Handbook of Weather Forecasting

CHAPTER 8

JET STREAMS, TROPOPAUSE AND LOWER STRATOSPHERE

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CHAPTER 8

JET STREAMS, TROPOPAUSE AND LOWER STRATOSPHERE

8.1 JET STREAMS

In Chapter 1 - Introduction - the two principal jet-stream systems of the troposphere were discussed briefly. The subtropical jet (STJ) is a result of the conservation of absolute angular momentum as air, having ascended near the thermal equator, moves polewards in the upper branch of the Hadley cell; it is a strong and fairly steady westerly current at about 150-200 millibars, more or less above the subtropical high-pressure cells. The STJ is associated with the break between the equatorial tropopause at about 100 millibars and the temperate-latitude tropopause at about 200-250 millibars.

The polar-front jet stream (PFJ) is very different; it exists as a result of the strong horizontal temperature contrasts built up through a substantial depth of the troposphere by the confluence of warm and cold air masses. The PFJ is constantly changing, each element lasting usually for only a few days. It usually blows from the western half of the compass, but at times may have an easterly component over part of its length. The axis of strongest winds is generally in the upper troposphere, in the region of 300 millibars. The jet stream is important, not only for aircraft in flight but also for its dynamical effects. The creation and maintenance of a band of strong winds implies marked acceleration and deceleration of the airflow, particularly at the entrance and exit of the jet stream. The accelerations lead to ageostrophic components of flow and associated vertical circulations which are responsible for the development of synoptic-scale features and which largely determine the activity of the frontal systems. The following account will refer almost entirely to the PFJ and will be divided into three parts, dealing with its structure, with the related distributions of cloud and humidity, and with the dynamical aspects of the jet stream and its relationship with surface features.

8.1.1 Structure of the polar-front jet stream

The average potential temperature and wind speed around the jet core, based on observations during 1959-61 by the Meteorological Research Flight,[1] are shown in Figure 1.

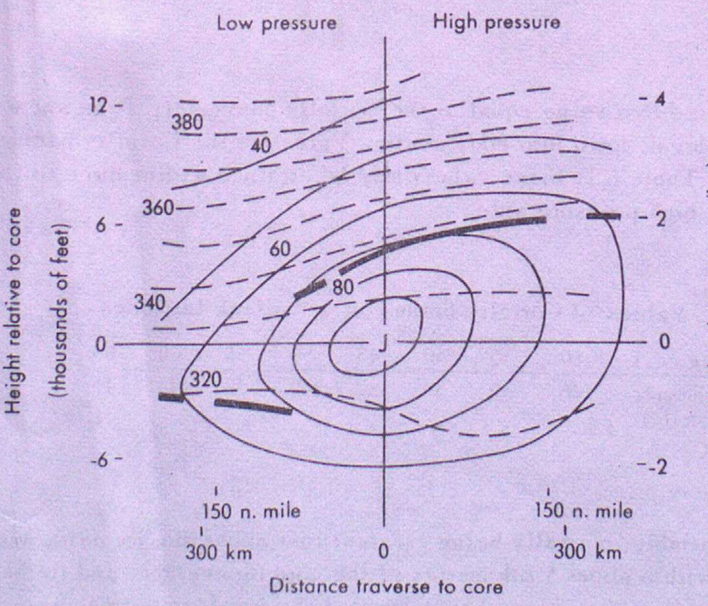

FIGURE 1. Average potential temperature and normalized wind speeds around the jet core

isotachs as percentage of core speed
potential temperature isotherms (K)
tropopause

The results are similar to those obtained by Endlich and McLean[2] for Project Jet Stream.

The axis of the jet stream is in the warm air, just below the tropopause, and above the surface cold air. It is often found vertically above the position of the front at about 400 to 500 millibars. On the low-pressure side of the jet core, horizontal wind shears are generally greater than on the high-pressure side; typical values are 0.4 hour $^{-1}$ (40 knots in 100 n. mile) on the low-pressure side and 0.25 hour $^{-1}$ on the high-pressure side. The magnitudes of the shears vary considerably from one jet to another, and very much greater shears can exist over shorter distances. Crossley[3] has estimated the extreme cyclonic shear to be expected and his results are shown in Figure 2.

