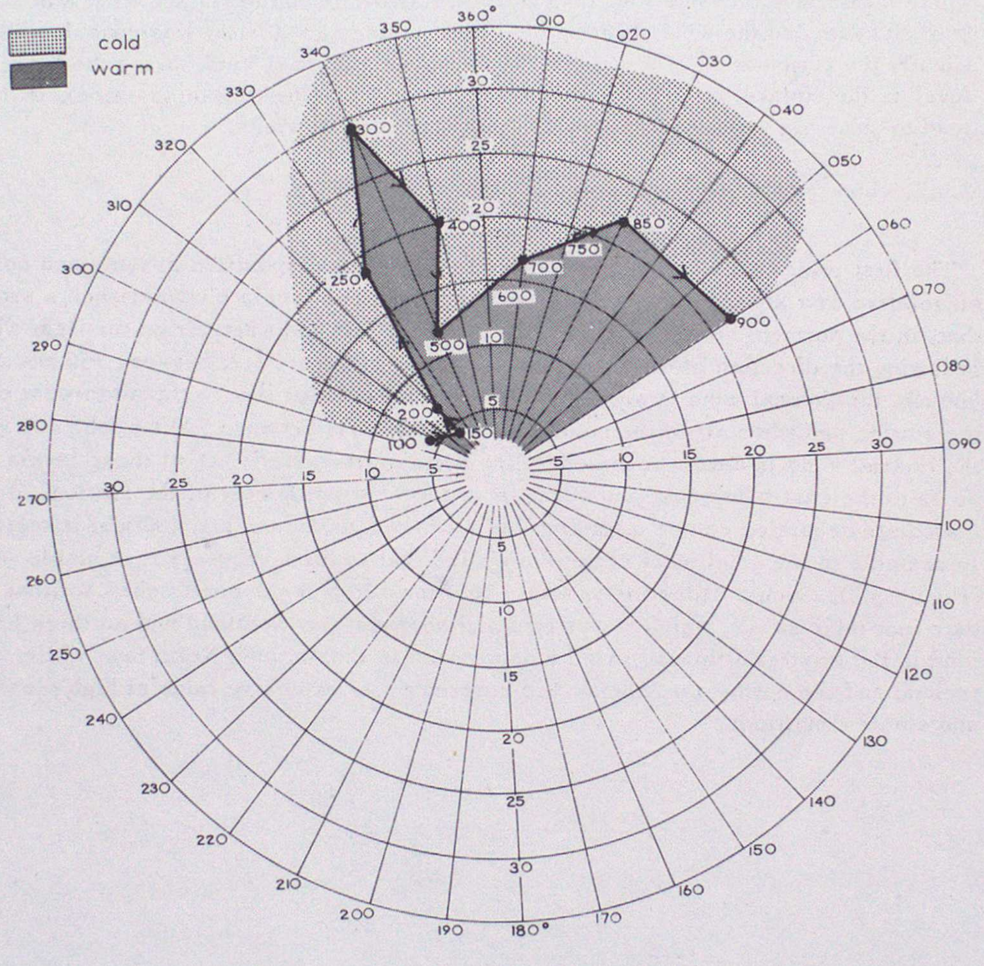

FIGURE 3.2 *Hodograph of Stornoway ascent, 0600 G.M.T., 9 June 1961*

mb.	degrees	kt	mb.	degrees	kt	mb.	degrees	kt
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 each 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 3.2 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.

3.4.2. *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. The thermal wind between two levels blows in such a sense that, in the northern hemisphere, cold air is on the left and warm air on the right when following the direction of the thermal wind. Thus in Figure 3.2, 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 3.3). 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.

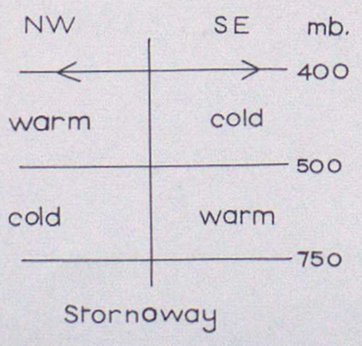

FIGURE 3.3 *Estimation of air mass stability from the hodograph*

The hodograph also shows the direction in which the warmer or colder air is moving at any level. For example in Figure 3.2, 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 3.2, 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 direction of warm air over the station, for the thermal wind direction is about 005°, hence warm air lies to the west and cold air to the east of Stornoway. (Figure 3.4) The actual winds through the layer are from 340°–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 3.2) 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.

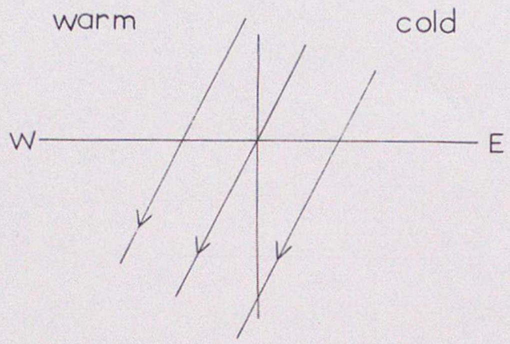

FIGURE 3.4 *Thickness lines 500–400 mb. for Stornoway, 0600 G.M.T., 9 June 1961*

3.4.3. *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 régime 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 later chapters (Chapters 7, 8 and 10) will show that frontal structure is more complicated. Let us imagine a sloping warm front surface $A_o B_o$ (Figure 3.5) 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.

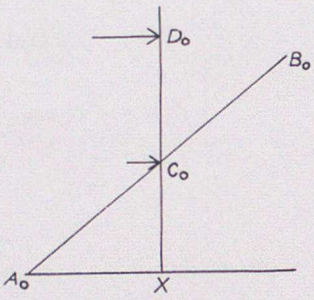

FIGURE 3.5 *Increase with height of the component of the wind normal to a front*

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 *anafronts* and usually give much rain and thick cloud. Fronts at which the warm air is subsiding are known as *katafronts* 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 3.6). 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.

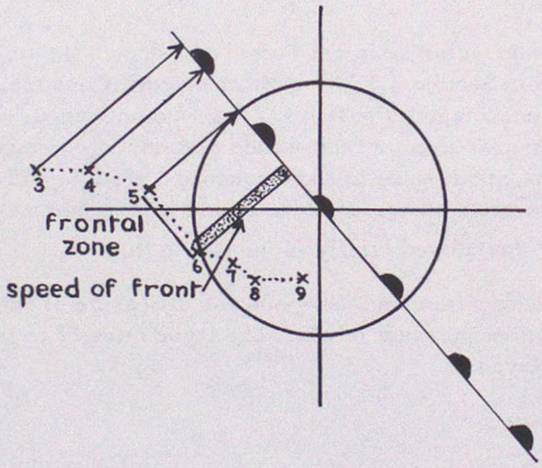

FIGURE 3.6 *Orientation of a warm front shown on a hodograph*

The speed of movement of the front, in a direction normal to its surface line of orientation, is the speed of the cold air first 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 radio-sonde ascents, can be very helpful in analysing fronts approaching the United Kingdom 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.

3.5. THE TEPHIGRAM

3.5.1. *General*

The upper air or thermodynamic diagram is a tool which has been designed to enable the meteorologist to analyse the structure of the upper air in terms of the observed thermodynamic properties pressure, temperature and humidity.

3.5.2. *Temperature and humidity diagrams*

The simplest way is to construct a temperature–height curve. This is not a proper thermodynamic diagram but may nevertheless have some practical convenience at times, for example the fine detail given on a Cardington BALTHUM is possibly more usefully displayed for some purposes on a rapidly constructed temperature–height graph than on any other available printed diagram.

The most commonly used diagram in use in the United Kingdom is the tephigram, which will be fully discussed in Section 3.5. Many other countries use tephigrams but they are not universal. Some diagrams possess $T \log p$ co-ordinates, which are the basis of the Väisälä diagram as used at radio-sonde stations in computing heights of isobaric surfaces. The emagram is used in some countries, also the Stüve diagram. All these diagrams have temperature as one co-ordinate and heights are measured in terms of pressure. They are considered briefly in Section 3.10.

On all these diagrams 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 humidity are clearly displayed.

3.5.3. *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

$$dq = c_v dT + dw = c_v dT + p d\alpha \qquad \ldots \quad (1)$$

and the equation of state of a perfect gas

$$p\alpha = RT \qquad \ldots \quad (2)$$

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 temperature in the Absolute Scale, 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 (1) is divided through by T it follows that

$$\frac{dq}{T} = c_v \frac{dT}{T} + p \frac{d\alpha}{T} \qquad \ldots \quad (3)$$

The term $\frac{dq}{T}$ which appears on the left-hand side of equation (2) is important in relation to the tephigram, and is defined as the change of entropy of the air. The integral $\int \frac{dq}{T}$ round a closed cycle of reversible changes is zero and so $\int \frac{dq}{T}$ 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 (3) becomes

$$c_v \frac{dT}{T} + p\frac{d\alpha}{T} = 0 \qquad \ldots \quad (4)$$

or in another form, derived from differentiating equation (2) and combining it with equation (4),

$$\frac{dq}{T} = c_p \frac{dT}{T} - R\frac{dp}{p} = 0, \qquad \ldots \quad (5)$$

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

Now

$$\int\left(c_p \frac{dT}{T} - \frac{R\,dp}{p}\right) = \int d\left(\log T^{c_p} p^{-R}\right) \qquad \ldots \quad (6)$$

or

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

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

Thus

$$\theta = T\left(\frac{1000}{p}\right)^{R/c_p} \qquad \ldots \quad (7)$$

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 mb. A value of the potential temperature therefore defines a given adiabatic process. The potential temperature must remain constant during an adiabatic process.

If equation (7) is differentiated logarithmically one obtains from equation (5)

$$\frac{dq}{T} = d(c_p \log \theta) \qquad \ldots \quad (8)$$

3.5.3.1. *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 $c_p \log \theta$. Since $c_p \log \theta$ is constant during an adiabatic process the straight lines $c_p \log \theta$ = 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 (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.

In the present form of the tephigram, (see specimen Form 2810A, 1956 edition), the pressure decreases upwards over the diagram and consequently when points representative of an aerological sounding are plotted the points corresponding to greater heights in the atmosphere are plotted higher up the diagram. The dry-adiabats or lines of constant potential temperature slope upwards at an angle of about 45° to the left. Values have been calculated for the thicknesses of standard 100 mb. layers at different temperatures. These values appear on Form 2810 1956 edition as numbers in tens of feet.

The thickness values of the consecutive 100 mb. layers bounded by the standard pressure surfaces 1000, 900, 800 mb. etc. may be calculated from the thickness equation

$$z_2 - z_1 = R\,T_m \log {}^{p_1}/{}_{p_2} \qquad \ldots \qquad (9)$$

where z_1, z_2 are the lower and upper heights of the corresponding pressure surfaces p_1, p_2 and T_m is the mean 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 3.7 the dashed line T_1T_2 represents the actual temperature sounding. T_m is the appropriate isotherm, the value of which, when substituted in equation (9) gives the thickness $z_2 - z_1$ of the layer bounded by p_1 and p_2.

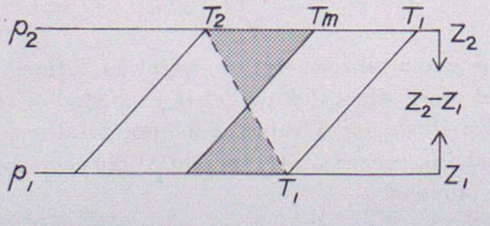

FIGURE 3.7 *Calculation of the thickness of a layer of the atmosphere*

The values of the thickness of layers between consecutive 100 mb. isobaric surfaces are indicated on Form 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 Form 2810A also possesses a nomogram to the left and top of the sheet. This is used 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, in Form 2810A (1956 edition) for a surface temperature of, say, 20°C a line may be drawn connecting 20°C on the upper horizontal temperature scale with 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 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°C. per kilometre, or about 5.5°F. per 1000 feet. Thus, if 9.8°C. or 5.5°F. are measured along a dry-adiabat the isobars intersecting the chosen segment of the dry-adiabat will include a layer of 1 kilometre or 1000 feet as the case may be. This device can be used on any part of the diagram to deduce thickness values.

The figures on the left of Form 2810 represent the height of the particular pressure level in the ICAN standard atmosphere (International Commission for Air Navigation) 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 figures on the left margin of the tephigram are given in feet and below (in brackets) in metres. These provide a useful approximate scale of height.

3.5.3.2. *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 $\int T\,d(\log p)$ is evaluated round a closed curve

$$\oint T\,d(\log \theta) = c_p \oint dT - \oint RT\,dp/p \quad \text{by substitution from equation (7).}$$

The first integral on the right-hand side vanishes round a closed curve and the latter may be evaluated by substitution from equation (2). Thus,

$$\begin{aligned} \oint T\,d(\log \theta) &= -\oint RT\,dp/p \\ &= -\oint a\,dp \\ &= \oint p\,da - \oint R\,dT \\ &= -\oint p\,da \qquad \ldots \quad (10) \end{aligned}$$

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

3.5.3.3. *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. Conversely, it is the rate at which dry air is heated as it descends to levels where the pressure is higher. In both cases it is assumed that the air undergoes an adiabatic process. The values can be derived from equation (5) together with the hydrostatic equation and the equation of state.

Thus $$\Gamma_d = g/c_p = 9.8°\text{C. per kilometre} \qquad \ldots \qquad (11)$$

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. The latter curve is sometimes called the process curve to distinguish it from the actual curve representing the change of temperature with height.

