CHAPTER 11

UPPER-AIR CHARTS

CHAPTER 11

UPPER-AIR CHARTS

Contents | Page

LIST OF DIAGRAMS

Figure | Page

CHAPTER 11

UPPER-AIR CHARTS

11.1 INTRODUCTION

Since weather occupies three dimensions in space, no study of it is complete without a study of the vertical structure of the atmosphere. Observations of temperature, humidity, pressure and wind in the free atmosphere are recorded regularly at a number of land and ocean stations but the number of these stations is small compared with the number of stations for surface observations. This is largely because of the cost of maintaining upper-air stations, particularly over the oceans. Fortunately, irregularities in the spatial distribution of the values of the temperature, humidity and wind are not as great in the upper air as at the surface. This is not to be construed as meaning that the present distribution of upper-air stations is satisfactory. All forecasters would wish to have a closer network of stations and more frequent soundings, particularly so over the oceans where the present network of ocean weather ships is much too sparse even to approach a satisfactory state. In consequence of this it is of the utmost importance in the analysis of upper-air charts to ensure that no observation is neglected unless there are good grounds for considering it to be grossly inaccurate. This is not meant to imply that all observations should be accepted as exact; critical weighting of the various items of information is essential for sound analysis. The importance of continuity from one level to another and from one time to another in upper-air analysis cannot be overstressed. Over the oceans it is quite easy for an important feature on one chart to be lost 12 hours later in the space between simultaneous soundings made many hundreds of kilometres apart. It is absolutely essential that any feature clearly portrayed on one chart should not be neglected at higher or lower levels or dropped from subsequent charts without very good reason.

The traditional observations used on upper-air charts are, in the main, instrumental and, with the exception of the wind, are measured by an instrument (radiosonde) remote from the station. The patterns of instruments are various and unfortunately the accuracy of all patterns or even of all instruments of one pattern is not uniform. The inaccuracies fall broadly into two groups – systematic errors characteristic of the design of the radiosonde and the techniques employed in its operation, and non-systematic errors affecting any particular ascent. Irregularities due to the former are frequently noticed (at the time of writing) at national boundaries at the higher levels (mainly 200 mb and above) and largely arise from radiation effects on the radiosondes. This effect is most marked at high levels in the day-time, and it is noteworthy that such instrumental errors are generally a function of height so that, if an error is found or suspected at one level, it must be allowed for at other levels in the same ascent. A non-systematic error on any ascent may or may not affect the accuracy of the information for levels other than that at which it occurs, depending on the nature of the error and on the technique used for calculation of the results of the sounding. The errors of radiosondes have been discussed in Chapter 10; the effect of these errors on the calculation of geopotential heights is dealt with more fully in 11.3 on page 11.

Since the first edition of this handbook was written, further data sources have become available with the advent of satellite photographs and infra-red sounding systems, and reliable aircraft observations of winds. Photographs from satellites in orbit may be used to locate systems in the upper air as well as at the surface, while those from geostationary satellites enable cloud movements to be measured and hence winds at the cloud height to be derived, filling in some of the large gaps in the observational network. Satellite measurements of atmospheric infra-red radiances at various wavelengths enable estimates to be made of temperatures at a number of levels in the atmosphere; these data also have the advantage that they fill in the large gaps in the network of observations, but they are of variable quality and must be carefully checked before use. Winds reported by aircraft are almost invariably measured by Doppler radar or inertial navigation systems and provide a very valuable and reliable source of data for, in particular, the 300-millibar level. They have compensated in some measure for the reduction in the ocean weather station networks and, although the observations are restricted in general to the main air routes, they provide an excellent and hitherto unavailable opportunity for detailed analysis of the wind field at levels near that of the jet streams.

The observations in the upper air enumerated above can be plotted in a variety of ways. One way is to plot observations on charts of the earth's surface, each chart having observations for a particular height or level plotted on it. These charts fall into two groups, charts portraying the variation of pressure over surfaces of constant height above mean sea level and charts showing the variation of height over levels of constant pressure. Surface charts are examples of the former type, but for the upper air it is more convenient in practice to use charts of constant-pressure levels. This system has the following advantages. One geostrophic wind scale can be used for all levels (see section 2.2.3 of Chapter 2) and this scale can also be applied to the thickness isopleths giving a simple determination of the thermal wind (see section 2.2.6 of Chapter 2). Further, isotherms on isobaric surfaces are lines of constant potential temperature. In addition, upper-air observations as received are, for the most part, basically in terms of pressure.

Another way of making use of upper-air data is to plot the data on vertical cross-sections of the atmosphere. Such analysis may be carried out for special occasions and the methods are dealt with in Chapter 12. Two methods of analysis not in regular use in the Meteorological Office, isentropic analysis and frontal contour analysis, are also described in Chapter 12.

The upper-air charts in daily use in the Meteorological Office are constant-pressure charts: they are drawn on an operational basis at a series of standard levels, 1000, 850, 700, 500, 300, 250, 200 and 100 mb, while charts at levels above this up to 10 mb are being drawn regularly but mainly for research purposes. On each chart is plotted, for each sounding, the temperature (°C), the height of the pressure level (geopotential metres), and the wind (degrees true and knots). Contours of height are drawn on these charts at intervals of 60 geopotential metres (gpm). (See section 2.2.6.1 of Chapter 2.)

A further set of charts of considerable use to upper-air analysts and forecasters are the thickness charts. On these charts are plotted the thicknesses of the air column between two chosen pressure levels and, in addition, the vector differences between the winds at the two pressure levels. The isopleths of equal thickness are drawn at intervals of 60 gpm. The most frequently used thickness chart is for the layer 1000–500-mb. For some aviation purposes, charts are drawn at the level of maximum wind, usually combined with tropopause analysis.

Contour analyses at levels from 1000 mb up to 100 mb and of the 1000–500-mb thickness pattern are now carried out on a routine basis by computer. It is still necessary, however, for manual analyses to be drawn, both as a check on the computer products and as a basis for amendment of the computed patterns if they are deficient in some way. The computer analyses are based upon (a) a background field, the 12-hour prognosis based upon the previous analysis, and (b) the observations. Quality control is applied to the observations, but the procedure, however comprehensive, must be inflexible, so that observations may be accepted which are in error, or some which are correct may be rejected. The human forecaster may, on the basis of his analysis, 'intervene' to remove any observations which he considers are in error or to reinsert those which he believes are correct. He may also intervene to amend the contour or thickness pattern in areas where he feels that the computer analyses could be improved, for example where data are sparse and the analysis relies heavily on the background field, or where a fairly intense synoptic system may not have been represented adequately. The intervention consists of the insertion of 'bogus' observations which, it is hoped, will lead to a better computer analysis, or the direct manipulation of the contour field by means of a visual-display unit showing the computer analysis. The topic of computer analyses is discussed in rather more detail in Chapter 3 – Background to computer models.

11.2 PRINCIPLES OF ANALYSIS

The basic principle of the analysis is the drawing of contours to fit the reported heights, using the observed winds to estimate the direction and magnitude of the slope of the isobaric surface, i.e. the orientation and spacing of the contours. In practice, the analysis is complicated by the presence of errors in both the heights and the winds, and by the fact that, although the geostrophic wind approximation is reasonable over much of the chart, there are areas where the ageostrophic component of the wind may be significant. The errors will be discussed briefly below, with a more detailed account in 11.3 (page 10).

11.2.1 Errors of geopotential height

Errors in the radiosonde lead to errors in the computed geopotential heights which generally increase with height. In the upper troposphere and the stratosphere a large part of the error is a result of solar radiation falling on the instrument. The total error is composed of systematic and random errors: the random errors cannot be spotted unless there is a fairly close network of stations; systematic errors would not greatly affect analysis if all sondes had the same systematic error.

This, however, is not so and it is only too obvious to the analyst of the 100-mb charts that allowance must be made for the various kinds of sonde and methods of working up the results (for example, whether or not a solar-radiation correction has been applied before transmission). It is possible to determine, on a rough basis, corrections to be applied to the various groups of sondes to bring their heights into better agreement. To do this it is necessary to choose one group of sondes as standard,and select charts when, according to the reported winds, the airflow is simple with slack gradient. It is then possible to calculate the gradients from the wind reports and to establish contour heights over the chart in terms of the standard sonde heights. The systematic errors of sondes in the various geographical groups (mainly national) according to the solar elevation can then be determined. A considerable number of occasions are required in order to eliminate random errors. The 'systematic' errors thus found can then be applied as corrections to future reports during analysis (or by additional plotting).

These 'corrections' are best found at 100 mb but they also apply to some extent to 200 and 300 mb. Hawson and Caton[1] have suggested that the following proportions should be applied to the lower-level charts (expressed as percentages of the 'corrections' at 100 mb):

200 mb – 60 per cent   300 mb – 35 per cent   500 mb – 10 per cent.

(These same percentages can also be applied as rough corrections to individual anomalous soundings if it is believed that the abnormality increases progressively with height.) It will be appreciated that any change of radiosonde type, or method of working up results, will affect these corrections so that any table of them may become obsolete without warning at any time. They thus need to be kept up to date and no table is published in this handbook.

Where there is a close network of reporting stations, as in Europe and North America, a compromise fitting of the majority of winds and heights is not difficult to draw, and erroneous data can usually be detected fairly easily. However, over the sea, and in other areas where the observations are far apart, each station's report must be examined critically; if an erroneous observation is incorporated in the analysis it may well introduce errors over a large area, and consequent errors in the forecast over even wider areas. It is essential to make use of all possible checks, such as continuity in time and in the vertical, and the reasonableness of the analysis in comparison with climatological limits and known patterns. Some background information which should be useful in this respect is given in Chapters 4 to 8.