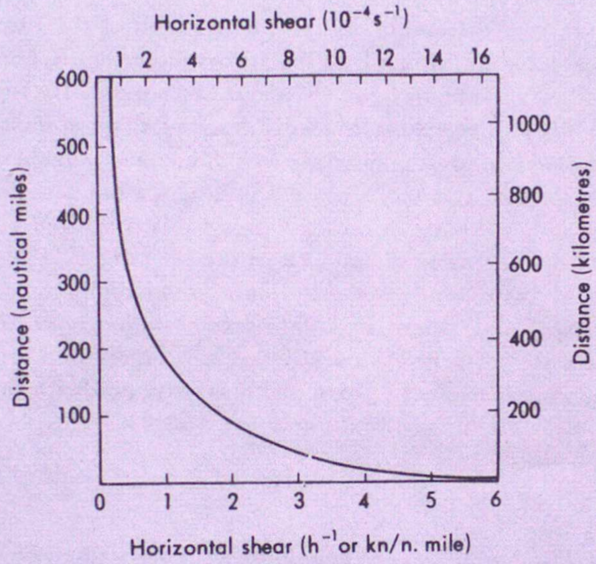

FIGURE 2. Curve of estimated extreme horizontal cyclonic shear

The anticyclonic shear is limited to a value equal to the Coriolis parameter; if the shear is greater than this the flow is likely to break down into eddy motion. Values of the Coriolis parameter for a range of latitudes are given in Table 8.1; these values may be attained within three to five degrees of latitude of the jet axis, on the high-pressure side.

TABLE 8.1 Values of Coriolis parameter at various latitudes

Latitude (degrees)	30	35	40	45	50	55	60	65
Coriolis parameter (kn/100 n. mile)	26	30	34	37	40	43	45	47

The vertical shear is very variable, normally being greatest just above the jet axis, with the speed falling to half its peak value within about 5 kilometres of the core on average, and in the troposphere on the low-pressure side of the axis. Crossley[3] has estimated the frequency of occurrence of various values of vertical shear and also the expected extreme shear: the results are shown in Figure 3.

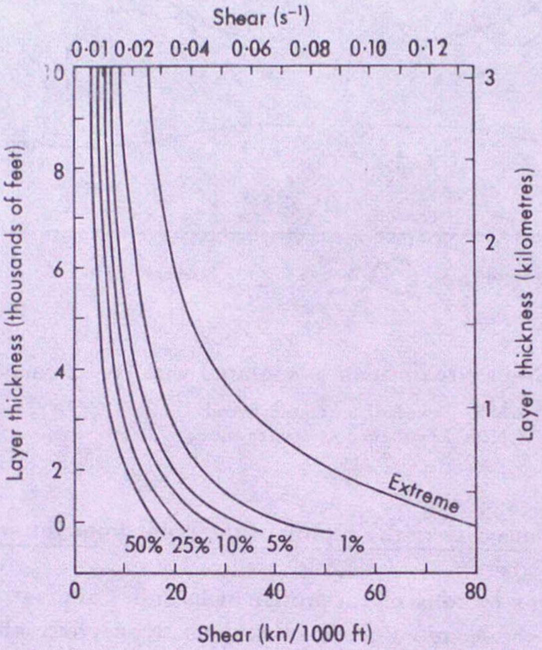

FIGURE 3. Estimated frequencies and extremes of vertical wind shear

Jet streams also exhibit fluctuations of speed along the axis, zones in which the wind speed reaches a maximum alternating with regions where the speed is appreciably lower, sometimes below jet-stream speed. The zones of highest wind speeds have sometimes been called 'jet streaks'. Newton[4] has shown that the length of a jet streak, defined as the distance along the jet axis between the minima on either side of the velocity maximum, is proportional to the wind speed in the jet streak. The patterns move downstream at speeds much less than those which occur in the strong-wind zones, so that the air moves through the pattern, alternately accelerating and decelerating. According to Newton, the variations are a result of inertial oscillations about the gradient wind, the strongest winds being 'super-gradient' and the minima being 'sub-gradient'.