3.5.3.4. *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 temperature which occurs 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 M is the mixing ratio $M=M_v/M_d$ where M_v is the mass of water vapour observed in M_d, the mass of dry air considered. The mixing ratio is usually of the order of a multiple of 10^{-3} but it can be greater provided that the temperature is high enough.

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, $S=M_v/(M_v+M_d)$ where S is the specific humidity. The specific humidity is not used as much as the mixing ratio and, in any case, $M=S$ to a first degree of approximation.

The mixing ratio and specific humidity are a measure 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 parametres 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 m_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 m_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_s. Clearly the relative humidity increases if the temperature decreases, providing the total vapour content remains constant. The relative humidity decreases if the temperature increases providing the total water 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. On Form 2810 (1956 edition) they 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 radio-sonde 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 with 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^*. The virtual temperature T^* is given approximately by the expression

$$T^* = T(1 + 0.61m) \quad \ldots \quad (12)$$

where m is the mixing ratio already defined.

3.5.3.5. *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 saturation-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 saturation-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 curves may also be derived graphically from a tephigram upon which the dry-adiabatic lines and the saturation-mixing ratio isopleths are printed. The latter method simply uses the idea of separating the change in temperature into the two parts discussed above. A small amount of water vapour is assumed to be condensed. The resultant release of latent heat warms the air by a certain amount at constant pressure. The dry-adiabatic curve through the point defined by the new temperature at the original pressure will intersect the new saturated-vapour-pressure isopleths at some other lower pressure. The small segment of the saturated-adiabatic curve is then constructed through the original point, say p_0, T_0, w_0, to the new point defined by the intersection of the dry-adiabat with the lower saturated mixing ratio isopleth, say p, T, w where $w = w_0 - dw$ and dw is the amount of water vapour condensed.

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

The saturated-adiabats may be used to find the equivalent 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-adiabat to 1000 mb.

3.5.3.6. *Normand's Theorem.* Normand 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.

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.

3.5.3.7. *Plotting of 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 3.8 illustrates an upper air ascent on the tephigram. The curve shows a shallow inversion or increase of temperature with height near the surface. This is

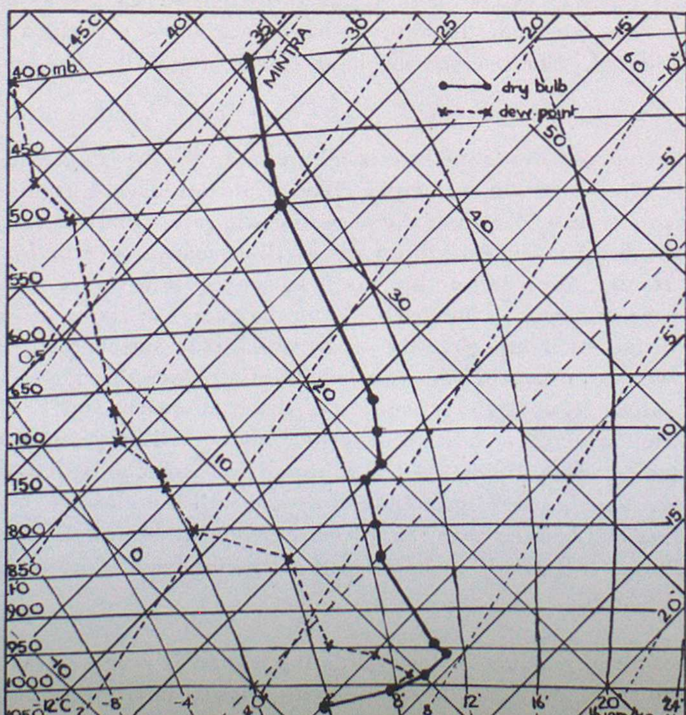

FIGURE 3.8 *An ascent plotted on a tephigram section*

followed by a shallow isothermal layer to about 960 mb. Above this corresponding height 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-adiabat tends to approach the dry-adiabat. 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.

Further examples of environment curves of temperature and dew-point are shown in Figures 3.21–3.27. These soundings representative of different types of synoptic situation are discussed in some detail in Section 3.11.

3.5.4. *Stability and Instability*

3.5.4.1. *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 upward or downward 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 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 the same. Thus the parcel of rising air will be warmer and less dense, or lighter 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 only occur 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 and heavier 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 (Section 3.5.3.2. Figure 3.9).

3.5.4.2. *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 3.9. 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.

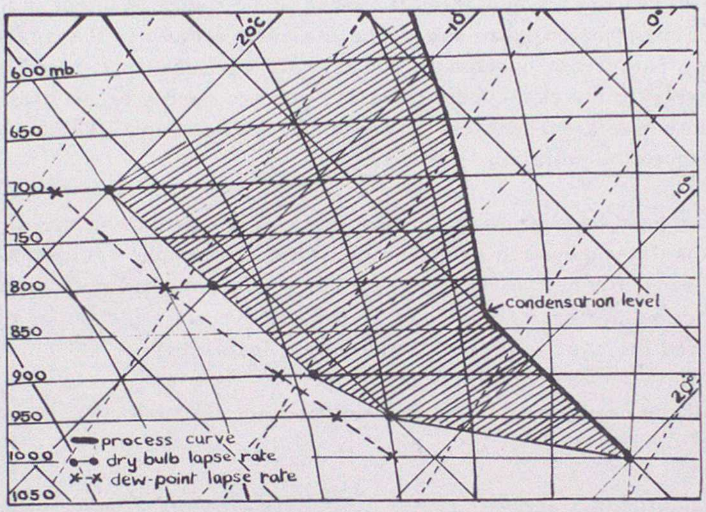

FIGURE 3.9 *Unconditional instability*

3.5.4.3. *Conditional Instability.* 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 3.10. An examination of the environment curve shows that, in the lower

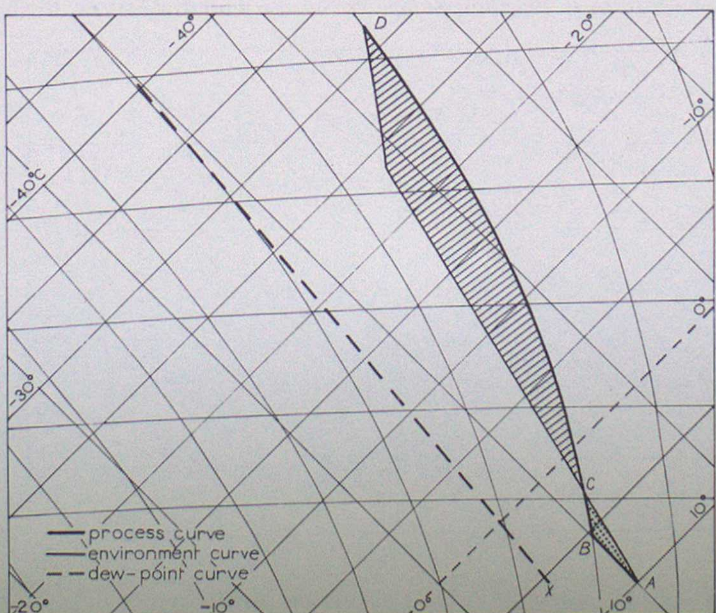

FIGURE 3.10 *Conditional instability*

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*.

3.5.4.4. *Latent Instability.* Latent instability is a term used, mainly in the United Kingdom to define the state of conditional instability as illustrated in Figure 3.10. 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 3.10 represents real latent instability since the area *ABC* is much less than the area *CDE*.

3.5.4.5. *Convective Instability.* Convective instability sometimes called potential instability refers to the realization of stability 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 convective instability is to consider the lifting of the air at the upper and lower boundaries of some chosen layer. For example in Figure 3.11 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 saturation-adiabatic to *B* where the

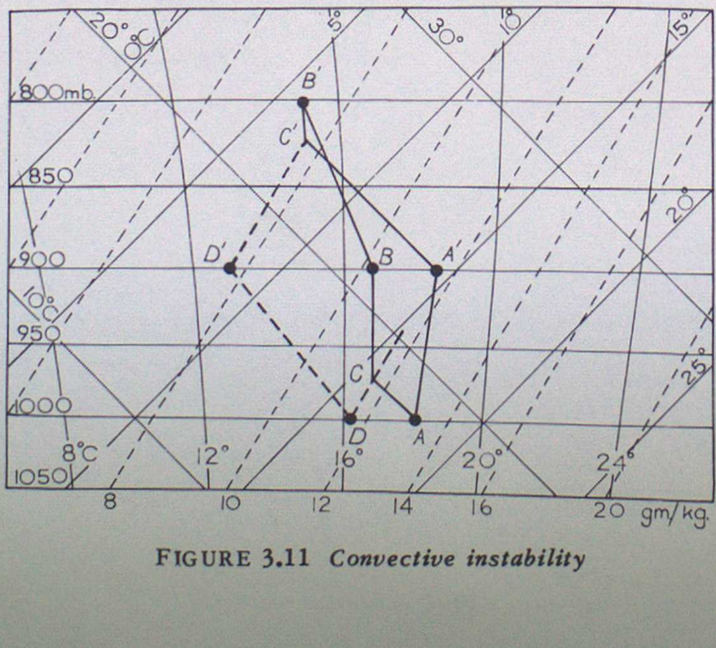

FIGURE 3.11 *Convective instability*

pressure is 900 mb. Similarly air at A' rises until it reaches its condensation level at C', and continues up to the saturation-adiabat 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 saturation-adiabat 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 saturation-adiabat 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 convective 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 convective instability is converted to true instability.

3.6. ENERGY RELEASE IN UNSTABLE ASCENT

3.6.1. *Parcel Method*

It was shown in Section 3.5.3.2. that work could be represented on a tephigram by an area bounded by a closed curve (for example, equation (10)). In Section 3.5.4.1. it was seen that the energy of instability is proportional to the area bounded by the environment curve and the process or adiabatic curve. When the air is unstable the amount of latent energy released may be assessed by measuring the area bounded by the process and environment curves.

A parcel of the environment at rest is subject to a gravitational force g/α' which is balanced by the vertical gradient of pressure in the undisturbed environment. If now a parcel of air is introduced which has ascended from another level and has specific volume α, then it will be subject to the gravitational force g/α, but the same upward pressure force g/α'.

Thus, the net upward force on the parcel is $g(1/\alpha' - 1/\alpha)$ per volume. If the parcel is displaced upwards through a pressure interval $-dp$ it will move through a distance $\alpha' dp/g$ (because of the hydrostatic relation $dp = -gdz/\alpha'$ and the work done on the parcel (that is the energy released) $= g\left(\frac{1}{\alpha} - \frac{1}{\alpha'}\right)\frac{\alpha'}{g}dp = \frac{\alpha' - \alpha}{\alpha}dp$. Thus the work done on unit mass is obtained by multiplying by α, namely $(\alpha' - \alpha)dp$ which is represented on the tephigram by the area between the process curves as shown in Figure 3.10.

3.6.2. *Slice Method*

The methods of stability analysis which have been discussed in the previous sections have been restricted to the so-called parcel method. This assumes that a parcel or small bubble of air moves through its environment without any change of heat or moisture through the boundaries. Such a condition is probably unreal in nature and only represents a rough approximation to the truth. A second method of analysis, called the slice method, attempts to take into account the warming of the environment by descent. Every rising parcel of air must be compensated by a descending parcel. The upward motion is assumed to be saturated-adiabatic while the downward motion is assumed to be dry-adiabatic.

The method is illustrated in Figure 3.12.

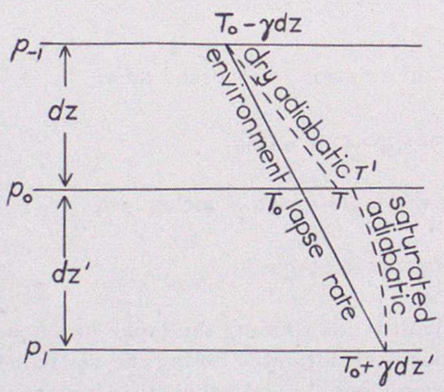

FIGURE 3.12 *The slice method*

A parcel is supposed to originate from some basic level p_1 and rise through a vertical distance dz' to another level p_0. At the same time the environment is assumed to descend to level p_0 from some higher level p_{-1}, located at a vertical distance dz above p_0.

We have the temperature T' of the saturated ascending air at some intermediate level p_0.

$$T' = T_0 + \gamma dz' - \Gamma_s dz' \qquad \ldots \quad (13)$$

where T_0 is the temperature of the environment at p_0. γ is the actual lapse rate, and Γ_s is the saturated-adiabatic lapse rate. The temperature T of the descending air at level p_0 is

$$T = T_0 - \gamma dz + \Gamma_d dz$$

where Γ_d is the dry-adiabatic lapse rate. At level p_0 the difference between the rising air and its environment is $T' - T$.

This difference would be $T' - T_0$ if the parcel method were used. The difference between the two methods is $T - T_0$. The level at p_0 is $T - T_0$ degrees warmer if the slice method is used.

Clearly, since the slice method takes into account the warming of the environment by dry-adiabatic descent it reduces the buoyancy forces as compared with the parcel method. Thus, convective phenomena may be overemphasized if the parcel method is used without any modification.

The warming of the environment depends on the size of dz compared with dz'. If the environment descends only slightly as when a few parcels or narrow towers ascend the result differs little from the parcel method. The saturated-adiabatic lapse rate which is the borderline for instability when analysed by the parcel method is also the borderline for narrow cloud towers when analysed by the slice method. As the lapse rate exceeds the saturated-adiabatic convection in broader towers (or larger parcels) becomes possible.