Occasionally it is necessary or useful to check, at least approximately, the height of one or more pressure surfaces above a given station. This can be done by the methods outlined in section 10.4 (page 5) of Chapter 10 – Upper-air ascents, using the tephigram. First, the height of the 1000-mb surface above sea level is found, and then the thickness of successive layers up to the required pressure level (using the mean virtual temperature instead of the mean dry-bulb temperature where the humidity correction is significant).

11.2.2 Errors of reported winds

The winds plotted on upper-air charts are mean values over a layer of the atmosphere traversed by the balloon in two, sometimes three, minutes and may therefore differ at times from the wind at the standard pressure level. In addition, the reported winds are subject to error from two sources. One is the random error of the range, bearing and elevation readings of the radar: the error increases only slowly with height. The second is a result of the geopotential height error, such that the reported wind is not appropriate to the height of the required pressure level: this error may be important in regions of strong vertical wind shear.

The errors of reported radar winds are discussed more fully in 11.3.2.2 (page 11).

11.2.3 Allowance for curvature of the flow

If the airflow is nearly straight and no rapid changes are taking place, the contour spacing can be simply derived with reasonable accuracy from the wind speed by use of the geostrophic relationship. If the flow is curved, an allowance must be made for the curvature (see section 2.2.3.2 of Chapter 2 and section 16.2 of Chapter 16). A single report cannot indicate whether the isobars are curved or not, but if this can be inferred from earlier or other data, a rough correction can be made to the spacing of the contours according to the expected departure of the wind from geostrophic. A relatively simple method of deriving this contour gradient has been proposed by Boyden.[2] Using the reported speed, visualize on the chart the circular path along which the air is moving and note the number of degrees of wind-direction change if this movement were continued for a number of hours (t) depending only on the latitude (ϕ). t and ϕ are related thus:

ϕ degrees	t hours	ϕ degrees	t hours
72	3½	42	5
56	4	37	5½
48	4½	33	6

Then the number of degrees (backing is counted as positive and veering as negative) is the percentage adjustment required to convert the reported wind to an equivalent geostrophic value which may be used with the geostrophic scale to space out the contours.

11.2.4 The 'gridding' technique

In the days before computer analyses were available, the technique of 'gridding' or the graphical addition or subtraction (sometimes known as 'de-gridding') of charts, was in widespread use. It enabled the analyst to 'build up' a series of charts from the surface to the upper troposphere in a way which ensured a high degree of consistency in the vertical and which made maximum use of the observations.

The advent of computer analyses made the gridding technique superfluous except in data-sparse regions and above 100 mb, and for a time it was little used. However, since the reduction in the ocean weather station network, it has once again become a valuable tool in ensuring vertical consistency in difficult areas, and a brief description has been included here.

The addition of two sets of isopleths to give a third set may be represented by the equation

$$S = F_1 + F_2 \qquad \cdot\ \cdot\ \cdot\ \cdot\ 11.1$$

where the quantities S, F_1 and F_2 can all be represented by sets of isopleths on the same basic chart, and at every point the value of S is the sum of the values F_1 and F_2 at that point.

If we differentiate this equation (in the plane of the chart) we get:

$$\nabla S = \nabla F_1 + \nabla F_2. \qquad \cdot\ \cdot\ \cdot\ \cdot\ 11.2$$

Thus the vector gradient of the sum S is the vector sum of the gradients of each of the two quantities F_1 and F_2.

The graphical process saves time because it obviates making separate additions for a large number of points and, if the original data for F_1 and F_2 contain errors, best-fit isopleths for F_1 and F_2 will produce best-fit isopleths for S. If there is doubt about the best fit of the isopleths for F_1 and F_2 they can be adjusted whilst obtaining the isopleths for S according to the operator's judgement. All sets of isopleths may be drawn on the same chart or, if a light-table is used, the isopleths may be on different charts which are superimposed in register on the light-table. In meteorology this technique is much used for differential analysis of heights of pressure surfaces and so the examples given here will be for the construction of contour charts for one level by the use of a contour chart for a lower level and the thickness chart for the intervening layer.

An example of the graphical addition of two sets of isopleths is shown in Figure 1. At each point of intersection A, B and C, the sum of the 1000-mb height and the 1000-500-mb thickness is 5340 geopotential metres (gpm), and at no other point of intersection of the isopleths drawn in the figure is the sum equal to this value. The 500-mb contour for 5340 gpm therefore passes through A, B and C and through no other point of intersection. The shape of the contour at intermediate points can be obtained by drawing in additional isopleths of height and thickness but this is rarely necessary in actual practice. It is, however, important to realize that the upper contour is defined at all points, and the temptation to draw it midway between the lower contour and the thickness line must be resisted. It should be noticed that, since from equation 11.2 the gradients are added, the direction of the third isopleth at any point of intersection can be obtained rapidly, since the vector geostrophic wind for the upper level is the resultant of the vector geostrophic wind of the lower layer and the thermal wind (see Figure 2), and in direction lies between the other two vectors. This fact is of assistance in ensuring that the isopleth produced in the addition passes through the intersection in the correct way, and allows the operator to start easily at any point of intersection.

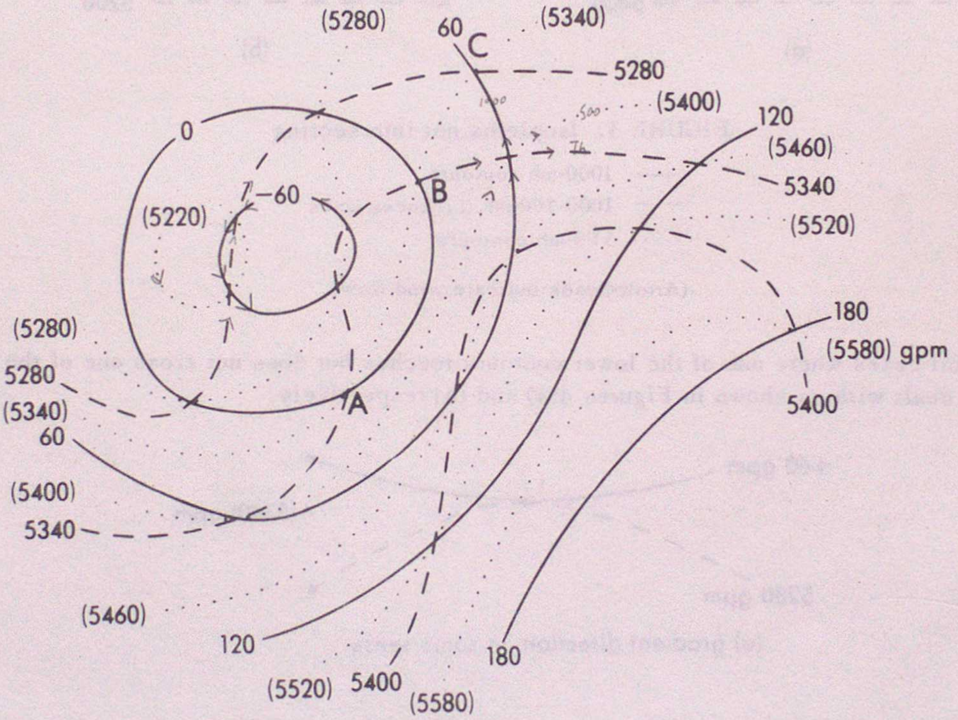


FIGURE 1. Gridding of 1000-mb contours (——) with the 1000-500-mb thickness lines (– – –) to obtain 500 mb contours (· · ·)

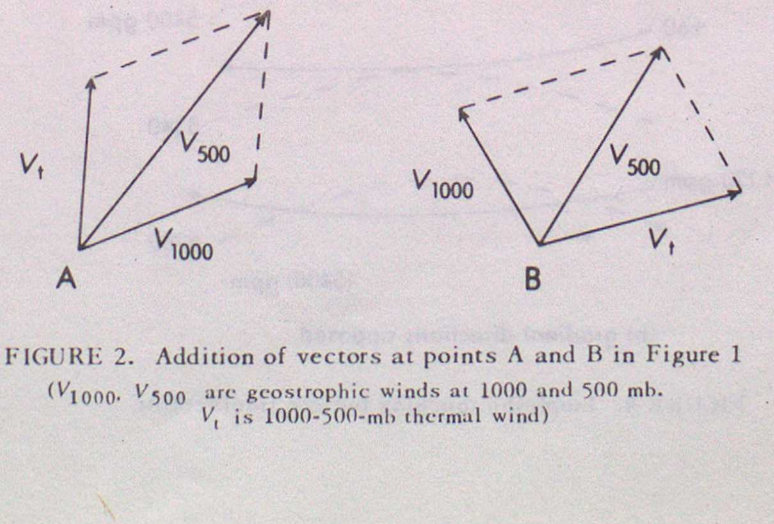


FIGURE 2. Addition of vectors at points A and B in Figure 1

(V_{1000}, V_{500} are geostrophic winds at 1000 and 500 mb. V_t is 1000-500-mb thermal wind)

The case of the two families of isopleths not intersecting but lying between each other needs careful drawing. In Figure 3(a) the lower geostrophic wind and the thermal wind are approximately in the same direction, and in Figure 3(b) they are opposed (that is, gradient in opposite directions). (On actual charts the lines are unlikely to be straight as in these diagrams but the argument is the same.)