The acceleration which air undergoes as it approaches a region of maximum speed gives rise to an ageostrophic component of motion to the left looking downstream, that is across the contours towards the lower contour heights. Deceleration results in an ageostrophic component to the right, towards higher contour heights. These components of the wind transverse to the axis at the entrances and exits of jet streams have been demonstrated by Murray and Daniels[5] by direct analysis of wind observations; the components averaged about 10 knots (5 metres/second). The transverse components would be consistent with the vertical circulations shown in Figure 4. Indirect support for the vertical motions in the upper troposphere is found in the distributions of humidity, clouds and rainfall associated with the jet stream. The vertical circulations are also consistent with the surface development areas shown in Figure 4 (A for an anticyclonic development area, C for a cyclonic development area), and in fact are an essential dynamical link in their formation.

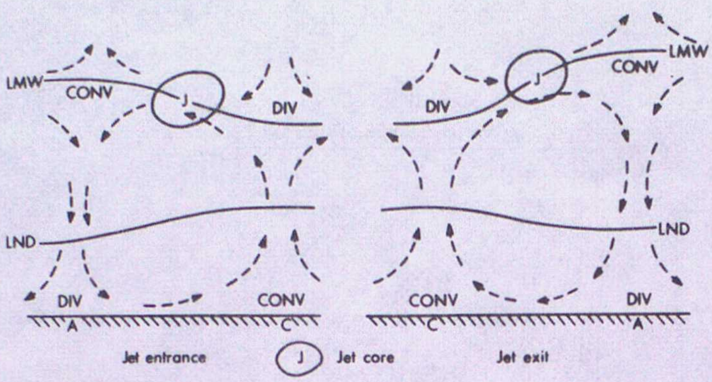

FIGURE 4. Indirect circulations associated with jet streams

LMW Level of maximum wind
LND Level of non-divergence

8.1.2 Distributions of cloud and humidity in the region of the polar-front jet stream

Murray[6] used aircraft observations to construct a profile of humidity across the jet stream. The distributions were very variable, but the mean values for the upper troposphere showed higher humidities on the high-pressure side of the jet stream than on the low-pressure side. Briggs and Roach[1] also studied the humidity distribution about the jet-stream core; the mean profile, for the same flights as in Figure 1, is shown in Figure 5. The main difference is between the tropospheric air on the high-pressure side and the stratospheric air on the low-pressure side.