3.6.3. *Entrainment*

In both the slice and parcel methods it is assumed that the rising air retains its properties and remains insulated from the environment. Such processes may be idealized and a more realistic appreciation may take into account the fact that air from the environment is entrained into the rising columns of air. Such entrainment would clearly influence both the temperature and the humidity content of the rising bubbles or parcels concerned.

A technique has been evolved to take entrainment into account. The procedure consists of mixing environment and rising air at constant pressure and then saturating the mixture by a wet-bulb process. The technique is illustrated in Figure 3.13.

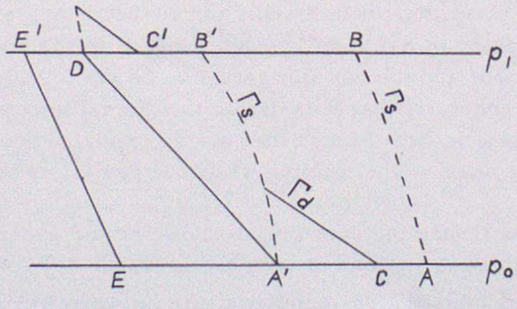

FIGURE 3.13 *The entrainment process*

AB is a saturated-adiabatic curve. Let point *A* denote the rising and cloudy air at some level p_0 at which the mixing with environment *E* by entrainment occurs. Point *C* represents the mixing before saturation and point *A′* the mixture after saturation has occurred as a result of the evaporation of part of the liquid water content of the cloud. Now if no further mixing by entrainment occurred the air at *A′* would ascend along the saturated-adiabatic curve *A′B′*. However, if further mixing with environment *E′* by entrainment occurs at a higher level p_1 the situation of level p_0 is repeated by points *C′* and *D*, respectively. Thus the curve *A′D* represents the actual curve of the rising air column instead of the saturated curve *AB* (no entrainment at all) or *A'B'* (entrainment at level p_0 only). The process shown at Figure 3.13 could be applied to an entire sounding ascent.

It is seen that the entrainment process results in a lapse rate for the rising bubble which is greater than the saturated-adiabatic lapse rate. It gives a smaller area of available latent energy between the parcel and environment curves. Consequently buoyant forces are reduced and convective activity is diminished in comparison with the ordinary processes which neglect entrainment consideration.

The slice method may be used to incorporate entrainment. It is difficult to evaluate the different methods except by statistical means. A test of various techniques of stability evaluation in the United States of America indicated that the best results were obtained by using the parcel method with entrainment.

3.6.4. *Estimation of convection during diurnal heating*

On many occasions the early morning ascent shows a stable lapse rate or an inversion in the lowest layers. At higher levels the lapse rate may be such as to give a condition of conditional instability which can be released if and when the stable lapse rate in the lower layer has been transformed into a dry- or saturated-adiabatic lapse rate according to whether the layer or part of a layer is unsaturated or saturated when it ascends. It is therefore of practical importance to devise a method for predicting the maximum lapse rate in the surface layer which is likely to be created during the course of a day. The method essentially involves the prediction of the maximum temperature for the day. The most widely used technique for predicting the maximum temperature is the Gold square method described in Section 14.7.1. Once the maximum temperature has been found a line can be drawn along the dry-adiabatic lapse rate passing through that temperature. When the parcel of air which is thus supposed to ascend becomes saturated (this will happen at the point where the dry-adiabatic intersects the mixing ratio line defined by the dew-point of the air corresponding to the maximum temperature already found), the line of ascent continues up the saturated-adiabatic curve which intersects the aforementioned point. A line is drawn along the latter curve until it intersects the environment curve. Then as previously discussed the energy released will be proportional to the area formed by the process curve on the right and the entrainment curve on the left of the tephigram. Various stages in the convection process during the diurnal heating may be estimated by drawing the dry-adiabatic line through the various ground temperatures which may be reached at different times of the morning while the heating is progressing.

In Figure 3.14 the early morning sounding is shown by the ascent marked *A B C D*. It is noted that there is an inversion between levels *A* and *B*. Above *B* there is a lapse rate which approaches the dry-adiabatic rate between 915 and 800 mb. It is assumed that the case refers to an early morning ascent in late spring or summer and that the conditions are such that the Gold square method predicts a maximum temperature of 22°C. The dew-point at *A* is 10°C. showing saturation at the time. The surface air may be assumed to retain the same amount of moisture, actually the surface layer will probably gain moisture through evaporation from the surface which may be dew covered or from low-lying fog or from vegetation generally. In the latter case the dew-point corresponding to *A* may be at some temperature between *A* and *A'*. However, if the moisture content is supposed to remain the same a dry-adiabatic line may connect points *A'* and *C*. The latter point is the level at which the surface air of maximum temperature 22°C. will become saturated. Beyond *C* the curve *CD'* follows the saturated-adiabatic curve. The shaded area represents the energy released by unit mass of air which rises from A'. If the dew-point is assumed to be at A'', then the dry-adiabatic curve from A' will go up to C' from whence the saturated-adiabatic will go up to D''. The surface air becomes saturated at C' in the latter case.

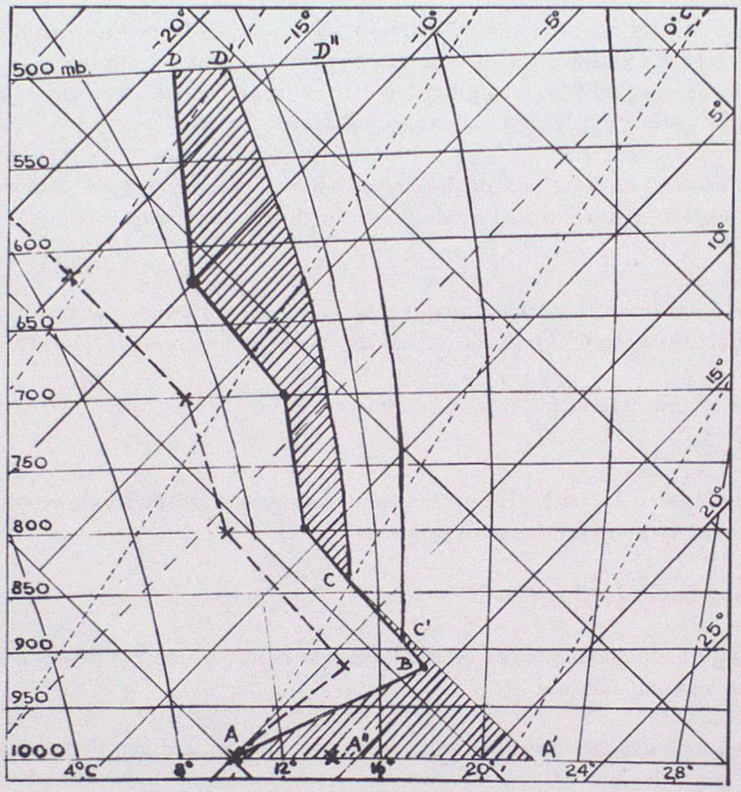

FIGURE 3.14 ***Estimating convection activity during diurnal heating***

3.7. INSTABILITY INDICES

Stability and lifting indices have been introduced as aids in the analysis of stability. The stability index is a measure of latent instability computed by lifting a parcel of air adiabatically from 850 to 500 mb. The theoretical temperature of the lifted parcel is then subtracted algebraically from the environment temperature at 500 mb. Positive numbers indicate stability and negative numbers instability. The lifting index is computed by similar means using a forecast maximum temperature. Such indices may be plotted on a synoptic chart and isopleths of equal index drawn. The resultant patterns may be a useful addition to chart analysis.

A further method is to calculate the actual area traced out on the tephigram when parcels of air at specified pressure levels, say 1000 and 700 mb. are lifted to other specified pressure levels, say 200 mb. The stability of a layer of air may be found in this way by averaging the areas found. The areas may be measured by a planimeter.

3.8. CONDENSATION LEVELS

In the preceding sections it has been frequently shown how rising air will become saturated during ascent when it reaches its dew-point. When the air becomes saturated condensation occurs.

Condensation may occur in the atmosphere as a result of several different processes. If, for example, air is lifted mechanically over a mountain barrier it will expand and cool adiabatically until it becomes saturated. The level at which this occurs is called the lifting condensation level.

It occurs at the intersections of the dry-adiabat drawn through the temperature, and the saturation mixing-ratio line drawn through the dew-point of the sample of air.

If, on the other hand, air rises because surface heating causes instability, the level at which saturation occurs is called the convection condensation level.

Examples of the estimation of the condensation level have been illustrated in Figures 3.9, 3.10, 3.11, 3.13.

Air may be cooled isobarically until it becomes saturated at its dew-point. No change in level is involved in such a process.

3.8.1. *Mixing condensation level*

The mixing condensation level is the lowest level at which condensation can occur due to vertical mixing.

It is found as follows: find the temperature of the air column after complete mixing and mark the corresponding dry-adiabatic line through the temperature. Then find the dew-point of the air column after complete mixing and draw the corresponding saturation mixing ratio line through that dew point. The intersection of these two lines indicates the mixing condensation level.

It is assumed that true mixing occurs when two parcels of air are brought adiabatically to the same pressure. Horizontal mixing can occur at the same pressure but it is vertical mixing after an adiabatic process which determines the temperature which defines the mixing condensation level. In Figure 3.15 it is desired to find the mixing condensation level of the layer 1000–900 mb. The temperature sounding in the layer is given by the curve T_0T_1 and the dew-point curve by W_0W_1. An adiabatic line is then drawn through T_0T_1 in such a way that the area $A = A'$. This adiabatic line θ_m corresponds to the mean potential temperature of the layer θ_m. Now a saturation mixing ratio line W_m is drawn through the dew-point curve W_0W_1 such that the area B is equal to B'. W_0 and W_1 are the saturation mixing ratios corresponding to the dew-points at 1000 mb. and 900 mb., respectively. The line W_m is a mean saturation mixing ratio for the layer after vertical mixing. The upward projection of θ_m and W_m meet at a point which is the mixing condensation level for the layer in question after vertical mixing.

In Figure 3.15 the mixing condensation level is higher than 900 mb. so that vertical mixing will not alone cause condensation or cause a low cloud base to form within the 1000–900 mb. layer. If, however, the intersection θ_m and W_m occurs

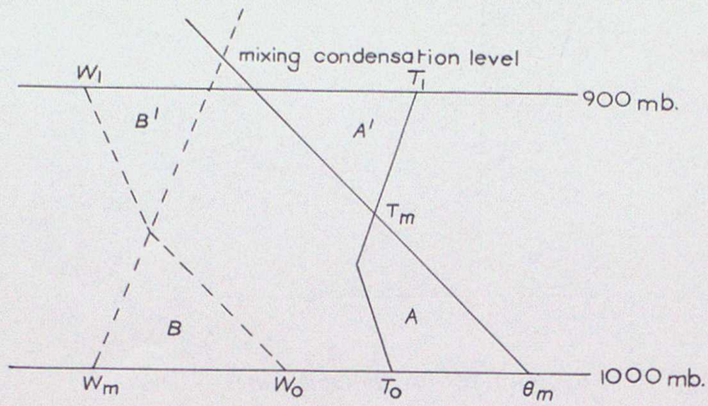

FIGURE 3.15 *Determination of the mixing condensation level*

within the 1000–900 mb. layer then condensation and low cloud will occur at the level of intersection. Above this level the lapse rate established by mixing follows the saturated-adiabatic instead of the dry-adiabatic curve.

3.8.2. *Convective Condensation Level*

The process of vertical mixing is most effective during diurnal heating. The method described in Section 3.8.1. above may be applied to the case of estimating convective activity during diurnal heating. A mean saturated mixing ratio may be determined by the equal area method, although slightly greater weight can be given to the lowest layer to take into account an increase of moisture content as a result of evaporation from the earth's surface. The point at which the dry-adiabat from the point of maximum temperature at the surface pressure intersects the mean saturated-mixing-ratio line is called the convective condensation level.

When the maximum temperature is being estimated from the "Gold square method" it may be more useful to apply the above method for successively deeper layers assuming a certain rate of increase of the surface temperature. When, ultimately the dry-adiabat through the surface temperature intersects the mean saturated-mixing-ratio corresponding to that surface temperature and the intersection lies to the right of the environment curve, then the point of intersection will give the estimated convective condensation level and at the same time the lowest surface temperature necessary to produce cumulus type clouds. It is preferable to construct each stage of the analysis on a separate tephigram to avoid confusion of overlapping areas and intersecting lines.

3.8.3. *Precipitable Water*

The saturation-mixing-ratio curve can be used to calculate the amount of precipitable water in a layer of air. Thus in Figure 3.16 the dew-point curve may be divided into two equal areas bounded by the pressure levels p_1 and p_2. The dividing line represents the mean mixing ratio for the layer W_m. The amount of water M is given by $M = \frac{Wm}{g}(p_1 - p_2)$ gm./cm^2. where W_m is in gm./kg. and p is in mb. The above quantity may be evaluated once W_m has been determined.