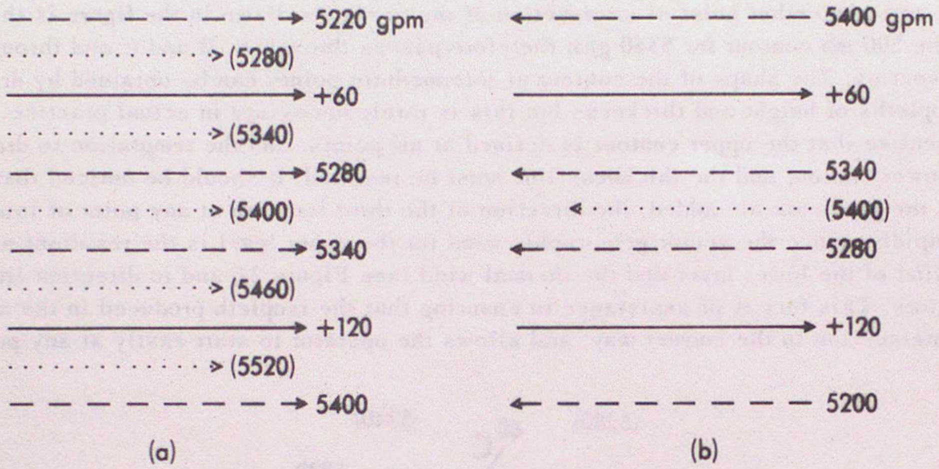


FIGURE 3. Isopleths not intersecting
——— 1000-mb contours
— — 1000-500-mb thickness lines
...... 500-mb contours
(Arrow-heads indicate wind flow)

The special cases where one of the lower contours touches but does not cross one of the thickness lines can be dealt with as shown in Figures 4(a) and (b) respectively.

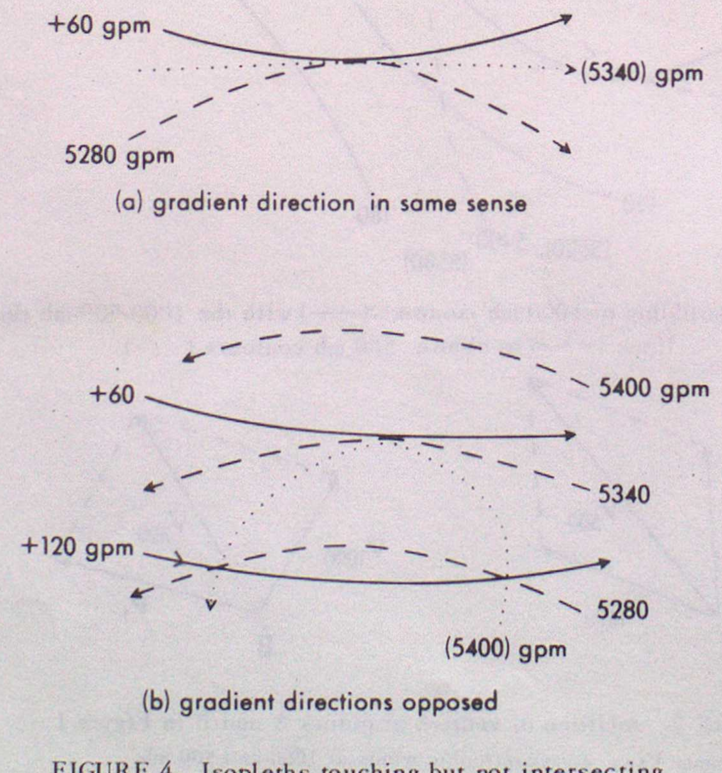


FIGURE 4. Isopleths touching but not intersecting

The reverse process of subtraction by graphical methods may be necessary on occasions: for example, to find the thickness between two layers p_1 and p_2 whose contours are given. Then to use the graphical process for $z' = z_2 - z_1$ it is necessary to note that:

$$\nabla z' = \nabla z_2 + (-\nabla z_1),$$

so that to draw the contours through the intersection it is necessary only to imagine the direction of the geostrophic wind vector(∇z_1) everywhere reversed and, using the reversed direction, proceed as in addition.

11.2.5 The use of satellite and aircraft data

Photographs of the earth's surface and clouds are transmitted from geostationary satellites at regular intervals, say every 20 minutes. The displacement of a given cloud element over the interval between the times of any two photographs may be used to determine the mean wind speed at the cloud height over that interval. There is usually some uncertainty about the cloud height, although a reasonable idea is often obtainable from the infra-red photographs. There may also at times be errors and uncertainties arising from the developments of the cloud. The area covered by each satellite is restricted both in latitude and longitude; data are not available at latitudes higher than about 40-45°, for example, and not all longitudes are covered at any one time. Nevertheless, the data help to fill what are normally large gaps between the routine radiosonde and radar-wind ascents.

Reports of winds determined by aircraft are a valuable help in the analysis of the 300-mb contour pattern. Although the associated contour heights are not available, contour gradients may be derived from the observed winds; normally this is done by assuming that the geostrophic relationship holds and relating the derived pattern to any available radiosonde data. Adjustments may be made, e.g. for curvature, to improve the fit if necessary. Near jet streams, accelerations and, therefore, ageostrophic motions are important, and corrections are more difficult and uncertain, but the overall result of the availability of aircraft winds has been a marked improvement in the quality of the analyses at 300 mb.

Two forms of data of direct use in upper-air analysis are available from satellites in orbit – cloud photographs, both visible and infra-red, and temperatures and thicknesses at a number of levels based on observations of infra-red radiation from the atmosphere.

Satellite cloud photographs have an obvious use in the preparation of surface analyses: for example, in helping to determine frontal positions; their use in upper-air analysis may be less obvious, but it is nonetheless very valuable. The main principle behind their use is that ascending motion leads to cloud, and descending motion is associated with relatively cloud-free air. Thus, between a trough and ridge in the upper westerlies there is usually an upward vertical component leading to a good deal of cloud. Conversely, in flowing from a ridge axis to a trough there is descending motion with only small amounts of cloud, mainly cumuliform. The boundaries of the cloudy air are related to the ridge and trough line in a way which depends upon the characteristics of the feature. In a broad, flat ridge the downward motion to the east of the ridge line is quite gradual and the cloud may extend some distance beyond it, whereas if the ridge is strongly curved there will be a sharp cloud edge near the ridge line.

A trough axis may often be located where it crosses a frontal cloud band: sinking air to the west of the trough causes the middle and high clouds to dissipate, leaving only a narrow band of low clouds, while to the east the cloud band is broader, more solid and of greater vertical extent.

Since the more extensive cloud bands extend roughly from a trough axis to a ridge line, an approximate idea of the amplitude of the flow may be gained from the latitudinal extent of the cloud bands. If the cloud band is quasi-linear the direction of the contour lines is roughly along the band. There may be vortices on the poleward side, representing short-wave, rapidly moving disturbances. There may also be, particularly in large-amplitude flow, cloud vortices or spirals which indicate closed circulations such as old occluded depressions or cut-off lows, the centre of the cloud spiral being generally coincident with the circulation centre. For further details and examples of the use of satellite photographs in upper-air analysis, the reader should consult WMO Technical Note No. 124.[3]

Satellite observations of infra-red radiation from the atmosphere enable temperatures at a number of levels in the atmosphere to be estimated and, from these temperatures and the corresponding pressures, thickness values for a number of layers may be derived. The satellite data are most readily interpreted over the sea in cloud-free regions, and in these cases provide a valuable addition to the forecaster's sources of data. It should be added, however, that some uncertainties exist, and some data may not be usable; regular checks are necessary and these are usually carried out by comparison with radiosonde data at 100 mb.

11.2.6 Analysis procedures

The traditional method of upper-air analysis was to start with a 1000-mb chart, traced directly from an analysed surface chart which can be regarded as a firm basis from which to work because of the plentiful supply of observations. The analyst would then build up from this to successively higher standard-pressure levels by using the gridding technique, which has the dual advantage that it ensures vertical consistency in the analyses and also makes the fullest possible use of the data.

The advent of computer-produced analyses and the introduction of the important new data sources mentioned in 11.2.5 have led to radical changes in the analysis scheme and in the uses of the products. The greatest difference in the analysis scheme is that there are now two levels for which observations are plentiful and reliable – the surface and 300 mb. The procedure is to use the surface chart as a base for drawing the analyses up to 500 mb, and to use the 300-mb chart as a base for drawing the analyses at higher levels and for checking, and if necessary modifying, the charts at 500 mb, and possibly below, to ensure vertical consistency.

The manual analyses are used as a check on the computer analyses; although the details of the 'man-machine mix' may change with time, it is worth while to give here a brief indication of present (1976) procedures.[4] The scheme is designed to make the best possible use of the computer, with its ability to process rapidly a large quantity of data, and of the human being, with his ability to exercise judgement based on experience and complex logical processes. The computer takes as a 'background' field the 12-hour forecast based on the analyses at the last main synoptic hour, 12 hours earlier; the background field is then modified by the new data. In regions where data are plentiful the new analysis is based mainly on the data, but where data are sparse the background field is little changed. The data, before incorporation in the analysis, undergo quality control in the computer, and those which are in error, according to certain fixed criteria, are rejected. The human analyst also examines the data to see whether

(a) any observations that contain errors can be corrected, perhaps after a detailed examination of the ascent or after comparison with neighbouring observations;

(b) any good observations have been rejected; or

(c) any poor observations have been accepted.

Any faults detected in the computer quality-control procedure can be corrected by intervention, although there is still room for human errors of judgement in deciding how the observational data should be treated. The intervention consists of reinserting good or corrected observations.

In some areas where there are few observations at the analysis time, the analyst may have some idea of developments from intermediate data or from other sources (aircraft, satellite data, etc.), and here the background field may be amended, by the insertion of 'bogus' observations, in an attempt to take the extra data into account.

The operational analyses and forecasts must be carried out according to a strict timetable, at times when not all the data for the analysis time have been received: more will be said about this aspect later. The analyses for the various levels are carried out in four stages. Firstly, the charts for the surface and for 100 mb are analysed; from the differences between the analysed 100-mb field and the radiosonde data for 100 mb, corrections for individual sonde ascents or for groups of ascents are derived by the method of

Hawson and Caton[1] for the 300-mb and 500-mb levels. These two levels are now analysed; the difference between the analyses and the background fields are used to modify the background fields at 850, 700, 400 and 200 mb and these four levels are then analysed. Each analysis is carried out in three 'sweeps', the first using observed heights only, the second using observed heights and winds reported together, and the third using all observed heights and winds. Finally, polynomials are fitted to the eight analysed levels in order to ensure vertical consistency and to derive the fields at 900, 800 and 600 mb for the 10-level model. The operational forecasts are based on these analyses.