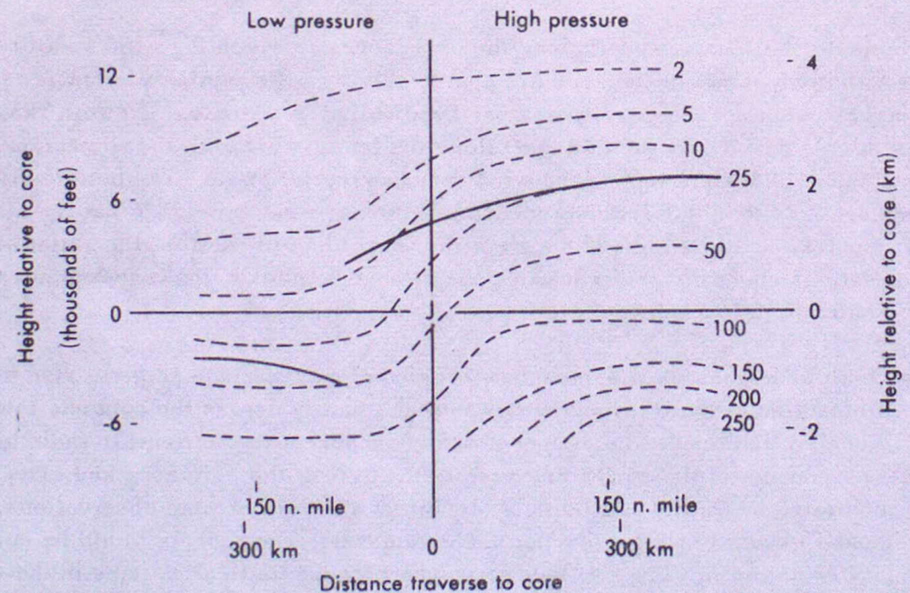

FIGURE 5. Average distribution of water vapour around the jet core

- - - - humidity mixing ratio
——— tropopause

Early investigations[6,7] of the clouds associated with the polar-front jet streams showed a marked predominance of clouds formed by convection on the left of the jet axis (looking downstream), and of layer clouds on the right on the axis. Murray[6] concluded from his studies that cloud above 400 millibars was generally found between 150 and 450 n. mile from the axis on the high-pressure side, but not beyond 100 n. mile from the axis on the low-pressure side, except occasionally for cumulonimbus tops or anvils. Medium cloud was much more common on the high-pressure side than on the low-pressure side, but low cloud was equally likely to be found on either side. Within 100-150 n. mile of the axis, amounts of high and medium cloud were very variable; extensive sheets of cirrus were generally found near jet streams associated with surface fronts.

These findings have been spectacularly confirmed by satellite pictures.[8] The poleward boundary of the high cloud is frequently very sharp, lying near or somewhat equatorward of the jet axis, and often casts a distinct shadow on the lower clouds or on the surface. Most commonly the edge of the cirrus shield is associated with the polar front and the jet stream on the eastern side of an upper trough, and the cirrus begins to dissipate as it is carried across the upper ridge line to the east. At times the cirrus near the edge may have transverse bands within it, in the region of strong horizontal shear. When the jet stream crosses an occlusion, the cirrus may or may not cast a shadow on the lower clouds to poleward, but the lower clouds will often show a lumpy texture resulting from the unevenness of the cloud tops.

Cirrus occurs much less frequently on the western side of an upper trough, and here the most common cloud pattern is the transition from open cellular cloud on the low-pressure, cyclonic side of the jet, to closed cells on the high-pressure side. The reader should consult WMO Technical Note No. 124[8] for excellent examples of satellite photographs showing jet-stream clouds.

8.1.3 Synoptic features associated with the jet stream

Although there is no unique correspondence between mid-latitude jet streams and surface synoptic features, the relationship between a jet stream and the associated synoptic systems often presents a coherent picture over a day or more, and the development of surface features is intimately related to the dynamical characteristics of the jet and to the events which bring about the formation of the jet in the first place. The variations in the individual relationships with time often occur in a systematic and relatively steady manner as the synoptic situation evolves, and there are some general statements which can be made and which should be of value to the forecaster.

The most common association is between the development of a family of depressions and the jet stream on the east side of an upper trough, as illustrated in Figure 2 (page 4) of Chapter 6 - Depressions and related features. The primary depression often lies nearly under the left exit of the jet stream, developing and perhaps moving with the upper pattern as the jet extends forwards or as the trough progresses. Once out of the left exit region, for instance as when the circulation of the low extends through most of the troposphere, development generally ceases and the low begins to fill. On the other hand, the right entrance region or, as in Figure 2 of Chapter 6, the region ahead of the upper trough and to the right of the jet-stream axis, is also a favoured place for development. A wave forming in this area is likely to lead to an intensifying low moving quickly through the pattern, across the jet axis, to replace or rejuvenate the old primary low.

Boyden[9] has summarized some of the relationships in the model shown in Figure 6. He came to the following conclusions about the average relationship between the jet stream and the surface fronts:

(i) The jet normally passes over tip of warm sector A in a direction roughly parallel to the warm-sector isobars near A.

(ii) The warm-front jet normally passes through a point G about 120 n. mile ahead of the surface ridge where this latter is farthest from the front (i.e. at F); G will normally be 300 to 450 n. mile from the front itself, the larger distances usually occurring when the front is slow moving.

(iii) The distance of the jet from the warm front remains roughly constant towards lower pressures until the point C, opposite B, where AB equals 7° latitude.

(iv) The distance of the jet from the warm front does not decrease towards higher pressures, i.e. DE $\geqslant$ BC.

(v) On average, cold-front jets diverge from the front by 140 n. mile for each 600 n. mile along the front.