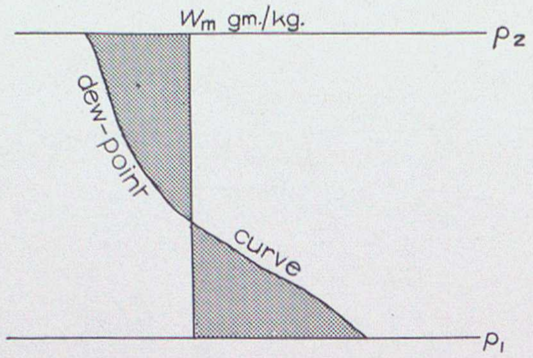

FIGURE 3.16 *Precipitable water*

The rate of precipitation may also be determined from the above method. It is assumed that the layer is saturated and undergoing vertical ascent along the saturated-adiabatic lapse rate at some estimated vertical velocity. The amount of precipitable water may be determined both prior to and after ascent. The difference in these two quantities represents the amount of water that has been precipitated during the time it takes for the whole layer considered to be lifted from levels p_1, p_2 to $p_1'p_2'$.

3.9. FURTHER PRACTICAL APPLICATIONS OF THE TEPHIGRAM

3.9.1. *Homogeneous air masses*

Tephigram soundings provide basic criteria for the identification of homogeneous air mass types (see Chapter 14 Section 14.8.1. referring to Belasco's classification). The specific identification of an air mass enables certain statistical limits to be assigned to its characteristic properties. These limits are useful in predicting maximum temperature and other values likely to be reached during diurnal heating and nocturnal cooling.

3.9.2. *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 3.17 shows a characteristic lapse rate through a warm front-surface.

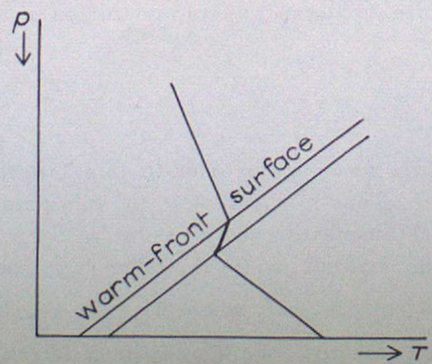

FIGURE 3.17 *Temperature ascent through a warm front*

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 is given in Section 3.11, Figure 3.23.

3.9.3. *Subsidence and ascent*

The temperature curve often shows an inversion when subsidence is occurring. This is illustrated in Figure 3.18. Suppose the layer 500 mb. to 600 mb. is subsiding. The initial lapse rate before subsidence is shown by the line *A B C D*. 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 were 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.

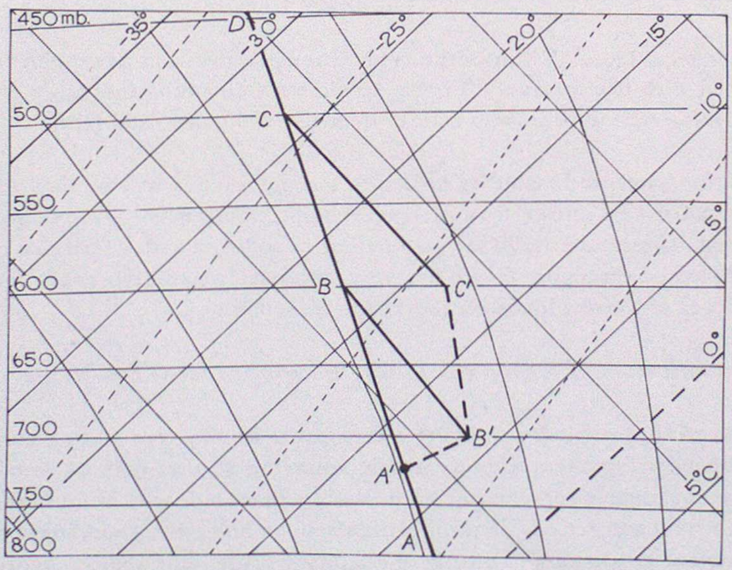

FIGURE 3.18 *Subsidence inversion on the tephigram*

Chapter 14 Figure 21 is a good example of subsidence.

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.

3.9.4. *Fog*

Radio-sonde ascents rarely show the temperature and humidity structure in fog, with much accuracy. As far as the United Kingdom upper air stations are concerned all but two of them are very near the sea where the incidence of radiation fog is low, but even at inland stations the sonde cannot register much detail of conditions in the fog layer because of the usually shallow depth of radiation fog. However, since it is an integral part of the analysis at inland sites to modify, say, a coastal ascent without fog from midnight, so as to represent the inland conditions with fog at about 0600 G.M.T., it is important that the typical temperature structure through fog should be well known by analysts and forecasters. In this connection, the regular Cardington BALTHUM messages, received by all outstations, can be very helpful and it should be widely encouraged that they be plotted and used where possible.

Figure 28 of Chapter 14 is a good example of the low-level portion of an upper air sounding in fog. The large inversion of temperature and the big difference in relative humidity as measured by the dew-point depression are characteristic features. If the analyst knows that fog is present at a given station he can modify the low level portion of the nearest available sounding, which may be fog free, to incorporate the characteristic features noted above.

3.9.5. *Low stratus and stratocumulus*

For the same reasons as mentioned in Section 3.9.4. above, it is very desirable to know the typical temperature and humidity structure through stratus cloud and equally the radio-sonde is often unable to give it.

Chapter 16, Figure 35 shows the typical temperature and dew-point lapse rate associated with low stratus. Figure 36 shows a sounding through a layer of stratocumulus whose base was mainly between 1000 and 2000 feet.

The temperature and humidity structure through clouds at and above safe flying levels is quite well known from aircraft reports. Currently the weekday BISMUTH flights from Aldergrove include two vertical soundings with details of cloud bases and tops passed through. These are very interesting and give some idea of the varied effect of these clouds on the shape of soundings.

3.9.6. *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.

The procedure is:

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

(ii) Connect the surface temperature and dew-point with the curves reported on the radio-sonde in a manner which takes account of the prevailing conditions, that is if fog, strong wind, heating or cooling is present, assume an appropriate lapse rate and hydrolapse in the lowest layers as indicated in the various sections outlined above.

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 present air mass may have been done far to the west over the sea some six–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 sea-breeze will modify the lowest layers and affect fog forecasts.

3.9.7. *Some information to be gained from inspectiom of the tephigram*

Part of the straightforward analysis of a tephigram consists of the direct extraction of certain frequently used values and parameters. One of these is the height of the significant temperature levels.

The 0°C. level is the most obvious and frequently used individual temperature level. It has a bearing on the formation of precipitation, on the nature of precipitation on reaching the ground, and on ice formation on aircraft.

The effect of continuous snow in lowering the 0°C. level is discussed in Chapter 16, Section 16.6.2.1. and the effect of continuous rain in Chapter 16, Section 16.6.2.2. The effect of flow over hills in lowering the 0°C. level is discussed in Chapter 14, Section 14.11.2.

Other temperature levels which are mentioned in Chapter 16 in connection with clouds and precipitation are: –5°C., –12°C., and –40°C. (Sections 16.7.2. and 16.7.3.). The 0°C. wet-bulb temperature level is important in winter, having a bearing of snow formation (Chapter 16, Section 16.7.6.1.) and glazed frost (Chapter 16, Section 16.7.8.).

The tropopause is fully dealt with in Chapter 7, Section 7.4. in relation to the tephigram.

Finally it should be emphasized that a careful study of each individual ascent can assist the construction of jet stream cross-sections, isentropic charts etc. by enabling more accurate interpolations to be made at intervening levels.

3.10. OTHER FORMS OF UPPER AIR DIAGRAM

There are other forms of upper air diagram in addition to the tephigram. As these are seldom used in the Meteorological Office, their structure will only be mentioned in brief.

3.10.1. *The emagram*

One of the co-ordinates of the emagram is T, as in the tephigram. The other co-ordinate is $-R \log p$. In such a system the isobars and isotherms are straight and perpendicular to one another. The dry-adiabats as deduced from equation (7) take the form of logarithmic curves. They become steeper with decreasing temperature but remain fairly straight within the range of values normally encountered in meteorology. The saturation-adiabats possess pronounced curvature, while the saturation mixing ratios are only gently curved. The angle between the isotherms and dry-adiabats is about 45°.

Summarizing the emagram has:

(i) area proportional to energy,

(ii) four sets of straight or nearly straight lines and one set which is curved,

(iii) a satisfactory angle of intersection between adiabats and isotherms. Figure 3.19 illustrates the emagram.

3.10.2. *The skew T – log p*

This is a modified form of the emagram constructed so that the isotherms are approximately perpendicular to the dry-adiabats as in the tephigram. However, in this diagram the dry-adiabats are gently curved lines from the lower right to the upper left of the diagram, running concave upwards. The saturation-adiabats are markedly curved as before, but the saturation-mixing-ratio lines are essentially straight.

Summarizing, this diagram possesses:

(i) area proportional to energy,

(ii) three sets of straight or nearly straight lines, and one of gently curved and one set more markedly curved,

(iii) an angle of intersection of nearly 90° between the dry-adiabats and the isotherms.

Figure 3.20 illustrates the skew $T - \log p$ diagram.

3.10.3. *The Stüve diagram*

This diagram has an ordinate representing p^{R/c_p} with p decreasing upwards. The abscissa represents T increasing to the right. The dry-adiabats are straight lines converging at $p = 1000$, $T = 0$ at the upper left-hand corner of the diagram. The saturation-adiabats are curved while the saturation-mixing-ratio lines are straight.

Summarizing, the diagram:

(i) has four sets of lines which are straight and one set which is curved,

(ii) has an adiabat–isotherm angle of 45°,

(iii) does not have area proportional to energy.

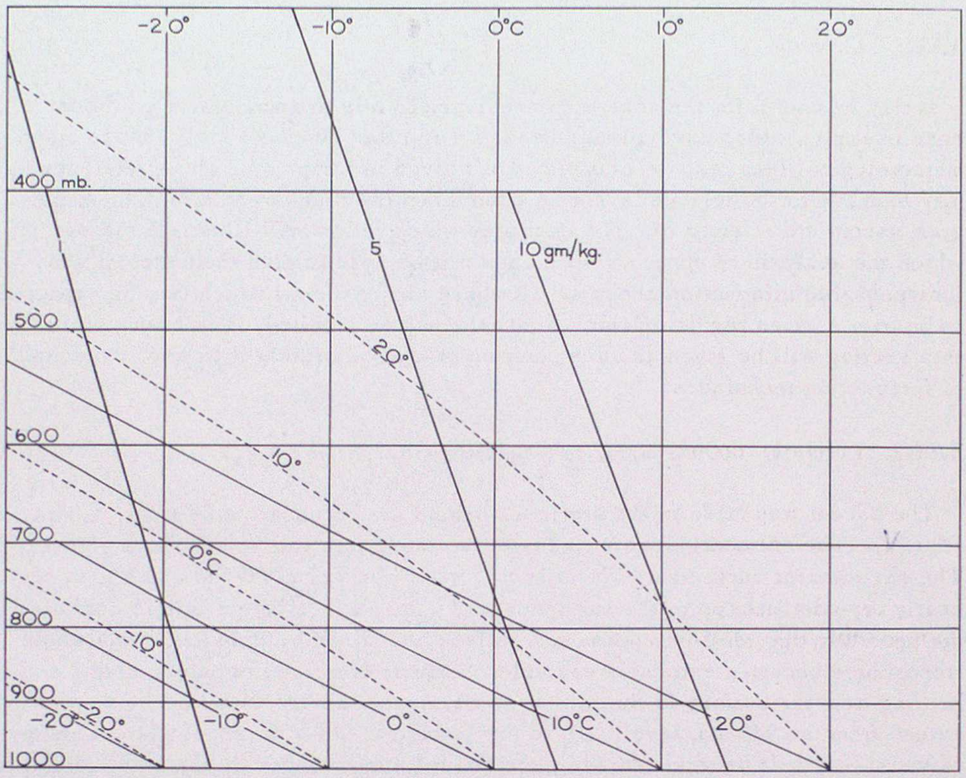

FIGURE 3.19 *Emagram*

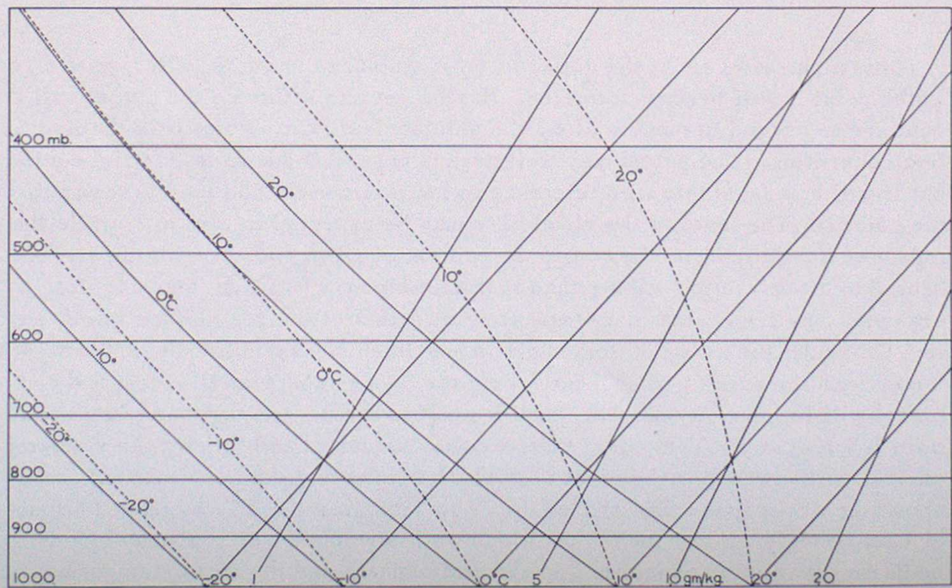

FIGURE 3.20 *Skew T – log p diagram*

3.11 CASE EXAMPLES OF TEPHIGRAM ANALYSES

3.11.1. *General*

It may be useful for the sake of convenience to rely to some extent on those case examples which have already been given in this handbook to illustrate other meteorological features, for example, those given in Chapter I. Thus, reference may be made to the relevant synoptic chart when the figures containing the tephigram ascent are referred to. The examples which follow will illustrate the way in which the analysis of upper air soundings can be done to give the meteorologist indispensable information about the air above him or the air which may be expected to be over a given region in which forecasts may be required. The examples in this section will be given as an exposition of analysis rather than as illustrations of forecasting techniques.