When later data are available the computer analyses are re-run – the 'update run'. Intervention may take any of the forms described above, and also by direct modification of the analyses fields via a visual-display unit. The intervention is carried out at the four main analysis levels, the surface or 1000 mb, 500, 300 and 100 mb, and influences the other levels indirectly through the procedures designed to ensure vertical consistency.

11.2.7 Jet-stream analysis

No set of upper-air charts for middle latitudes and extending to the tropopause can be considered complete unless some attention is given to the jet streams. A jet stream is defined by the World Meteorological Organization as 'a strong narrow current concentrated along a quasi-horizontal axis in the upper atmosphere, characterized by strong vertical and lateral wind shears and featuring one or more velocity maxima'. The vertical wind shear is of the order of 10-20 knots per thousand metres, the lateral shear about 18 knots per hundred nautical miles and an arbitrary lower limit of 60 knots is assigned to the speed of the wind at the core. A full account of jet streams and their features is given in Chapter 8, and the present section is restricted to their representation on the constant-pressure charts considered in this chapter.

On such a chart, the presence of a jet stream is indicated by a belt of more or less closely packed contours lying along the belt of strong winds. The belt may be hundreds or thousands of kilometres long; it is usually curved and may have a few simple branches. In middle latitudes, the jet streams are often at about 300-250 mb so that the standard level at which these jets are most marked is usually 300 mb. In the case of subtropical jet streams, the 200-mb chart is more useful since the cores of these jets are mainly at about 200 mb; however, these jets are well marked at 300 mb as well. It is therefore usual to mark the jet streams on the 300-mb chart first and then to proceed upwards and downwards. Usually jets are not found below 500 mb or above 150 mb. The core of a jet is, of course, rarely coincident with a standard level so that, in general, no chart will portray the maximum winds. Another point to remember is that the axis of the various isobaric cross-sections of a jet are not vertically above one another but, both above and below the core, are often progressively displaced towards the tropospheric warm air.

The remarkable wind shears of the jet streams are perhaps best shown in vertical cross-sections perpendicular to the line of the core. On contour charts, the close packing of the contours along the quasi-horizontal section of the jet does not indicate directly to the eye either the location of the jet axis or the wind strengths and shears. These latter features are brought out in a much more prominent manner if isotachs (lines of equal wind strength) are drawn on the chart. For such purposes, the isotachs can be drawn at intervals of 20 knots and, where possible, to fit the actual winds reported. On such charts, the jet axis is usually indicated by a double or thick line with the actual values of the velocity maxima along it inserted at appropriate places. Over the continents of Europe and North America, the network of upper-air observations is sufficient to allow a fairly accurate placing of the jet axis and isotachs on each chart almost on wind reports alone. Over the oceans, the paucity of data makes such recognition impossible and one must then resort to more indirect methods of analysing the chart. Once again, continuity in time and space becomes of the utmost importance and every observation must be given full weight unless there is very definite evidence against it.

In the placing of jet streams over the oceans, the relation of the jets to the surface frontal systems should be borne in mind. In middle latitudes, the jet streams usually exist in close association with the strong thermal gradients of well-marked fronts and, although the core of the jet will lie in the warm air a little below the tropopause, the slope of the frontal surface is such that the core actually lies well to the cold side of the surface-front position. In fact, an approximate rule is that the core lies vertically above the position of the frontal surface at 500 mb. It should be noted, however, that the subtropical jets are not so related to frontal features and that, when their meanderings are sufficiently large to bring them or branches of them northward, no attempt should be made to tie them to surface fronts. Similarly, branches of middle-latitude jets sometimes occur which cross surface cold fronts in the rear of depressions.

Although there are material differences between one jet stream and another in middle latitudes, it has been noticed by several workers in this field that there is a strong family resemblance and, over considerable areas of the world, the horizontal velocity profiles perpendicular to the core conform to a pattern. As a general rule, one can assume that the wind speed drops to 75 per cent of the axis value at 100 nautical miles (185 km) from the axis on the tropospheric cold side, and at about 150 nautical miles (≈ 280 km) from the axis on the warm tropospheric side. According to Johnson[5] this rate of fall continues to about where 50 per cent of the axis value is reached, but Endlich and McLean[6] found a very considerable decrease in the horizontal wind shear, especially on the cold side, beyond the wind speed of 70 per cent of the axis value. In the absence of other more definite information, the analyst can make use of this mean profile in the placing of the contours and isotachs about a jet stream. The winds on the axis of a jet stream vary along its length having one or more maximum values. When the wind is accelerating along the axis, the airstream can be expected to cross the contours towards the low side and vice versa. Simla's[7] approximate calculations show the order of magnitude of the cross-contour angles. At about latitude 50°N the inclination to the contours is about 10° when the speed changes by 40 knots in 600 miles (965 km) (measured along the jet axis), 20° when the 40-knot change occurs in 250 miles (400 km) and 30° when it occurs in about 150 miles (240 km). At 60°N the angles are about 10 per cent smaller; at latitude 30°N they are nearly 50 per cent larger. Since, in general, the axis of a jet can be assumed to be a streamline, this means that the jet axis will cross the contours in like manner. Occasions will be found, however, when the axis of maximum wind is not completely coincident with any streamline and then the jet axis may not cross the contours in the manner just described.

11.2.8 Tropopause analysis

In section 8.2 (page 7) of Chapter 8, a description is given of the features of the tropopause of which the analyst must be aware. Strict application of the WMO definition of the tropopause (quoted on page 7 of Chapter 8) may occasionally give misleading reports which could, in particular, omit the lowest tropopause; continuity from previous charts should help to indicate when this has happened, and recourse to the individual ascents may be necessary to obtain the true value for the analysis. It is useful to remember that the potential temperature at the tropopause does not change much over 24 hours, and that in areas well away from tropopause breaks the pressure at the tropopause will rarely change by more than 2 mb h^{-7} at a given location.

Tropopause funnels or lows may cause some difficulties in analysis as they are often relatively small and may not be indicated directly by the observations on a particular chart. If they extend below the 300-mb level they are normally aligned vertically with warm centres of the 300-200-mb thickness pattern. On the other hand, tropopause domes or highs are usually aligned with warm areas in the 500-300-mb thickness pattern.

Tropopause contour gradients often show little slope over large areas, while the gradient is concentrated into narrow bands at tropopause breaks which are, in general, associated with both the subtropical jet stream and the polar-front jet stream.

11.3 ANALYSIS OF HIGH-LEVEL CHARTS

11.3.1 Introduction

The purpose of this section is to discuss the circumstances which are encountered in the analysis of synoptic charts at isobaric levels between 100 and 10 mb, and to describe the techniques which are employed to reduce the difficulties which arise.

When the network of observations is sufficiently dense, a cursory examination reveals conflict between the reported winds and the geostrophic winds associated with the reported heights. Unlike the state of affairs which obtains at levels in the troposphere, the geopotential and wind fields no longer interlock and complement each other.

The first question to be considered is: do the familiar laws of motion and physics which apply in the troposphere break down? The answer is no. They continue to apply, because no basic changes in composition of the fluid or forces encountered occur until dissociation and associated electrical forces become involved. Perceptible ionization caused by the sun's ultra-violet radiation begins at about 60 km. In some problems, e.g. radio-wave propagation, the ionized component present in the 60-90-km layer is already important. But, in analysis of atmospheric motions below 100 km, the ionized component can be neglected (Kochanski[8]). Furthermore, the amplitudes of the atmospheric tides in the stratosphere are smaller than the errors of single wind observations.

The main reason for the apparent conflict on the high-level charts is found to lie in errors of observation associated with reported values, especially in reported geopotentials. We therefore begin with a review of the elements observed and the inaccuracies of their measurement relative to our problem.

As a preliminary, it should be noted that, except for relatively minor effects of mean temperature, the thickness of a layer is broadly controlled by the ratio of the pressure surfaces bounding the layer. Thus readers already familiar with the typical heights of surfaces from 1000 to 100 mb are readily able to familiarize themselves with the typical heights of higher surfaces. For example, the heights of the 1000- and 100-mb surfaces are typically 0 and 16 km, so that the thickness of the 1000 to 100-mb layer, or any other layer of the same mean temperature bounded by pressure surfaces whose ratio is 10:1, is about 16 km. Consequently we may expect the height of the 10-mb surface to be about 16 + 16 = 32 km (or, allowing for the lower temperature in the 100 to 10-mb layer, about 1 km less). Similarly we may infer that the height of the 1-mb surface is about a further 16 km higher than the 10-mb surface, that is 31 + 16 = 47 km, or use knowledge of thicknesses associated with other tropospheric pressure ratios. This provides a broadscale view of the general heights of isobaric surfaces, but of course on closer inspection isobaric surfaces vary in geopotential with time and location. When winds are strong the actual geopotential of an isobaric surface at one time can vary over the Northern Hemisphere by as much as 3 km (for example, at 10 mb in winter).

11.3.2 Observations

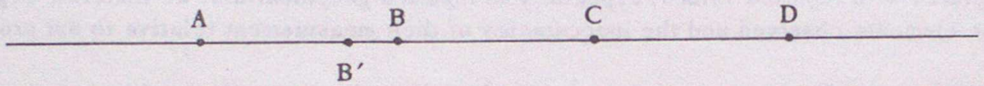
11.3.2.1 Humidity. The low-temperature region around the tropopause separates the stratosphere from the major water sources at sea level. The saturation vapour pressure of water at these temperatures is very low and this region acts as a barrier which largely inhibits upward passage of water vapour. In consequence, the humidity mixing ratio of the stratosphere is small, and effects of inaccuracies of humidity measurements on the calculations of geopotential in the stratosphere, via the virtual temperature increment and thickness, are negligible. Nevertheless, by the process of successive additions of pressure-layer thicknesses adopted in the calculation of geopotential, any such errors arising in the lowest layers of the troposphere are carried upwards almost unchanged to the highest levels reported.