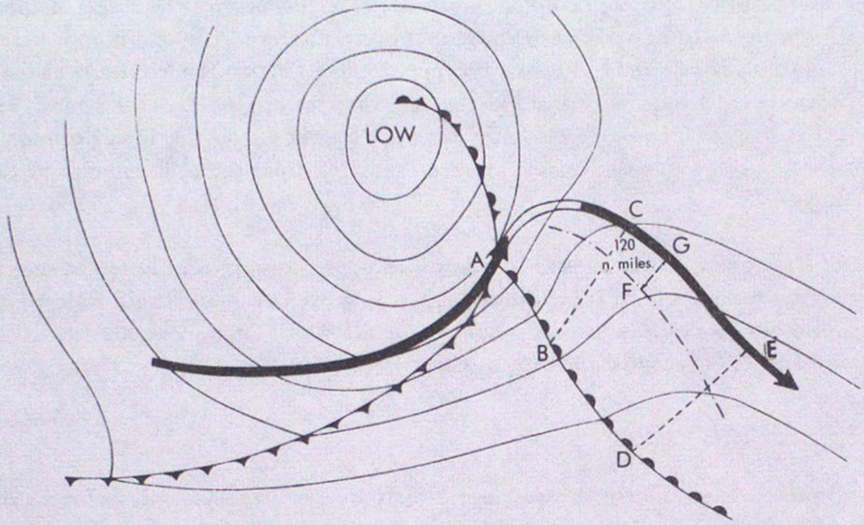

FIGURE 6. Jet stream in relation to surface fronts
—.—.— axis of surface ridge associated with warm front.
The broad arrow represents the 300-mb jet, the unshaded portion being weak or broken.

Boyden[9,10] also studied the relationship between the jet stream and the flow at other levels. For the strongest jets, the flow at 900 metres was usually greater than 30 knots and had a direction within about 20° of that of the 300-millibar wind. A reasonably close relationship between the position of the axis of the jet stream and the position of the peak 1000-500-mb thickness gradient was found, the former being on average 60 n. mile to the cold side of the latter and less than 100 n. mile on 70 per cent of occasions. The maximum wind speed at 300 millibars was roughly twice that of the maximum 1000-500-mb thermal wind speed.

Normally the jet axis is curved anticyclonically relative to the 300-millibar contours, as in Figure 7, and, in general, the amplitude of the jet stream in the long waves is greater than that of the contours.

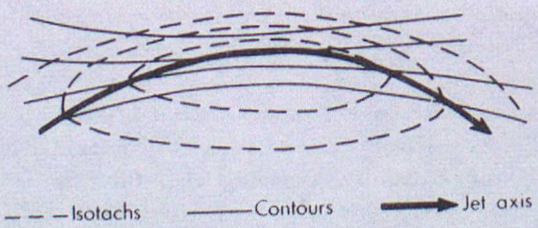

FIGURE 7. Position of jet axis relative to contours

A comprehensive account of jet streams has been given by Reiter,[11] and there is an excellent summary of many of the main features in Palmén and Newton.[12]

8.2 THE TROPOPAUSE

The formal definition of the tropopause given by the World Meteorological Organization[13] is as follows:

(i) The first tropopause is defined as the lowest level at which the lapse rate decreases to 2 degC/km or less, provided also that the average lapse rate between this level and all higher levels within 2 km does not exceed 2 degC/km.

(ii) If above the first tropopause the average lapse rate between any level and all higher levels within 1 km exceeds 3 degC/km, then a second tropopause is defined by the same criterion as under (i). This tropopause may be either within or above the 1-km layer.

Note

A level otherwise satisfying the tropopause definition but occurring at a height below that of the 500-mb level will not be designated a tropopause unless it is the only level satisfying the definition and the average lapse rate fails to exceed 3 degC/km over at least 1 km in any higher layer.

The tropopause is the boundary or narrow zone of transition between the troposphere, the region characterized by marked vertical mixing on a wide range of scales, and the stratosphere, in which vertical mixing plays a very much more restricted role (see 8.3, page 11). It is characterized by a change in lapse rate from, on average, somewhere near the saturated adiabatic value in the troposphere below, and a much smaller lapse rate, an isothermal layer, or an inversion in the stratosphere above. The tropopause is recognizable as a surface or zone which is almost continuous over substantial areas of dimensions of hundreds or thousands of kilometres. However, breaks or discontinuities do occur. The tropopause is seldom level and sometimes slopes quite sharply in the vicinity of active synoptic systems. On some occasions it is possible to recognize more than one tropopause in the vertical.