3.11.2. *Lerwick; 0000 G.M.T., 19 January 1958. (Figure 3.21)*

The ascent was made in the arctic air behind the secondary cold front, (Chapter I, Plate V). The extraordinary steepness of the lapse rate can be seen at a glance. The environment curve apart from a short interval between 760 and 680 mb. is very nearly dry-adiabatic up to the tropopause. As the surface layers of this cold air are heated by the relatively warm sea surface and mixed by turbulence, the whole troposphere becomes extremely unstable. Suppose combined turbulent mixing and heating from the surface in the lowest 40 mb. cause the dry-adiabatic lapse rate to extend from the 940 mb. level, (*B*), to the surface at 980 mb. (*A*), the latent energy of instability will be represented by the shaded area bounded by the lapse rate curve and the process curve. The latter will be the dry-adiabatic line to the condensation level C and the saturated curve thereafter. The level C is obtained by using the saturation mixing ratio line drawn through the dew-point (A') of the surface air (see also para. 3.8).

Thus, a parcel of air at the surface (*A*), if displaced upwards, will rise to *C*, at which point it will become saturated. Having become saturated the parcel will continue to ascend to point *D* along the saturated-adiabat. Point *D* is the limit of free convection. The parcel may well ascend beyond *D* due to its inertial motion but it will now be stable with reference to its environment and tend to return to the point *D*. The level of the cloud base may be expected to form at *C* while the tops may rise beyond *D*. Now suppose parcels of air in contact with the sea are heated to a temperature warmer than that existing at *A* (−3°C.); suppose, for example, they are heated to a temperature near that of the sea surface itself, say to 5°C. Under the situation this might indeed happen. We follow the 980 mb. isobar from the point *A* until it intersects the 5°C. isotherm at *E*. From *E* the parcel will follow a dry-adiabat, until it reaches the 2.5 gm./kg. saturation-mixing-ratio isopleth at *F*. The parcel then becomes saturated and follows the saturated-adiabat until it reaches the point *G* at which it intersects the environment curve above the tropopause. The shaded area now indicates a vastly increased amount of latent energy. Under these conditions well developed cumulonimbus clouds could be expected to penetrate into the stratosphere and the weather might be marked by heavy snow showers.

The conditions postulated have, however, been somewhat idealized. In particular it has been assumed that parcels of air in the surface layer possess the saturation mixing ratio indicated by point *A'*. This may only be true for a very thin layer. It is noted that the saturation mixing ratio at the next point on the tephigram (940 mb. at point *H*) is considerably less, only 1.5 gm./kg. Thus in treating the whole 40 mb.

layer it might be better to assume that the layer possessed a mean saturation mixing ratio of 2.0 gm./kg. In this case the areas of latent instability are somewhat less than previously. A parcel at point *E* will not become saturated until it reaches *F′* from whence it will follow a saturated-adiabat to *G′*.

The difference in the latent energy which arises as a result of the determination of the dew-point by these two methods is shown by the cross hatched area. The new area is still sufficiently large for active instability phenomena to develop.

The case illustrates an unusual example for the British Isles because of the extreme coldness of the air mass.

3.11.3. *Camborne; 0000 G.M.T., 21 January 1960. (Figure 3.22)*

This upper air ascent was done in the warm air shortly after the passage of the warm front (Chapter I Plate II). The lapse rate is stable for both dry and saturated air from 970 mb. to 600 mb. The surface layer from 970 to 1000 mb. is, however, unstable for saturated air. The latter condition has probably been brought about by turbulent mixing in the frictional layer. The humidity observations between 1000 mb. and 650 mb. show that the dew-points are about one degree Celsius below the air temperature. Only a small amount of upward vertical motion would saturate the air column up to 650 mb. Consideration of the turbulent layer will help to determine the low cloud conditions. Examination of the inset part of figure 3.22 shows that the lowest 30 mb. layer has a mean saturation mixing ratio of about 7.3 gm./kg. and that total turbulent mixing of this layer will produce saturation at about 970 mb. i.e. at about 800 to 900 ft. It is clear, however, that complete mixing through this layer has not taken place and the air near the surface has maintained a higher mixing ratio than the air only a little higher. The condensation level is thus likely to be rather lower with the base of the turbulence cloud somewhere between 970 mb. and point B, which is where a parcel of air rising without mixing from the surface would become saturated. The slightly lower cloud bases reported in the vicinity are probably due to slightly moister air upwind. (Note the temperatures in the sounding have been plotted in whole degrees, with fractions of a degree ignored, so that detailed consideration of the shallow turbulent layer cannot be wholly successful.)

Since the lapse rate is stable the clouds would be of layer form and if any precipitation fell one might expect it to be light and steady in character.

Figure 3.23 shows the ascent for the same time at Crawley. The surface warm front is still 150 kilometres away so that the balloon would be expected to pierce the upper frontal surface at about 1 to 2 kilometres above the surface. The actual ascent shows the base of a strong temperature inversion at 850 mb. The inversion extends to 790 mb. above which there is an isothermal layer to 750 mb. Snow was falling in south-east England at the time which is in accord with the reported temperatures, the saturated state of the atmosphere up to 750 mb., and the fairly moist layer between 750 and 500 mb. The lapse rate is stable up to about 620 mb. above which it approaches the saturated-adiabatic rate.

3.11.4. *Camborne; 0000 G.M.T., 11 September 1957. (Figure 3.24)*

This example illustrates the case of an ascent undertaken in the warm sector of a young and developing wave depression. The sounding is absolutely stable up to 770 mb. Above this level it follows the saturated-adiabatic curve quite closely. At the surface the air is wholly saturated and it remains very moist up to 550 mb., above which it becomes quite dry. Rain or drizzle is reported throughout the area, indicating that the air is undergoing upward vertical motion.

~~above which it becomes quite dry. Rain or drizzle is reported throughout the area, indicating that the air is undergoing upward vertical motion.~~

3.11.5. *Valentia; 1200 G.M.T., 6 May 1959. (Figure 3.25)*

The region in which the ascent was made lay in a ridge of high pressure. The layer 1020–890 mb. has been heated sufficiently to create what appears to be a super dry-adiabatic lapse rate. Between 890 mb. and 810 mb. there are two ~~exists an~~ inversions. Above 810 mb. the lapse rate is absolutely stable. Let one consider a parcel of air at *A* (1020 mb.) having a saturated mixing ratio of 5.2 gm./kg.; this value is obtained by averaging the water vapour content within the thin layer near the surface. If the parcel is given an initial impulse it will rise along the dry-adiabatic curve to point *B*, the lifting condensation level, at which it becomes saturated. It will consequently rise along the saturated adiabatic curve *BD* to point *D* the top of free convection. Cloud should form with its base at about 930 mb. The reported base of well developed cumulus at 2000 feet agrees well with the calculated level. Above the level of free convection the air is very dry, so that any cloud tops penetrating above this level would be eroded by evaporation. The hatched area of latent instability ACDBA ~~BCD~~ is too small for instability showers to occur. For this to happen the temperature would have to be much higher. Since it is already midday and Valentia is near the coast and probably subject to sea-breeze effects it is unlikely that the temperature will rise much above the reported 11°C. It must be remembered that the surface temperature can only increase to a value which creates a dry- or saturated-adiabatic lapse rate; heat is then carried upwards from the surface to higher levels.

3.11.6. *Crawley; 0000 G.M.T., 29 June 1957. (Figure 3.26)*

The night ascent preceded a day during which maximum temperatures in south-east England exceeded 30°C. A noteworthy feature of the ascent is the pronounced surface inversion which extends from the surface to 970 mb. The temperature increases throughout this shallow layer from 18°C. at the base to 24°C. at the top. The humidity sounding is somewhat erratic and shows a humidity lapse from 970 to 790 mb. and then an increase to 710 mb., at which level the air is very moist. The temperature lapse rate is unstable for saturated air above the top of the hydrolapse inversion.

The Gold square method (see Chapter 14) can be applied to give an estimate of maximum temperature. How great will this have to be in order to release latent instability? We may assume that the average saturation mixing ratio in the inversion layer is 9.5 gm./kg. Then if a dry-adiabat is drawn tangent to *B* it will intersect the required saturation mixing ratio value at *C* and the surface pressure at *A*. The latter point is at 34°C. Thus if the surface temperature reaches this value, the surface air will rise to *C*, become saturated, and continue to rise along the saturated-adiabat to point *D*. The large shaded area represents the available latent energy of the surface layer. The maximum temperature at London Airport on this day was 35°C. This value was reached comparatively early in the afternoon after which the sky rapidly became cloudy. Thunderstorms broke out fairly generally over south-eastern England in the afternoon and evening.

3.11.7. *Crawley; 0000 G.M.T., 1 July 1957. (Figure 3.27)*

This example relates to the situation 48 hours after the case illustrated in Figure 3.27. The hot spell had given way to a thundery depression. The most noticeable difference is the great increase in the moisture content, particularly below 700 mb.

There is an isothermal layer from 1000 mb to 840 mb. but above the latter level the lapse rate is definitely unstable for saturated air.

ratio of about 11.5 gm./kg. in the surface layer, a maximum temperature of 28°C. would set off a huge area of latent instability. This is shown in Figure 3.26 by the shaded area bounded by the environment curve *A* *B D* and the process curve. The latter is dry-adiabatic from *A′* to *B* and the saturated-adiabatic from *B* to *D*. In Figure 3.27 the saturated-adiabatic curve follows the 20°C. wet-bulb potential temperature. Figure 3.27 also illustrates an example of convective instability. If, for example, the layer between 900 and 800 mb. is lifted by 100 mb. the lapse rate within the layer will become convectively unstable. Thunderstorms were in effect reported in the vicinity at the time of the ascent. In this case the instability was triggered by vertical motion resulting from convergence in a shallow depression rather than by surface heating. Although the humidity observations do not indicate saturation at any level, cloud was reported at various levels. It can be seen from the ascent that if cloud and therefore saturation existed at *C* or above, instability would arise. The area of latent instability is bounded by the process and environment curves connecting *B* and *D*. Figures 3.26 and 3.27 both show areas of real latent instability.