When only small quantities of water vapour are available to activate humidity sensors the lag coefficient of the sensor becomes very high. This is why the lag coefficient of all radiosonde humidity units increases rapidly as atmospheric temperature decreases and the routine instruments are regarded as useless when the temperature is lower than −40°C. In fact lag difficulties are evident at −20°C.

In practice, humidity is rarely reported at the levels under consideration here and errors contained in the reported geopotentials stemming from humidity are negligible compared with those from other sources.

11.3.2.2 Wind. Wind observations are mostly derived by tracking free-flying ascending balloons. The reported winds are usually obtained by measuring the displacement of a balloon over time intervals of 1 to 5 minutes (depending on the uncertainties of the observation, with 2 minutes perhaps the most common choice), and are mean values for layers about 350-2000 m in depth. In some systems the track is independent of any airborne pressure and temperature observations, in others these elements are invoked to derive the track. In many systems the airborne pressure observations are used to determine the (pressure) level to which the measured wind is assigned. When radar is involved, an alternative independent measurement of (height) level is readily available and is sometimes better. Errors in the pressure (or height) observations can result in the measured winds being ascribed to erroneous levels, and thus, through the effect of the shear of wind with altitude, introduce errors in the reported winds. Usually wind shears with height are not large and this effect is small. However, when the vertical wind shear is very strong, especially at low pressures, this effect can be a major source of error, much bigger than those usually arising from other sources. Indeed, in the early stages of a disturbance of the winter stratosphere, vector errors of this type in excess of 50 knots in the vicinity of 10 mb have been encountered.

Measuring a displacement over a layer effectively yields a mean wind over the layer so that this measurement process tends to reduce maximum and increase minimum winds actually encountered. Errors in fixing the location of the balloon at a common boundary of two adjacent layers affect winds derived from these layers in the opposite sense. This is illustrated below in a simplified form.



Here the points A, B, C, D represent the true locations at unit intervals of time of a rising balloon travelling at a uniform speed and direction. B′ represents the apparent location of B in an over-simplified situation, where B′ lies on AB and all the other points are presumed to be accurately located. The wind over the layer BC will be apparently represented by B′C/unit time and be too large by an error B′B/unit time, whilst simultaneously the wind over the layer AB will be measured as AB′/unit time and be too small by the same error B′B/unit time. The errors in the two adjacent layers are negatively correlated. In practice, all the locations are likely to be more or less in error in two dimensions rather than one. This tendency for negative correlation nevertheless remains for adjacent layers which have a common boundary and affects shear winds between such layers. It does not hold, however, for adjacent layers which do not have a common boundary location, e.g. winds measured over the layers AC and BD.

Tracking the balloon is usually carried out from some location effectively centred near the balloon-launching site. Errors in wind measurements tend to increase with altitude as the range between the balloon and the observing equipment increases. This is due to the effects of geometry, sometimes coupled with a slight increase in the uncertainty of angular measurements with range. (With radar the errors of range measurement do not increase with range, so that if the track is predominantly radial the errors of wind-speed measurements in such circumstances do not increase with range.) Wind errors in general are not cumulative with altitude and, as indicated above, errors in touching layers tend to oppose one another.

The connection between the basic errors of radar equipment and errors in the computed wind is, of course, a matter of geometry. Although basic radar errors show some variation from one system to another and revised (smaller) basic errors will be appropriate as equipment is developed, it is instructive to present tables of root-mean-square vector errors of wind as functions, first of slant range and then of height and mean wind (surface to the height) for one simple set of basic radar errors. This is done in Tables 11.1 and 11.2, for basic root-mean-square radar errors of 16 yards (15 m) in range, and 0.1° in both bearing and elevation; these are values appropriate to the GL III, otherwise known as Meteorological Office wind-finding radar, Type 1. The tables are extensions of originals by Bannon.[9] They assume no correlation between individual errors of the different measured radar parameters and are for a balloon rising at 6 metres/second with winds derived from individual locations at one-minute intervals. For winds derived from the more usual two-minute intervals, errors of half those shown in these tables would be appropriate. For more modern equipment with basic errors likely to be about 70 per cent of those illustrated, errors for one-minute and two-minute winds would be about 70 per cent and 35 per cent respectively of those shown.

These root-mean-square errors refer only to the casual errors of observation. Systematic errors also occur in all the observed variables but, except in conditions of light vector mean winds when the elevation is large, these systematic errors do not affect the calculated winds to any great extent, because these winds depend predominantly on differences between consecutive observations, and changes in systematic error between such observations are small.

TABLE 11.1 Root-mean-square vector error in wind as a function of slant range and elevation to the nearest knot

	Slant range in thousands of yards											
	5	10	20	30	40	50	60	70	80	90	100	200
All elevations	Less than 1	1	2	2	3	4	4	5	6	6	7	14

TABLE 11.2 Root-mean-square vector error in wind as a function of height and mean wind to the nearest knot

Mean vector wind	Height in thousands of feet										
knots	3	10	20	30	40	50	60	70	80	90	100
10	Less than 1 knot	1	1	1	1	2	2	2	2	3	3
20		1	1	2	2	2	3	3	4	4	5
30		1	1	2	3	3	4	5	5	6	6
40		1	2	3	3	4	5	6	7	8	8
50		1	2	3	4	5	6	7	8	9	10
60		1	3	4	5	6	7	8	10	11	12
70		2	3	4	6	7	9	10	11	13	14
80		2	3	5	7	8	10	11	13	15	16
90		2	4	5	7	9	11	13	15	16	18
100		2	4	6	8	10	12	14	16	18	20

Figures to the right of and below the lines refer to slant ranges from the radar set of more than 100 000 and 200 000 yards respectively.

<u>Note</u>. These tables are derived from a radar with basic root-mean-square errors of 16 yards (≈15 metres) in range, 0.1° in bearing and 0.1° in elevation, for a free balloon rising at 1200 ft/min (≈ 6 m/s), and winds derived from individual fixes at one-minute intervals.

We now digress to mention two phenomena of background interest to analysts. As radar equipment improves, both in its accuracy of performance and its capacity for frequency of measurement (vertical resolution), one might expect the accuracy and resolution of calculated winds to increase progressively. However, the problem becomes complicated because balloons, even ascending in still air, typically exhibit lateral self-induced motions which introduce errors when the balloons are tracked as sensors of wind motion. The flows and motions depend on (i) Reynolds number (*Re*) which determines the flow regime; (ii) the relative mass (*RM*) of the sphere (balloon) to the fluid it displaces, because, for a given *Re*, the lower the *RM* the greater the lateral motions; and (iii) the sphere's rotational inertia and minute details of the surface roughness, sphericity and random orientation. In the sub-critical *Re* regime, where wake separation is laminar, the motion tends to be a fairly regular zigzag or spiral of wavelength of the order twelve times the balloon diameter. The magnitude of the lateral motion is roughly related to the factor $(1 + 2RM)^{-1}$. In the super-critical *Re* regime, where the wake separation is turbulent and the wake smaller, the motion tends to be an irregular meandering spiral. In general, *Re* for a rising radiosonde balloon decreases with increasing altitude and passes through the critical value somewhere on the ascent, usually at a little below 12 km (200 mb). Experiments with special balloons, (constructed with knobs on and known as Jimspheres), have demonstrated that the height at which the critical *Re* is encountered can be increased. However, by suitably averaging high-resolution radar data to smooth out radar-induced noise (e.g. radar data at 0.1-second intervals are averaged over 4 seconds) the periodic balloon motions can also be effectively removed. This phenomenon presents no practical difficulty with routine radiosonde wind-finding. It is quite distinct and separate from that associated with an intricate vertical layering of the air often found by research balloons in the (lower) stratosphere. In this latter phenomenon quite large measured individual speed variations tend to persist for a matter of hours, sometimes exceeding 6 hours. These variations are not necessarily turbulent and the observed wind oscillations must be associated with ageostrophic flow since the detailed shears cannot be accounted for quantitatively by the thermal wind equation.

For further detail on the motion of balloons through the atmosphere the reader may like to consult References 10 to 12.

On PILOT soundings the height of the balloon is calculated from radar range and elevation (allowing for the earth's curvature) and plotted against time; the times at which specified heights are reached serve to determine the time intervals over which the wind plot is measured. For British stations the rule (as for TEMP soundings) is that if the time at which the standard height is reached is between N.4 and N.6 minutes (where N is an integer), then the translation over a three-minute interval centred on this time is measured; for all other decimal fractions of time, e.g. N.3 etc., a two-minute interval is used. The maximum wind is measured over a one-minute interval. The specified heights are mostly standard heights selected by international or regional agreement as overall equivalents to standard pressure surfaces; they are not, in general, synoptic equivalents to standard isobaric surfaces. The difference is occasionally material when considering a mixture of TEMP and PILOT reports in a region of strong vertical wind shear.

On ocean weather ships, radar is also used for tracking but, because of the rolling platform, which is only partially stabilized, a different radar is used, and the accuracy of the angular measurements is greatly inferior to that of land-based radar already discussed. Computational techniques based upon the rather wild fluctuations of apparent rate of ascent and the likely smooth rate of ascent (also directly indicated by the radiosonde-pressure/time plot for full radiosonde soundings) are first used to reduce the errors of the elevation observations; the projection of each balloon position on the tangential plane through the ship is then plotted from these smoothed plan ranges and the corresponding azimuth observations. This plot is then subjectively smoothed to reduce the inaccuracies of the azimuth observations on the basis of the operator's experience. A correction for the earth's curvature is incorporated in the procedure. Wind reports for individual levels are influenced by observations made over layers of considerable depth (of about 10 000 ft, ≈ 3 km). As a side issue, synoptic experience of the vertical cohesion of wind reports from ships is as much a measure of the performance of the smoothing techniques employed as independent evidence of the quality of the wind observations. As a rough guide, root-mean-square vector errors of wind observations from ships are likely to be less than three times those shown in Tables 11.1 and 11.2, but the errors are likely to vary with the state of the sea and of the surface wind.