It is not surprising that there is a zone of minimum temperature somewhere between two major sources of atmospheric heating – at the surface and in the ozone layer between 35 and 70 kilometres. The general form of the temperature profile may be calculated from consideration of the radiative properties of the atmosphere, with adjustment for convective motions in the troposphere where the radiative equilibrium calculations yield an unstable lapse rate (see, for example, Murgatroyd[14]). It is less easy to explain the sharpness of the discontinuity of temperature gradient, or why the tropopause appears to behave for some of the time almost as a material surface. Dynamical processes are likely to be involved, with synoptic-scale vertical motion in the lower stratosphere playing an important part.

On the global scale it is important to recognize two tropopauses, the tropical and the polar. The tropical tropopause is associated with tropical air and is usually the sole tropopause between the equator and about latitude 30°. This latitude is subject to seasonal, geographical and day-to-day variations. The tropical tropopause is high (pressure about 100 millibars) and cold (temperatures about 190-200 K) in the tropics.

Polewards of about latitude 50° there is commonly only the polar tropopause which is both lower and warmer than the tropical tropopause. In general, pressures and temperatures at the tropopause increase towards the poles. In polar regions there is often a very low and relatively warm tropopause associated with the very cold tropospheric air near the poles. Schumacher[15] has, however, remarked that over the Antarctic some extremely low temperatures were measured in the stratosphere in winter, associated with the apparent disappearance of a well-defined tropopause.

Over the zones of both hemispheres between the parallels of about 30° and 45°, both tropical and polar tropopauses occur on more than about 10 per cent of occasions. There are both seasonal and geographical variations. In some localities in some seasons of the year the double tropopause may be by far the most common occurrence. In another season it may be rare and, at another place near the same latitude, occur less than half the time. For example, at Nicosia during 1950-51 two tropopauses occurred on 91 per cent and 95 per cent of occasions in January and April but on only 2 per cent of occasions in July. For the same period and months the figures for Gibraltar were 48 per cent, 43 per cent and 15 per cent respectively.

Over the British Isles the mean pressures and temperatures at the tropopause are normally between about 220 and 280 millibars and 210 and 220 K. There are wide variations beyond these limits which are usually readily recognizable on the synoptic scale from routine data. When the troposphere over the British Isles is composed of very warm air, both pressure and temperature will be substantially lower. Conversely, in a marked outbreak of very cold air from the north, pressure and temperature at the tropopause are higher. More detailed data covering temperatures and pressures at the tropopause, as deduced from upper-air soundings, may be obtained by consulting the various parts of *Upper air data for stations maintained by the Meteorological Office*.[16] The statistics and graphs are not reproduced in this handbook.

The rest of this subsection will be of a qualitative nature describing some features of the tropopause which should be known by forecasters.

Sawyer[17] has conducted a detailed synoptic analysis of the topography of the tropopause for the month of May 1949. From the synoptic and analytical viewpoint an important conclusion is that, at most times, the tropopause may be regarded as a material surface moving with the air, but that at infrequent intervals the tropopause dissipates at one level and re-forms at another. It was also found that the tropopause usually retains its potential temperature within the limits of measurement (about 2-3 °C) over 24 hours. Accordingly, over such periods, behaviour of the air near the tropopause level may be treated as adiabatic.

Although the tropopause may be regarded as a material surface it does not always have a simple form. Several authors have noted and described 'folds' in the tropopause. These folds are usually observed in association with fronts and the simplest fold is shown schematically in Figure 8. A sounding at point D would probably show a double tropopause.