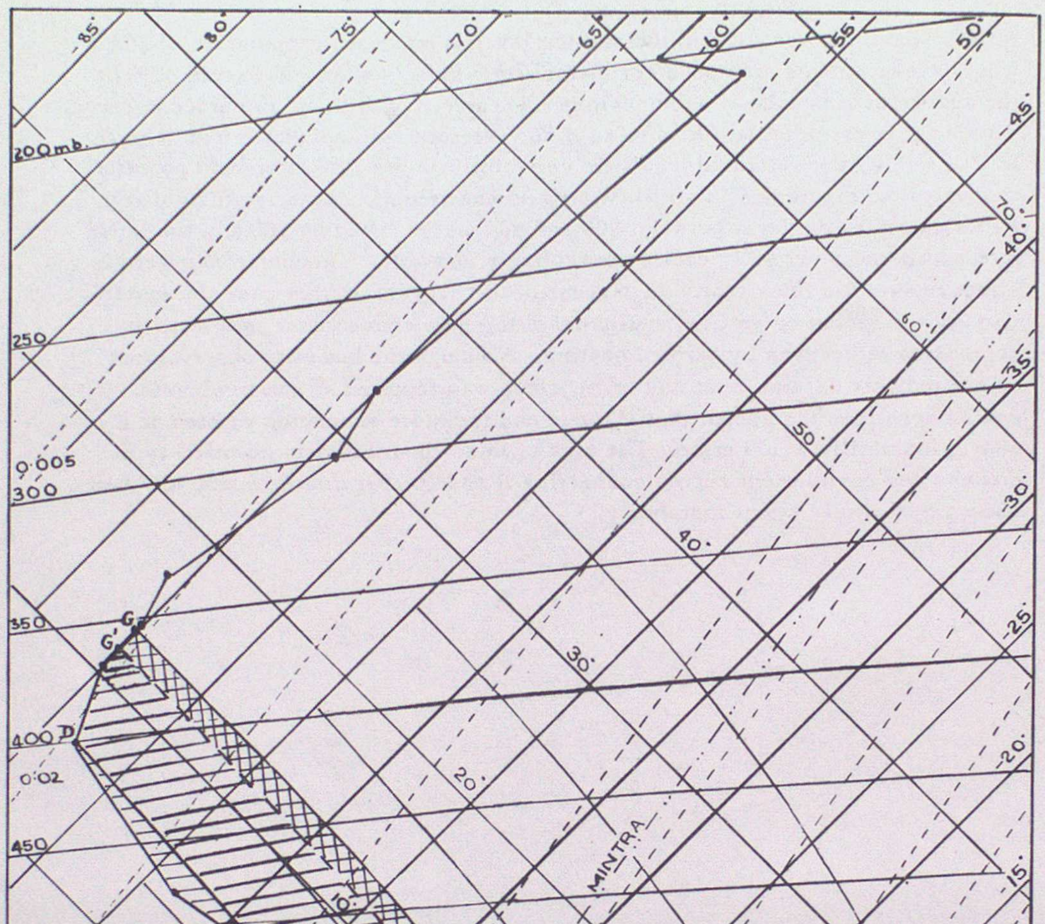

FIGURE 3.21 *Tephigram for Lerwick, 0000 G.M.T., 19 January 1958*

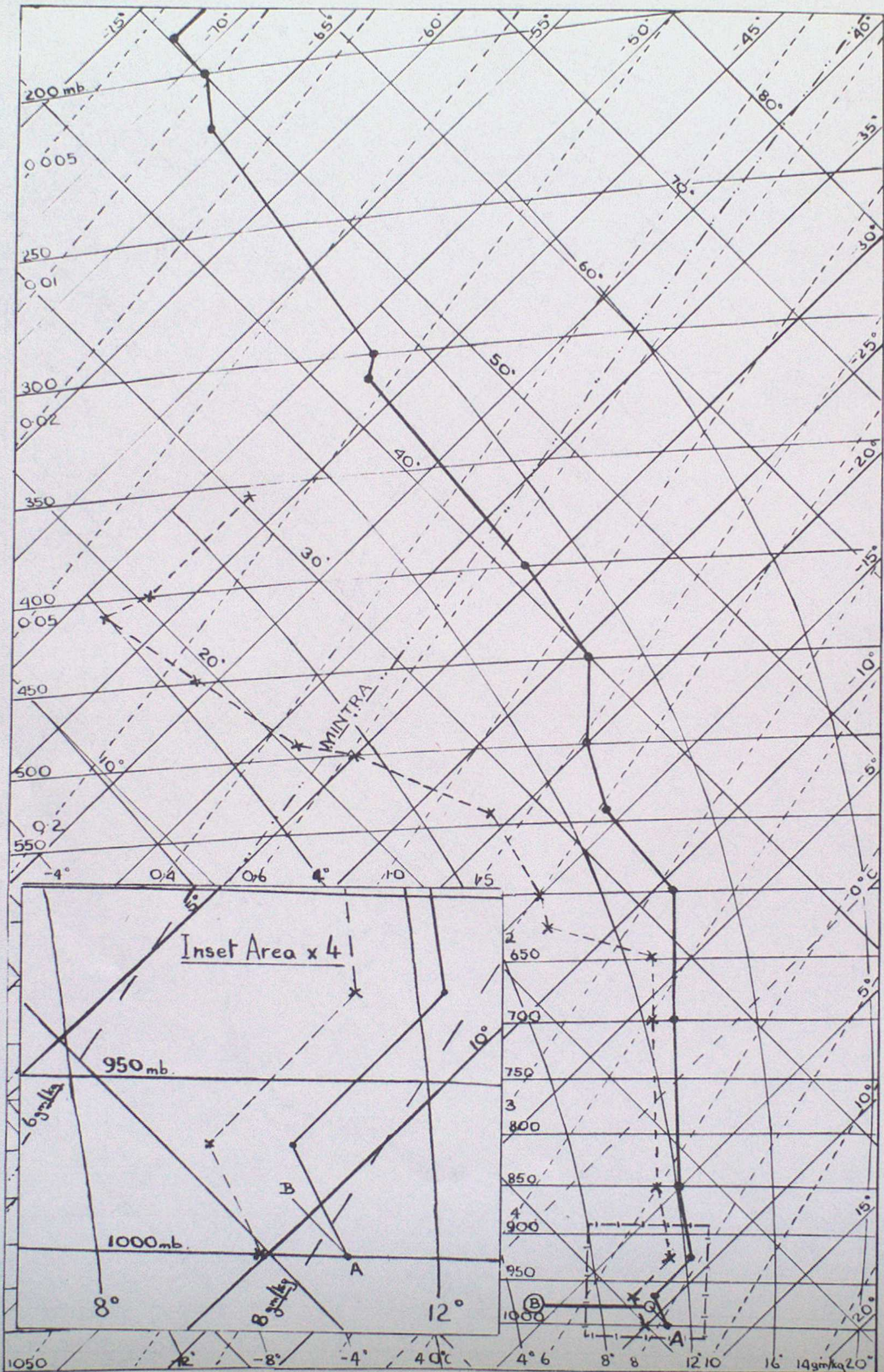

FIGURE 3.22 *Tephigram for Camborne, 0000 G.M.T., 21 January 1960*

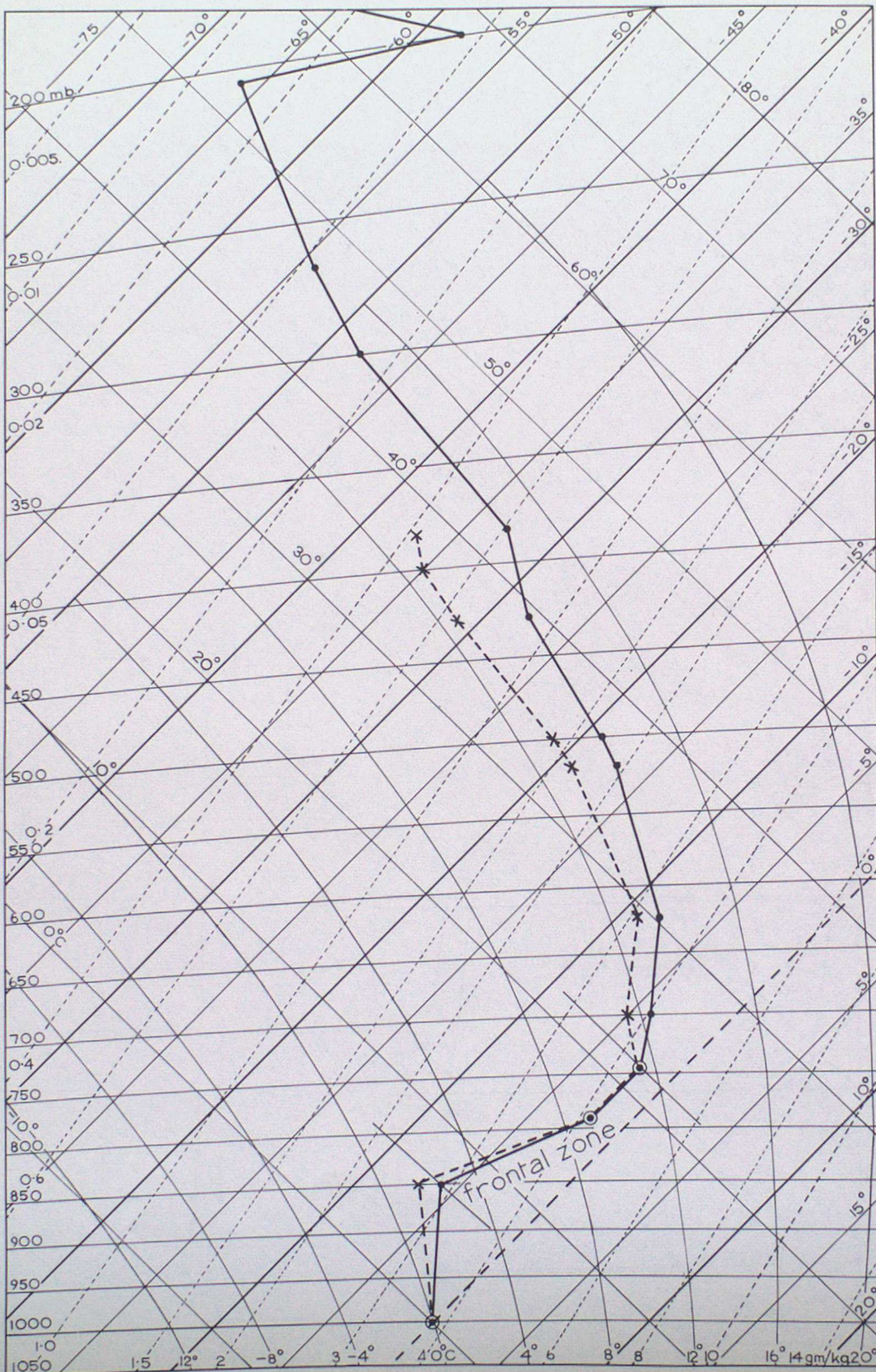

FIGURE 3.23 *Tephigram for Crawley, 0000 G.M.T., 21 January 1960*

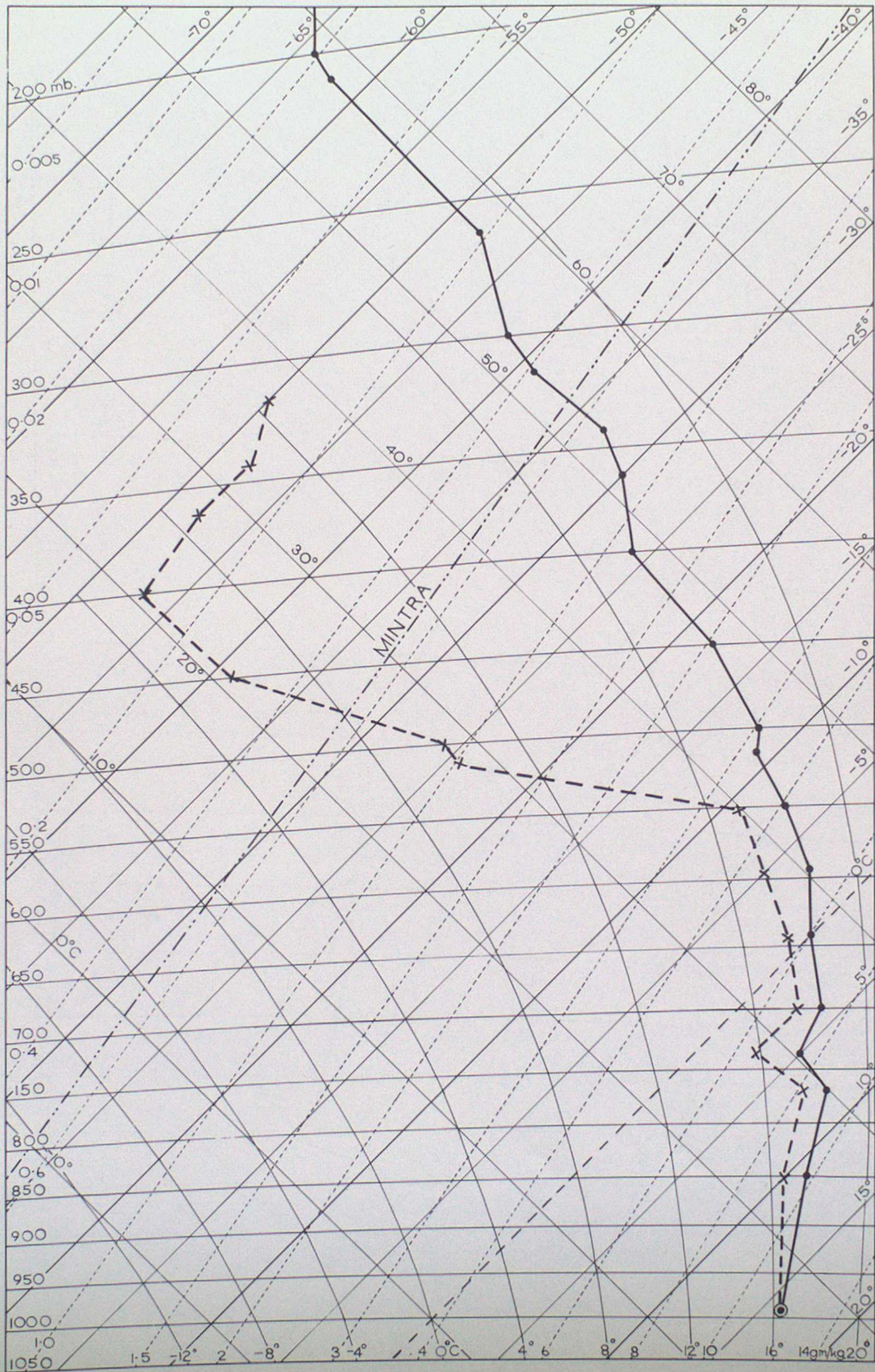

FIGURE 3.24 ***Tephigram for Camborne, 0000 G.M.T., 11 September 1957***

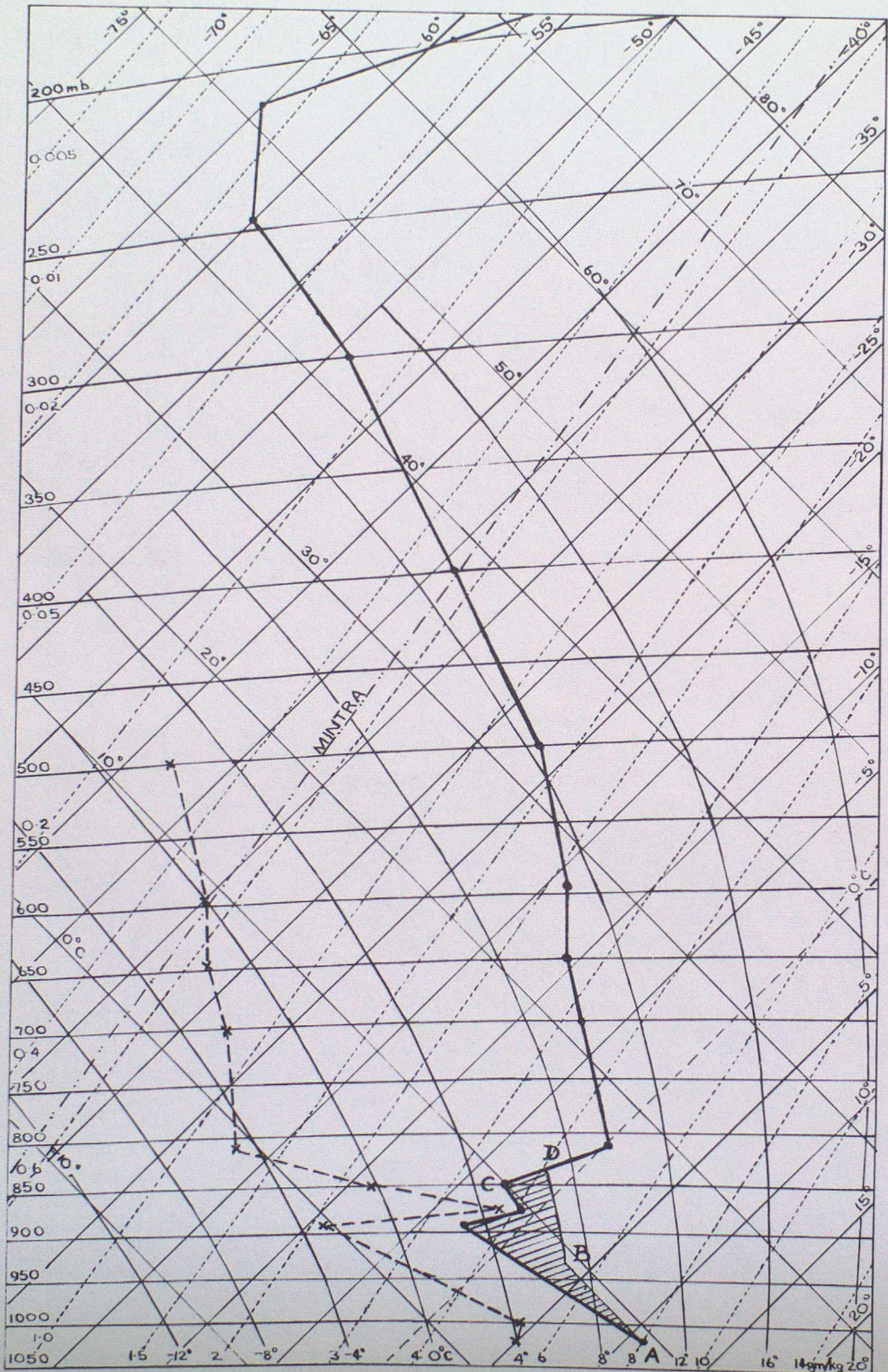

FIGURE 3.25 *Tephigram for Valentia, 1200 G.M.T., 6 May 1959*

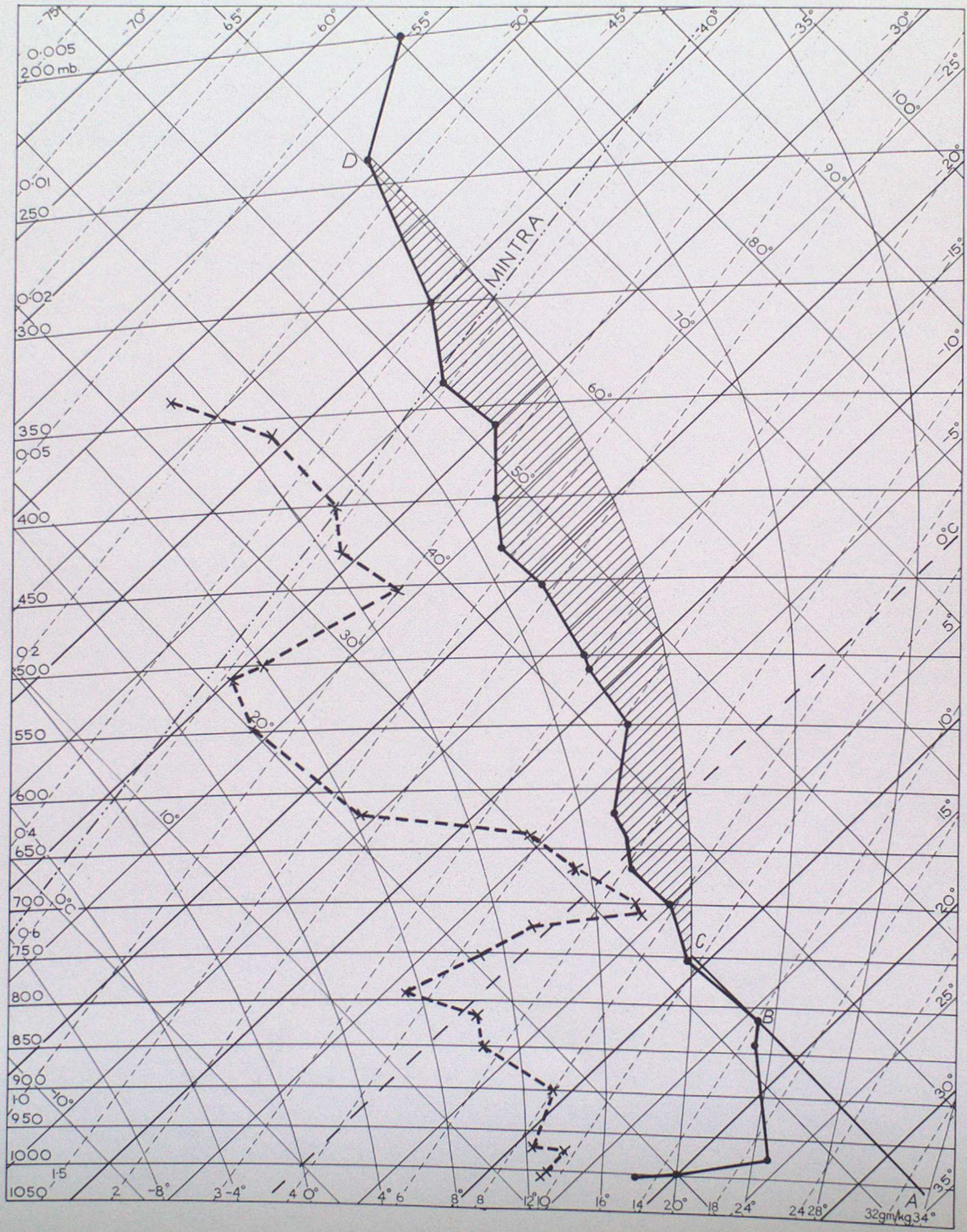

FIGURE 3.26 *Tephigram for Crawley, 0000 G.M.T., 29 June 1957*