Some wind-finding systems, both on land and at sea, provide angular measurements but no slant-range observations (at least, not on all soundings). Their reduction technique is based upon the tangent relationship, heights given by the radiosonde pressure observations and observed angles of elevation. This technique is particularly vulnerable to wind-finding errors when the elevation angles are low, i.e. in situations involving strong vector mean winds. A correction for earth's curvature is usually incorporated into the computation nowadays. Some observations before 1955 did not include such corrections and, for these, high-level winds were systematically too strong.

Some wind-finding systems depend on measurements of phase difference, recorded at fixed ground antennae, of the radio wave transmitted by the sonde. The height of the balloon, either calculated from the pressure and temperature readings or estimated from the rate of ascent, together with the tangent relationship, are used in the wind determination. Many of the wind reports from stations using this equipment seem to fit into coherent synoptic patterns. However, the author has experienced some which did not and suspects that site troubles involving reflections or ghosts from geographic features such as mountain ridges or rivers, particularly those accidentally orientated along parts of ellipses with the station and the radiosonde as foci, and therefore erratic in their incidence, give rise to substantial wind errors often enough to be a nuisance.

To summarize, upper-wind observations vary in quality, their quantity decreases slowly as higher and higher levels are considered, and also when strong vector mean winds are encountered which carry the balloon beyond the range of the tracking equipment, although such failures are relatively rare except at very high levels in winter. Methods depending on the tangent relation are particularly susceptible to error in winds for which the mean vector wind from the surface to the level of interest is strong. However, many wind-finding stations mostly achieve a high standard of accuracy and their observations cohere well in space and time. From time to time wind observations are encountered which stand out from their space/time environment. Occasionally such reports can be attributed to transmission errors, sometimes to gross (e.g. 180°, 100°, 90° or some multiple of 10°) systematic direction errors of orientation, sometimes these reports bear such eccentric relationships with the winds reported by the same station at other levels, particularly at adjacent levels, that they can be attributed to poor positional fixes of the balloon track, and sometimes they arise because the wind has been attributed to the wrong level. A few cannot be explained in any of these ways.

Despite the high quality of many of the wind observations some qualifying remarks have to be made concerning thermal and vertical shear winds. First the shear wind is usually found from vector subtraction of two observed winds, each of which is reported within limitations imposed by the reporting code. Assuming no correlation between the errors of the two winds, the standard error of their vector difference is $\sqrt{2}$ times the standard error of the observed winds themselves. When the winds are sufficiently strong and similar, an accident of throwing to the nearest code number for direction can yield a shear of apparently significant strength and exaggerated cross-contour flow. A further source of difficulty arises when trying to equate shear winds to thermal winds. Thermal winds, by definition, relate to differences between geostrophic winds at different levels. Shear winds are based on actual winds which may well depart from geostrophic values at each level. These geostrophic departures are related to the trajectories and accelerations experienced by the air at the upper and lower boundaries of the layer. Sometimes these departures can be materially different and do not necessarily cancel themselves in the vector subtraction process. The practical point to remember is that shear winds derived from the observations and commonly, but wrongly, called 'thermal winds', provide a rather inferior guide to thickness gradients than do observed winds to contour gradients.

11.3.2.3 Geopotential height. In making an observation, indications of temperature, pressure, and humidity as sensed by balloon-borne radiosondes are usually recorded against time, the actual readings being in some unit such as frequency, or its inverse. These readings are converted into conventional values by means of calibration curves or tables. The process incorporates various correction procedures, usually including one (the control correction) for checking the calibrations as experienced on the ground immediately before the sounding. The temperature and humidity observations so found are then plotted against pressure interpolating through time, and geopotential differences (thicknesses) between convenient isobaric surfaces calculated for the atmosphere sampled. The essence of this calculation is that the thickness of the layer between specific (i.e. defined) isobaric surfaces is proportional to the virtual temperature of the layer when this is averaged against the logarithm of pressure. The geopotentials of defined standard isobaric surfaces are then found by adding successive thickness values to the height of the 1000-mb surface above mean sea level. This 1000-mb height is itself derived from the station surface pressure and height above MSL, with a little assistance from the surface temperature and humidity (and near-surface temperatures and humidity when the station surface pressure is above 1000 mb). Thus each standard isobaric geopotential height observation is made up of a component dependent upon the barometric pressure and air temperature and humidity measured at the surface at the launching station, together with a series of components each dependent upon the mean virtual temperature of the atmosphere (as observed) between successive defined isobaric surfaces employed in the computation. The essential features of these procedures are:

(a) Errors involved in one layer are carried upwards to all higher surfaces, so that systematic errors tend to accumulate on each ascent as geopotentials of higher and higher isobaric surfaces are determined. This effect applies to all errors which are systematic – or correlated from level to level – on single soundings, including those which may be otherwise random in their incidence from sounding to sounding at the same station. To be specific, errors arising from parallel shifts of calibration curves between radiosonde calibration and flight are corrected through the control-correction procedure, but any twists which may have occurred to the calibration curves are not, and such twists are associated with errors of radiosonde temperature, pressure, and humidity values, which are correlated from level to level on a particular flight.

(b) Any errors in the surface-pressure observation contribute directly (as a constant) to the geopotential error of all isobaric surfaces on that sounding. They also, through the control correction, affect the errors of the in-flight pressure measurements which themselves affect the thickness values indirectly, through the temperature changes with height experienced, by assigning observed temperatures to erroneous pressures.

For an isothermal atmosphere the effects of such in-flight pressure errors on the calculated geopotential are, of course, zero. But in-flight pressure errors can also affect the in-flight temperatures (and thus the geopotentials) by assigning erroneous radiation corrections to the temperature readings. This effect is usually swamped by other forms of temperature error at altitudes below 100 mb but can become important at higher levels. Radiation corrections are used only for daylight soundings made by some radiosondes.

(c) Errors in both temperature and pressure measurements may take either sign in specific circumstances and their individual effects on the calculated geopotentials are not necessarily additive. Indeed they can be more or less self-compensatory up to specific isobaric surfaces. In such circumstances the ratios of the geopotential errors at 100 mb to those at other levels on single soundings often depart substantially from more common values, even the signs of the errors may change with altitude. This limits the efficiency of practical techniques (to be described later) designed to reduce the effects of errors of reported geopotential heights on contour analysis of synoptic high-level isobaric charts.

Notes

Pressure observations are usually corrected for temperature although the correction procedure is imperfect, partly because of lag and radiation effects.

The geopotential of a feature such as the base of an inversion, as determined by a radiosonde, is significantly less accurate than the geopotential of a specific defined isobaric pressure surface of comparable altitude as determined by the same sonde.

Temperature measurements are usually corrected for lag and also for radiation effects. The radiation corrections are usually based, among other things, on a constant albedo below the sonde for all daylight soundings (either assumed or averaged for a specific location and season). For British Mk 2b sondes, radiation corrections based upon individual forecasts of albedo (to the nearest 0.2) for each flight are now used. This helps to reduce residual radiation errors, but the correct albedo cannot always be forecast. Radiation 'corrections', when used, do not necessarily eliminate radiation effects: the corrections may, in practice, over-compensate as well as under-compensate. Effects of radiation on uncorrected temperature measurements always increase as pressure decreases, although some sondes are far more susceptible to solar radiation than others. A rough guide is that, in terms of temperature, radiation effects (before correction) for the British Mk 2b sonde in the range 100 to 10 mb increase by about ⅔ each time the pressure is halved. Another guide is that the radiation effect at a level changes by about 7 per cent for each 0.1 change of albedo from 0.4 (albedo experienced may range from less than 0.1 to about 0.8). Radiation corrections vary with the elevation of the sun above the horizon. This elevation can be accurately forecast, and errors stemming from this part of the radiation-correction technique are usually small.

In general, the larger the radiation effect the greater the residual error in practical correction techniques. The American sondes, which are relatively insensitive to solar radiation and are not corrected for radiation effects, nevertheless yield routine measurements containing less radiation error than some sondes for which radiation corrections are applied in the reduction routine.

Rates of ascent of sounding balloons are usually in the range 6 to 7 $m\,s^{-1}$ and generally increase with altitude.

11.3.3 Geopotential errors in relation to associated geostrophic winds

Relative errors of geopotential for a specific isobaric surface and synoptic time, between adjacent stations, are associated with erroneous apparent geostrophic-wind components at right angles to the lines joining adjacent stations. Such components are independent of the true wind field and the isobaric level involved. They are, of course, dependent on the magnitude of the relative geopotential errors, the inter-station separation and the latitude. As an example, for a 120 n.mile interstation separation (equivalent to about 2 degrees of latitude, or roughly the distance Crawley - Hemsby) and a 60-gpm relative error between the geopotential determinations for the two stations, the associated geostrophic-wind error, in a direction normal to the interstation line, would be 46 knots at 50° latitude, about 40 knots at 60° latitude and about 60 knots at 35° latitude etc., for any isobaric level, and any actual wind field. The 60-gpm difference assumed in this example is reasonably typical of standard errors between single observations of 50-mb geopotentials for stations using the same radiosondes. Such geostrophic-wind errors, of course, are inversely proportional to the interstation differences involved. If we were to attempt contour analysis such that the contours followed the direction of the observed winds, the geostrophic-wind errors in this direction associated with the relative errors of geopotential would be inversely proportional to the magnitude of the projection of the interstation distance on a direction normal to that of the wind. Indeed, at 100 mb and above, errors in the implied wind fields associated with contour analyses which fit the reported heights are far too great to be acceptable. A compromise analysis has to be reached, aiming at contour fields associated with wind fields that correspond to the observed winds as far as possible, and utilizing to this end reported geopotential heights within their limitations of error.