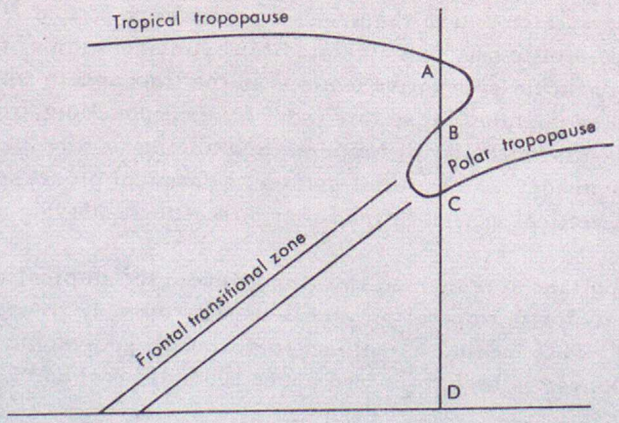

FIGURE 8. Schematic cross-section through a warm front showing a folded tropopause

Sawyer found no fewer than 12 distinct examples of folds in the tropopause charts for May 1949 in association with warm-front systems, and stated that a folded tropopause occurred with all major warm-front systems for which adequate observational coverage was available during this period, although the folded structure did not necessarily extend throughout the length of the front. The relationship of the fold or discontinuity of the tropopause with the surface system is shown schematically in Figure 9. These diagrams bring out only significant features as the details varied considerably from case to case.

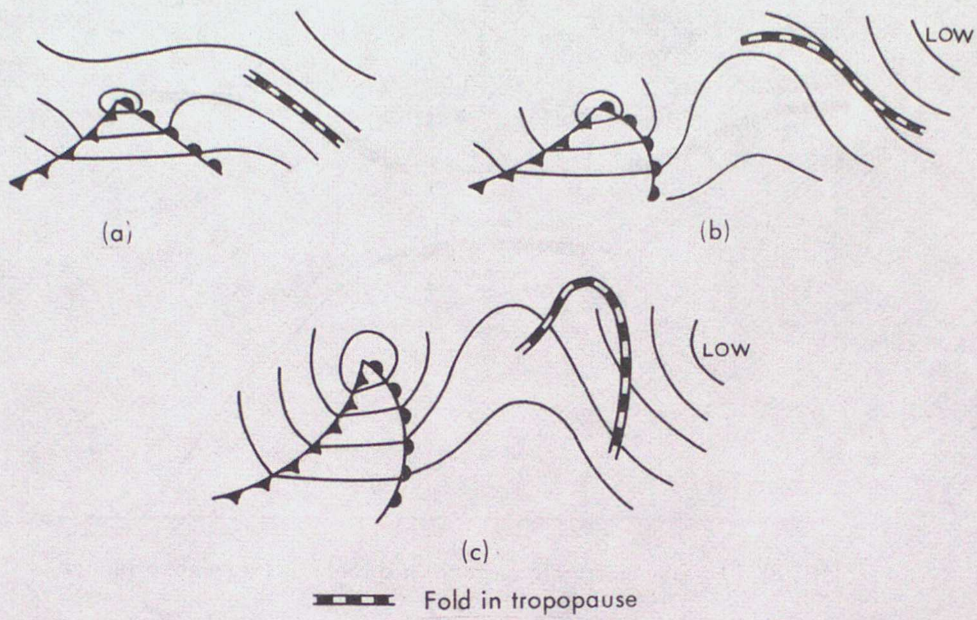

FIGURE 9. Schematic relation of the folded tropopause to the frontal system in the horizontal

Sawyer found that the first formation of the folded tropopause appeared to take place some 800-1500 kilometres ahead of the surface warm front and to be well to the right of the track of the depression centre and that subsequently the folded structure might become more extensive, tending to conform to the shape of the front. The fold was found to be above a variety of pressure systems and was usually east of the axis of the ridge of surface high pressure. Sawyer concluded that it was not possible to identify its position by any feature of the surface chart. He did find that there was a close association between the folded tropopause and the jet stream; all the cases noted occurred in association with the jet stream and most of them first appeared near the axis of the jet, with a slight tendency for the discontinuity to be a little to the left of the jet axis. Various authors have suggested or deduced that the entrance to a jet stream should have a transverse circulation (see 8.1 on page 1).

Discontinuities exist in the tropopause surface and are particularly well marked with very low tropopauses. To describe this phenomenon Palmén[18] has coined the expression 'tropopause funnel'. A tropopause funnel is shown schematically in Figure 10.