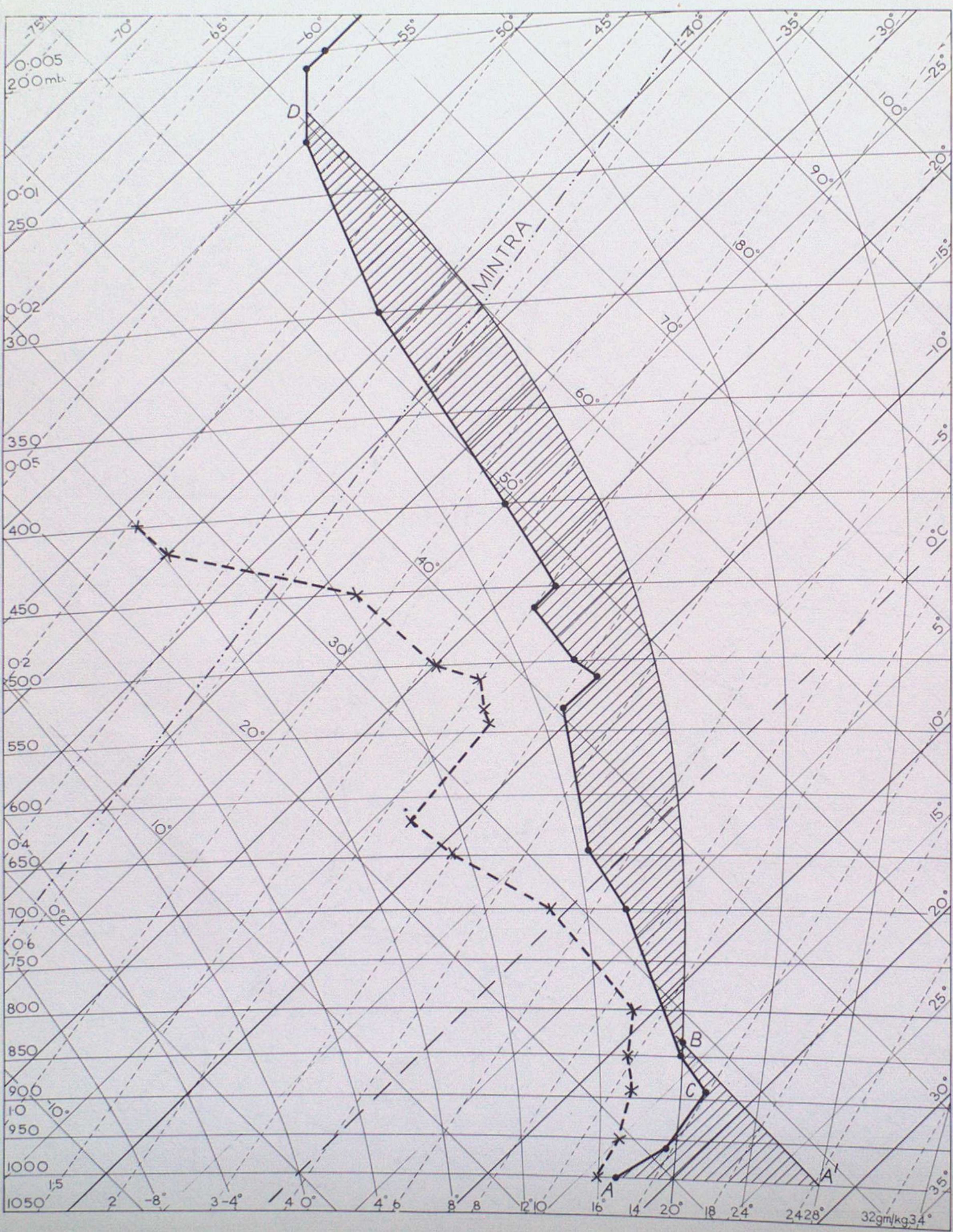

FIGURE 3.27 *Tephigram for Crawley, 0000 G.M.T., 1 July 1957*

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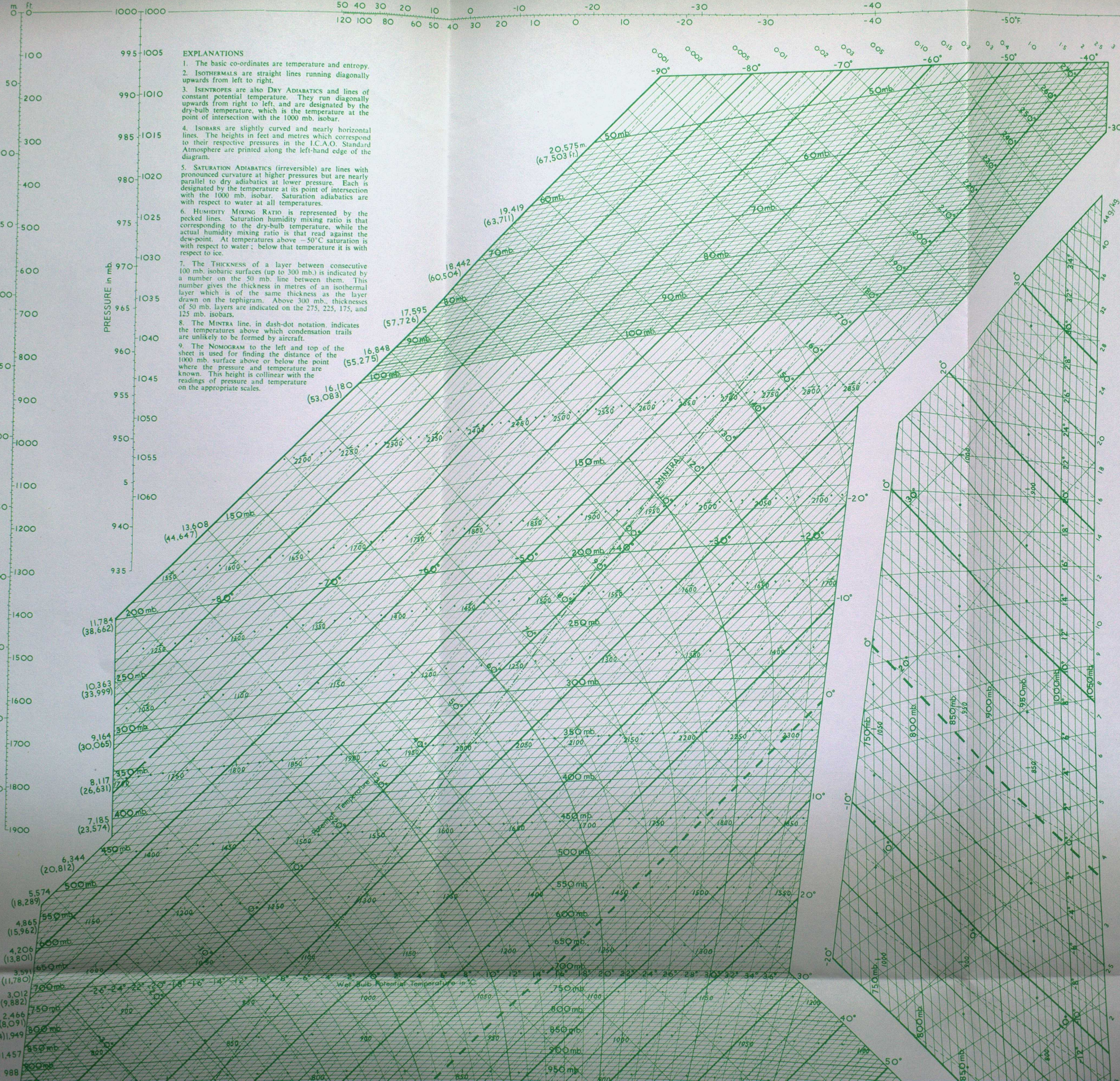
TEPHIGRAM

AIR MINISTRY METEOROLOGICAL OFFICE

Form 2810B (1956 edition)

Date 19



EXPLANATIONS

1. The basic co-ordinates are temperature and entropy.

2. Isothermals are straight lines running diagonally upwards from left to right.

3. Isentropes are also Dry Adiabatics and lines of constant potential temperature. They run diagonally upwards from right to left, and are designated by the dry-bulb temperature, which is the temperature at the point of intersection with the 1000 mb. isobar.