11.3.3.1 Errors in reports of geopotential. Geopotential reports are subject both to errors and to mistakes. By mistakes are meant all those aberrations arising externally to the actual observation, such as mistakes in simple arithmetic, in transcription, in transmission, and in plotting. No specific rules can be laid down for dealing with mistakes, but they can often be discovered by inspection of associated so-called redundant data – for example, hydrostatic checks of layer thickness against reported temperatures. Such mistakes are found at all levels, for instance a number containing a repeated digit with the wrong digit repeated. It is necessary to eliminate mistakes by whatever means are available in the circumstances before dealing with errors. Techniques designed to reduce the effects of errors will not deal effectively with mistakes.

It is convenient to consider errors in reports of geopotential for a specific isobaric level from an individual station to be made up of the following components:

Component S: A systematic part which holds a constant value for long periods of time (years) for soundings made in darkness.

Component I: A variable part which changes systematically in an organized manner between soundings made in darkness and in daylight.

Component R: A variable part which changes in an unpredictable manner from sounding to sounding and is characterized by its standard deviation E.

Notes

A low value of E is a more desirable quality in a sounding system than low values for components S and I.

Although, in the absence of an absolute standard, it is not possible to determine absolute values of the constant component S, it is possible to determine relative values of this component between some stations, i.e. differences of S from some practical arbitrary standard. For synoptic contour-analysis purposes such relative differences contain as much useful information as would knowledge of absolute values.

The appropriate values of the component I, when added to the differences in the associated components S, yield systematic differences between soundings made in daylight (for specific isobaric levels between specific stations under specific conditions). Such 'daylight' differences, in general, change more or less steadily, depending on geographical location, and climate through the year.

The variable component R for a specific sounding is assumed to be drawn at random from a (normal) distribution at each isobaric surface. The standard deviation E of this distribution at each isobaric level is, like the other components, associated with the particular type of sonde in use, with the particular methods adopted by the authority controlling the station and with the skill of the staff involved. In other words, all error components are characteristic of the individual station and may sometimes be different even for different stations of the same meteorological service using the same type of sonde.

Uncertainties in the values adopted for the components S and I are presumed to be included in the random component R.

Although the components R are considered to be random in their incidence from sounding to sounding at an individual station, these components can be highly correlated from level to level on the same individual sounding, that is to say there is a major contribution from the so-called 'sonde error' described by Harrison[13] (see section 10.3 of Chapter 10 - Upper-air ascents). Indeed, from the nature of the geopotential computation and the manner in which errors arise, it is to be expected that, when they are not small, geopotential errors contributed from errors of pressure and temperature observations separately will both be highly organized from level to level in an individual sounding. When both these contributions have the same sign their sum will also be strongly correlated from level to level. Thus we may expect interlevel error correlations and interlevel error ratios to be more stable for large errors than for small ones.

Contributions to the geopotential errors from pressure errors are, of course, dependent upon the temperature lapse rates encountered on the particular soundings as well as on the particular pressure errors. Thus, for a specific pressure error changing relatively slowly with altitude, the contributions to the geopotential-height errors in the temperature-lapse environment of the troposphere will vanish in an isothermal environment and change sign in a thermal incline environment (that is, a very deep inversion) as that often encountered in the lower and in the high stratosphere.

Contributions to the geopotential errors from temperature errors, on the other hand, enter directly into the computation whatever the lapse rate. However, where the calibration graph or device for changing the units in which the temperature-dependent radiosonde signal is measured to conventional temperature units is approximately linear, and pre-launch ground controls have been applied, it is likely that the error ΔT (excluding contributions from radiation error) will be a function of the difference between the in-flight temperature, T, experienced and the pre-launch ground-level temperature, T_s. Specifically, for a mechanical as opposed to an electrical-type device,

$$\Delta T \approx C(T_s - T). \qquad \cdot \cdot \cdot \cdot \quad 11.3$$

These restricted conditions apply to the Meteorological Office Mk 2b radiosonde (and probably others). Harrison[13] derived errors of this sonde from twin flights. His data enable some check on the validity of equation 11.3 to be carried out by dividing his standard deviations of sonde temperature errors (obtained from about 50 twin soundings) by associated values of $T_s - T$ for each isobaric level. A constant value (related to C) should result if equation 11.3 is valid. In the absence of specific values of $T_s - T$ for Harrison's data, values of $T_s - T$ taken from the ICAN Standard Atmosphere were used. Results are shown in Table 11.3. This result is not, of course, proof that equation 11.3 is valid even for the Mk 2b sonde, let alone others: it may arise from other circumstances. Nevertheless it supports the suggestion and reasoning implicit in equation 11.3.

TABLE 11.3

Isobaric level (*mb*)	700	500	300	200	150	100	80
Test ratio	.018	.013	.013	.013	.013	.013	.012

$$\text{Test ratio} = \frac{\text{Standard deviation of sonde temperature error by night (Harrison)}}{\text{Difference in temperature between ground and level (ICAN Atmosphere)}}$$

Relative values of S for a level may be derived from mean differences (observed values minus analysed values) at each station averaged over a series of night-time charts. For this purpose the chart analyses must weigh heavily on wind observations to determine contour gradients over the area, and winds must not be too strong or too variable. Technique errors involved in this comparative process will be smallest when the wind fields are lightest and steadiest. Nevertheless, such technique errors will contribute to individual assessments. However, by averaging over a sufficient number of examples in relation to the scatter of the results, the standard error of the relative values of the S component may be reduced almost at will. Once known, relative values of the S component by definition can be applied to individual observations in future analyses, and in these analyses increased weight may be given to the geopotential reports so adjusted. The process is, to some extent, self-generating. Details of an early application of this method to the determination of relative S values were given by Hawson and Caton.[1] Nowadays, some types of objective analysis by numerical computer provide a ready source from which to monitor the relative values of S components.

The standard deviation of the individual differences between observed and analysed values at a station, used to derive relative component S, gives a measure of the standard deviation E of the random components R. This particular estimate, which for descriptive purposes we will call $\hat{E}_1$, contains a contribution from the errors of the synoptic analyses as well as from E itself. If the errors of these synoptic analyses are not locally correlated with the R components, or are negatively so correlated, then $\hat{E}_1$ will overestimate E. If, on the other hand, errors of synoptic analyses are locally positively correlated with the components R, as is more likely if the analysis gives weight to the observed geopotential values (especially if these have been adjusted for the component S), then $\hat{E}_1$ will be an underestimate of E.

Values of the component I may be derived by taking the mean value of $\overline{D}$ from a sufficiently long series of individual values of a quantity D_i, where D_i is defined as a time-centred-difference between one (the i th) observation in darkness and the average of the two observations made in daylight 12 hours before and after that time. (Differences between one day-time and two adjacent night-time soundings can be used instead of that indicated above if more convenient.) Since $\overline{D}$ may well vary with season, a series cannot be made as long as we wish in order to reduce the standard error of our estimate of I to as small as we choose. Judgement has to be used in choice of periods and a compromise reached, depending on the seasonal variation of $\overline{D}$ encountered and the scatter of the individual D_i values. $\overline{D}$ is made up of a contribution from the error component I and a contribution from the actual diurnal variation of the atmosphere. The latter is not necessarily zero, but is a function of local time. However, it is very probably quite small for differences between local noon and local midnight. For all other local times the component of $\overline{D}$ arising from the diurnal variation of the atmosphere has to be considered and removed from $\overline{D}$ to obtain a measure of the instrumental component I. The true atmospheric diurnal variation is not well established, but for middle latitudes in summer at 100 mb a linear change with time over a range of about 30 gpm between a dawn minimum and a dusk maximum is suggested. At higher levels this range will increase, perhaps to about 100 gpm at 10 mb.

Alternative estimates, which we will call $\hat{E}_2$, of the standard deviation E of the random-error components R can be made from the standard deviation σ_D of a series of values of D_i (for a single isobaric level and station). It can be shown that

$$\sigma_D^2 = \sigma_x^2 \, (3 - 4r_{12} + r_{24})/2 + 3E^2/2 \qquad \ldots \ldots \quad 11.4$$

where σ_x is the standard deviation of the true values of the geopotential,
r_{12} is the auto-correlation over 12 hours of these true values, and
r_{24} is the auto-correlation over 24 hours of these true values.

The first term on the right-hand side of equation 11.4 cannot be precisely or uniquely evaluated, but for geopotentials at levels of 100 mb and higher, especially in summer and by present-day radiosondes, the first term is usually small in relation to the second term. By neglecting the first term we can then reduce equation 11.4 to:

$$\hat{E}_2 = \sqrt{2/3}\,\sigma_D \qquad \ldots \ldots \quad 11.5$$

Subject to the usual statistical errors of sampling, $\hat{E}_2$ will, in these circumstances, be a small over-estimate of E. In the derivation of equation 11.4 one of the assumptions is that the standard deviation of the random-error components R is the same for day-time as for night-time soundings. If this is not so, $\hat{E}_2$ will tend (when 1 night and 2 day soundings are used to obtain D_i) towards the night-time value of E as well as towards the larger value.

By carrying out on a series of day-time synoptic charts a process similar to that outlined above for night-time charts to derive relative values of S and the estimate $\hat{E}_1$, a series of relative systematic errors by day (associated $S + I$ values and $\hat{E}_3$), an estimate – possibly an underestimate – of E for daylight soundings can be obtained for individual stations. This completes the circle. The strength (or weakness) of the system lies in the extent to which results from such investigations are complementary. Each part of the system has its limitations, but these limitations are interdependent so that cohesion of the whole strengthens faith in the parts and in the system. Further support is lent by Harrison's independent results derived from the twin soundings referred to earlier. These gave a value of 40 gpm for E at 100 mb for the Mk 2b sonde in the 1950s. $\hat{E}_2$ derived for Crawley from routine reports month by month from 1959 to 1961 also averaged 40 gpm. (Subsequently, $\hat{E}_2$ for Crawley has decreased a little to between 30 and 35 gpm, indicating that some improvement in performance has been achieved by various changes introduced into both the instrument and the techniques after 1961.)

By using the above techniques in research investigations, values for S, I and E can be obtained for 100 mb and higher isobaric levels for each radiosonde station. These investigations have to be repeated at intervals to update values which change as new or revised sondes and routine reduction procedures are introduced. Consequently no attempt is made to give detailed values here. However, to provide some guide on magnitudes in the early 1970s at 100 mb: (i) relative S components range over about 100 gpm, (ii) I components range from 0 (or even negative) to a few tens of gpm for most stations, but are over 150 gpm in some circumstances, (iii) E usually ranges between 20 and 50 gpm (but E can be 70 gpm or more at a few stations).

11.3.4 Contour analysis

Having first corrected 'mistakes', the next step is to remove the error component S in darkness or $S + I$ in daylight from the reported values of geopotential so that the values so adjusted will all scatter around a common (arbitrary) standard, and then to reduce the effects of the random components R by suitably averaging these 'adjusted' values from several independent stations over sub-areas which the wind observations indicate are suitable. Thus a series of 'measurements' of geopotential are obtained with materially smaller standard errors than the original reports. This is gained at the cost of larger inter-measurement spacing and therefore reduced capacity to resolve synoptic systems. Reported winds are then used further to construct local contour gradients around each 'corrected' geopotential and the whole region analysed by patching these together, having regard to the evolution of the field in space and time. In this process the 'corrected' geopotential values are not considered to be necessarily correct but to have standard errors appropriate to their own derivation, that is approximately $\hat{E}/\sqrt{n}$, where $\hat{E}$ is some representative value of the various values of E involved and n is the number of stations used in the sub-area averaging. Where it is impractical to reduce the effects of component R by suitably space-averaging over sub-areas, some reduction can be achieved by consideration of time series of reported values at the isolated stations. Construction of the analysis can often be facilitated by considering first the location of an individual contour line near the lowest latitude of interest, and then one near the highest latitudes,before constructing intermediate contours.

In practice, formal averaging of the 'adjusted' geopotential values is not usually specifically carried out, but the 'adjusted' values and the associated standard deviations of their random components (values of E) are considered in bulk. The analyst aims to construct gradients compatible with the observed winds which fit the 'adjusted' geopotential values with roughly the appropriate characteristics of a normal distribution with appropriate E. The main points to remember are that at 100 mb (after adjusting for the systematic parts of the errors) with present-day sondes, only about a quarter of the geopotentials can be expected to have values within ± 10 gpm of the analysis ($\pm E/3$ for many sondes). Many should have only reasonably small (say ± 30 gpm) differences from the analysis and only about a twentieth should exhibit differences from the analysis in excess of $\pm 2E$. The analyst has to resist the temptation to construct so that too many of the adjusted observations fall into either the $E/3$ or the $2E$ categories relative to the analysis.

The above procedure can be adopted at any of the high levels where the data are sufficiently plentiful and winds sufficiently steady. Currently it is used at 100 mb where, in middle latitudes, winds are usually relatively light and approaching a level of minimum variability in both space and time. It could, with the data density now available, be carried out at 50 mb and there would be some advantage in carrying it out at both 100 and 50 mb.

An alternative procedure for correcting 'the reported observations of geopotential' is, however, available and currently used for other isobaric levels once an analysis of adequate quality is available from one of the high levels. This alternative procedure 'corrects' each reported value in one step for the effects on each individual sounding of the three error components S, I and R by using standardized multiples for each level of the difference A between reported and analysed values at the high level already analysed. For use with values of A derived for 100 mb in middle latitudes the standardized multiples are 125 per cent for 50 mb and 145 per cent for 30 mb. Although, for practical purposes, a constant percentage of A (125 per cent etc.) is used to 'correct' each associated 50-mb etc. level, as explained earlier, it is possible for the ratios of the real instrumental errors at 100 and 50 mb etc. to assume a wide range of values and even to change sign. Fortunately this range of uncertainty decreases very materially as A itself becomes large. Consequently the numerical size of the uncertainty with which the 50-mb corrections are deduced by the expedient of using a constant percentage of A does not vary as much as might be feared. For example, if $A = 20$ gpm, the standardized correction of 125 per cent A for 50 mb might itself be 100 per cent A, i.e. 20 gpm, in error. On the other hand, if $A = 200$ gpm, the standardized correction of 125 per cent A for 50 mb is itself likely to be less than 10 per cent A, i.e. also 20 gpm in error.

Errors in the analysis at 100 mb affect the A values themselves and these errors are, of course, carried into the corrections deduced for higher levels. However, provided the gradients on the basic chart accord with the winds, such errors tend to change only slowly with distance, so the gradients on the upper chart are little affected. The effects of this type of error on the correction are more troublesome when considering changes with time involving a series of charts.

In principle, if on individual soundings independent measurements of A could be made at two levels with sufficient accuracy, for example at 100 mb and at 50 mb, individual 'corrections' could be inferred for other levels, which should be more effective than those derived using standardized multiples of A from a single level. However, the method based upon standardized multiples of A at 100 mb works reasonably up to about 30 or 20 mb. Above that level, because of the usually increased thermal incline and poor correlation between the pressure errors in the troposphere and those above 30 mb on individual soundings, the method is often unsatisfactory and is not recommended. Even at 50 and 30 mb it sometimes yields unsatisfactory results. This becomes apparent from the distribution between different stations of the 'corrections' over sections of the chart and can usually be traced to erroneous analysis at the basic high level (100 mb), or to an abnormal environment.

Reported geopotentials 'corrected' by this alternative procedure based upon standardized multiples of A are not of course error-free. In general, although the original error content of the reported values is substantially reduced, residual errors remain. Where these residual errors are randomly distributed between stations in sufficiently dense networks they can be further reduced by suitably averaging for several stations over sub-areas or time series (as was done in the basic system of analysis to reduce the effects of the original components R). Reported winds and 'corrected' geopotential values, having regard to the evolution of the field in space and time at a given level and the next lower one, are then used to construct contours over the whole region at the higher levels. In this process, formal partial thickness and upward-gridding techniques, formerly used in the troposphere, are not generally employed but can make useful contributions, particularly in data-sparse areas. This upward-gridding technique was particularly useful in the troposphere when upper-air data were sparse, but is prone to error in layers when the wind speed is decreasing with altitude so that the resulting 'grid field' is a residual between two opposing and generally larger fields, as for the 200 to 100-mb levels. At levels above 50 mb the wind speed usually again increases with altitude and the data become sparse. In such circumstances, gridding technique recovers its utility.

Above 30 mb the efficiency of the procedure for 'correcting' the reported values by standardized multiples of A_1 at 100 mb usually deteriorates substantially, and it is better to use differences A_3 between reported and analysed values at 30 mb in attempts to 'correct' any reported or extrapolated geopotentials for 10 mb. It is difficult to improve on simple corrections of A_3 to associated 10-mb reported heights. Because of the very sparse amount of either wind or geopotential data at 10 mb, and the errors to which these are subject even after 'correction', the analyst has to rely heavily on continuity in space and time. A de-grid of the apparent partial thickness field between the already analysed 30-mb contours and the 10-mb contours under construction can help in some circumstances. In considering continuity in time, contributions from the real diurnal variation of the atmosphere, as well as from the diurnal variation of the error components I, are material (i.e. $\bar{D}$ rather than I alone is relevant).

For an account of the climatology of the stratosphere and background information of use to analysts, the reader is referred to section 8.3 (page 11) of Chapter 8.

BIBLIOGRAPHY

1. HAWSON, C.L. and CATON, P.G.F.; A synoptic method for the international comparison of geo-potential observations. Met Mag, London, 90, (Dec.) 1961, pp. 336-344.

2. BOYDEN, C.J.; A method of fitting isobaric contours to the gradient wind. Met Mag, London, 89, (March) 1960, pp. 68-71.

3. World Meteorological Organization. The use of satellite pictures in weather analysis. Tech Notes, Wld Met Org, Geneva, No. 124, 1973.

4. SINGLETON, F.; Human intervention in the operational objective analysis. Met Mag, London, 104, (Nov.) 1975, pp. 323-330.

5. JOHNSON, D.H.; A further study of the upper westerlies; the structure of the wind field in the eastern North Atlantic and western Europe in January 1950. Q J R Met Soc, London, 79, 1953, pp. 402-407.

6. ENDLICH, R.M. and McLEAN, G.S.; The structure of the jet-stream core. J Met, Lancaster, Pa, 14, 1957, pp. 543-552.

7. SIMLA, J; The relation of the jet-stream axis to contour direction. Toronto, Met Br, Circ No. 3018, 1958.

8. KOCHANSKI, A; Atmospheric phenomena in the height region from 70 to 160 km. The circulation of the stratosphere, mesosphere and lower thermosphere, Chapter VII. Tech Notes, Wld Met Org, Geneva, No. 70, 1965, pp. 140-169.

9. BANNON, J.K.; Errors in winds measured with GLIII radar equipment. Met Res Pap, London, No. 406, 1948.

10. MacCREADY, P.B., Jr.; Comparison of some balloon techniques. J Appl Met, Boston, Mass., 4, 1965, pp. 504-508.

11. Boston, Mass., American Meteorological Society. Symposium on rising superpressure balloons. J Appl Met, Boston, Mass., 4, 1965, pp. 130-148.

12. SAWYER, J.S.; Quasi-periodic wind variations with height in the lower stratosphere. Q J R Met Soc, London, 87, 1961, pp. 24-33.

13. HARRISON, D.N.; The errors of the Meteorological Office radiosonde, Mark 2B. Scient Pap, Met Off, London, No. 15, 1962.