4. Isobars are slightly curved and nearly horizontal lines. The heights in feet and metres which correspond to their respective pressures in the I.C.A.O. Standard Atmosphere are printed along the left-hand edge of the diagram.

5. Saturation Adiabatics (irreversible) are lines with pronounced curvature at higher pressures but are nearly parallel to dry adiabatics at lower pressure. Each is designated by the temperature at its point of intersection with the 1000 mb. isobar. Saturation adiabatics are with respect to water at all temperatures.

6. Humidity Mixing Ratio is represented by the pecked lines. Saturation humidity mixing ratio is that corresponding to the dry-bulb temperature, while the actual humidity mixing ratio is that read against the dew-point. At temperatures above −50°C saturation is with respect to water; below that temperature it is with respect to ice.

7. The Thickness of a layer between consecutive 100 mb. isobaric surfaces (up to 300 mb.) is indicated by a number on the 50 mb. line between them. This 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., thicknesses of 50 mb. layers are indicated on the 275, 225, 175, and 125 mb. isobars.

8. The Mintra line, in dash-dot notation, indicates the temperatures above which condensation trails are unlikely to be formed by aircraft.

9. The Nomogram to the left and top of the sheet is used for finding the distance of the 1000 mb. surface above or below the point where the pressure and temperature are known. This height is collinear with the readings of pressure and temperature on the appropriate scales.

STATION				TIME	G.M.T.
STATION LEVEL PRESSURE				M.S.L. PRESSURE	
PRESSURE	HEIGHT	THICKNESS	TEMP.	PRESSURE	WIND
100		1/2		100	
200		2/3		200	
300		3/5		300	
500		5/7		500	
700		7/10		700	
1000		5/10		1000	
TROPOPAUSE				FREEZING LEVEL	
CLASS.	PRESSURE	HEIGHT	TEMP.	PRESSURE	HEIGHT

STATION				TIME	G.M.T.
STATION LEVEL PRESSURE				M.S.L. PRESSURE	
PRESSURE	HEIGHT	THICKNESS	TEMP.	PRESSURE	WIND
100		1/2		100	
200		2/3		200	
300		3/5		300	
500		5/7		500	
700		7/10		700	
1000		5/10		1000	
TROPOPAUSE				FREEZING LEVEL	
CLASS.	PRESSURE	HEIGHT	TEMP.	PRESSURE	HEIGHT

STATION				TIME	G.M.T.
STATION LEVEL PRESSURE				M.S.L. PRESSURE	
PRESSURE	HEIGHT	THICKNESS	TEMP.	PRESSURE	WIND
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200		2/3		200	
300		3/5		300	
500		5/7		500	
700		7/10		700	
1000		5/10		1000	
TROPOPAUSE				FREEZING LEVEL	
CLASS.	PRESSURE	HEIGHT	TEMP.	PRESSURE	HEIGHT

STATION				TIME	G.M.T.
STATION LEVEL PRESSURE				M.S.L. PRESSURE	
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1000		5/10		1000	
TROPOPAUSE				FREEZING LEVEL	
CLASS.	PRESSURE	HEIGHT	TEMP.	PRESSURE	HEIGHT

NOTES

TEPHIGRAM

AIR MINISTRY METEOROLOGICAL OFFICE

EXPLANATIONS

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7. The THICKNESS of a layer between consecutive 100 mb. isobaric surfaces (up to 300 mb.) is indicated by a number on the 50 mb. line between them. This 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., thicknesses of 50 mb. layers are indicated on the 275, 225, 175, and 125 mb. isobars.

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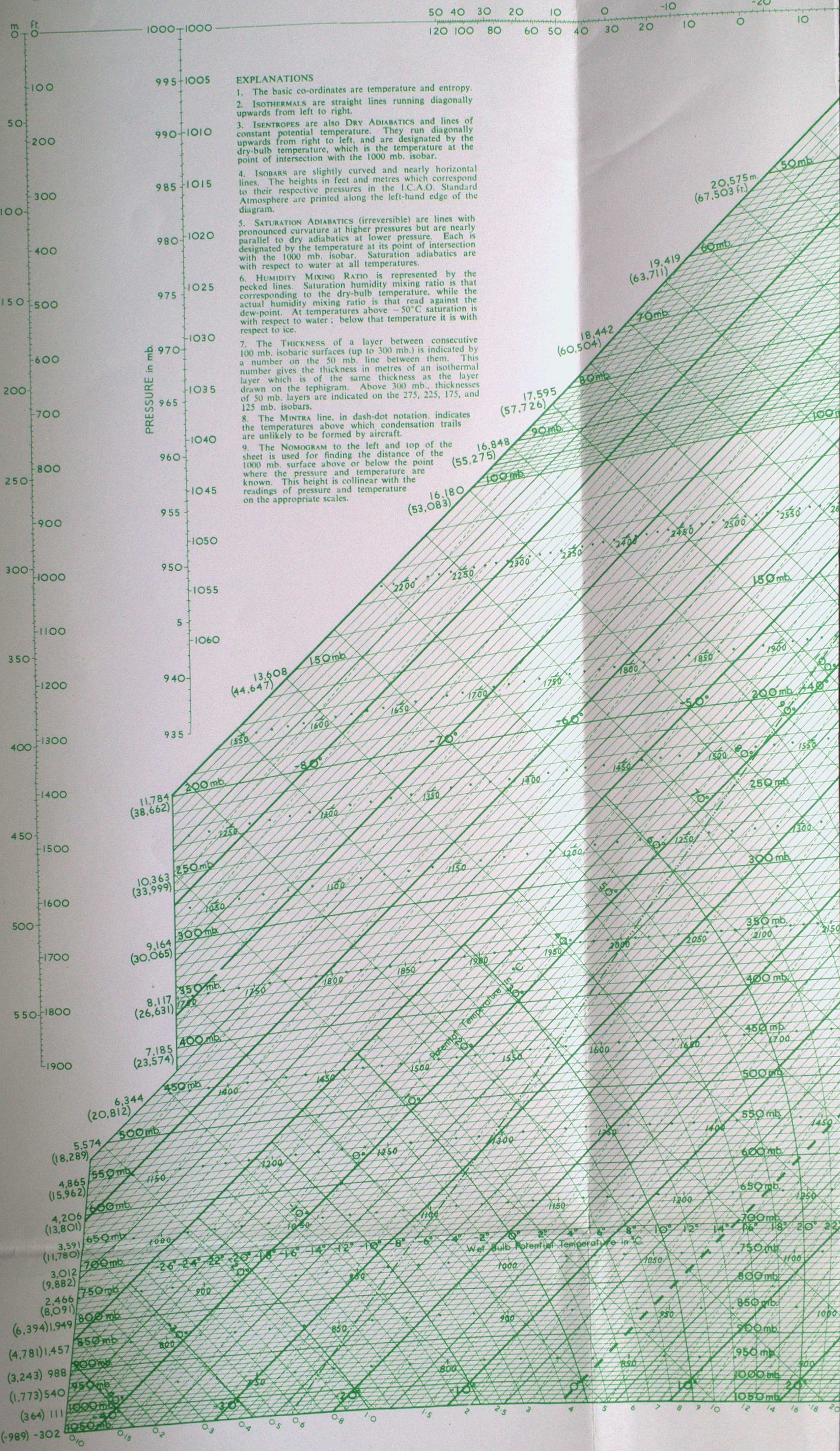

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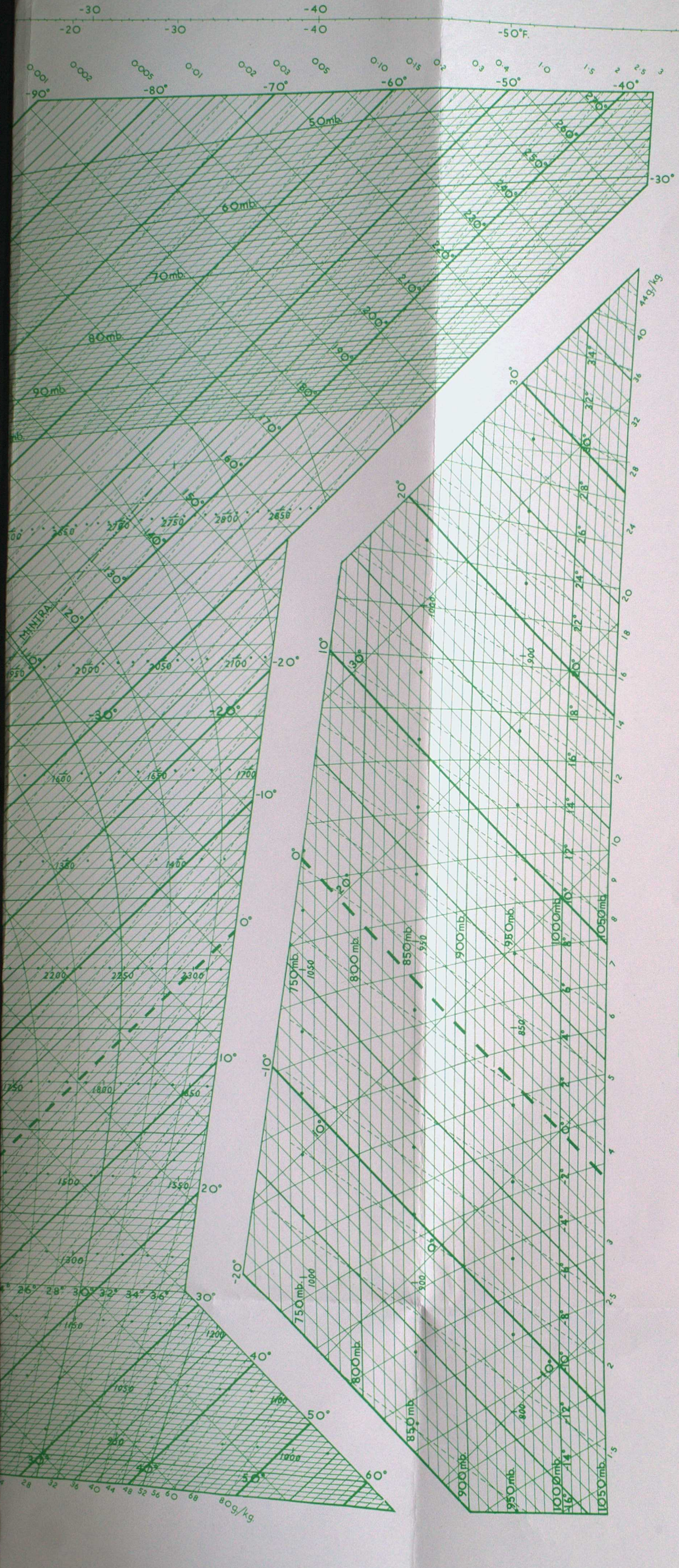
Form 2810 B (1956 edition)

Date...19.........



STATION				TIME	G.M.T.
STATION LEVEL PRESSURE				M.S.L. PRESSURE	
PRESSURE	HEIGHT	THICKNESS	TEMP.	PRESSURE	WIND
100		1/2		100	
200		2/3		200	
300		3/5		300	
500		5/7		500	
700		7/10		700	
1000		5/10		1000	

TROPOPAUSE				FREEZING LEVEL	
CLASS.	PRESSURE	HEIGHT	TEMP.	PRESSURE	HEIGHT

STATION				TIME	G.M.T.
STATION LEVEL PRESSURE				M.S.L. PRESSURE	
PRESSURE	HEIGHT	THICKNESS	TEMP.	PRESSURE	WIND
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200		2/3		200	
300		3/5		300	
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700		7/10		700	
1000		5/10		1000	

TROPOPAUSE				FREEZING LEVEL	
CLASS.	PRESSURE	HEIGHT	TEMP.	PRESSURE	HEIGHT

STATION				TIME	G.M.T.
STATION LEVEL PRESSURE				M.S.L. PRESSURE	
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CLASS.	PRESSURE	HEIGHT	TEMP.	PRESSURE	HEIGHT

STATION				TIME	G.M.T.
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CLASS.	PRESSURE	HEIGHT	TEMP.	PRESSURE	HEIGHT

NOTES