The centres of low tropopause shown centrally in Figure 10 develop frequently in association with vigorously deepening depressions and subsequently become centred over the old depression when it has become slow moving and has started to fill. Sawyer[17] found 12 centres of low tropopause during May 1949, all being persistent features with a life of from two to seven and a half days. The majority formed in association with active but rather slow-moving depressions and Sawyer considered it probable that depressions developing close to the tip of a cold trough or to the left of a diffluence in the thermal pattern were particularly likely to lead to the development of a low tropopause. The formation of tropopause funnels seems to be closely associated with the flow pattern near the tropopause level. Nine of the 12 occurrences in May 1949 were associated with well-defined troughs at 300 millibars which, in several instances, subsequently developed closed circulations; the remaining examples were associated with a closed circulation at 300 millibars from the earliest traceable stage. An important feature of a tropopause funnel is the very low tropopause in the central region.

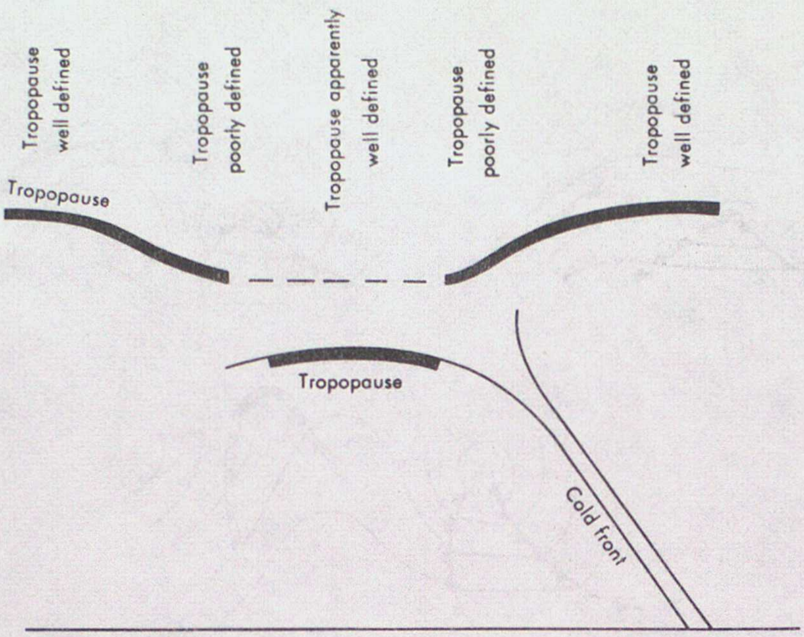

FIGURE 10. Schematic cross-section through a tropopause funnel

The tropopause is usually well below the 300-millibar level and sometimes below 400 millibars. The potential temperature at this low tropopause in the neighbourhood of the British Isles is usually between about 27 and 35 °C which is often well below the potential temperature at the tropopause in any of the air masses entering into the original circulation. The low central tropopause in many cases appears to be continuous with the lower boundary of an adjacent cold front. Sawyer found that the tropopause is well defined within the central core but it is often difficult to trace on its fringes; this is consistent with the discontinuous structure shown schematically in Figure 10.

Tropopause funnels with closed wind circulations seem to be dynamically stable. Their decay is often associated with the overrunning of a higher tropopause associated with the approach of a vigorous warm-frontal system. The decay is then rapid and the tropopause funnel may effectively disappear in 12 to 24 hours. Sawyer believed that the decaying process was associated with the rapid descent of air in the tropopause funnel and its ultimate absorption into the troposphere.

After excluding changes in tropopause levels due to folded tropopauses or tropopause funnels, Sawyer noted several occasions in May 1949 in which the tropopause appeared to have been replaced simultaneously over a fairly wide area by a new tropopause at a different level. On most of the occasions the displacement was downwards. The changes in the tropopause level were not large; 20 to 40 millibars was common with a corresponding change of potential temperature of about 6-11 degC. Sawyer was unable to prove from his data whether the change was continuous or discontinuous but he considered that a discontinuity of lapse rate developed at a new level and subsequently became the dominant discontinuity and that the potential temperature at both levels was approximately conserved during the process. He found that no feature of the surface synoptic chart could be connected directly with such developments but several of the cases occurred on the fringe of anticyclones.

One further aspect of Sawyer's work is important for forecasters. From an analysis of trajectories Sawyer concluded that advection and vertical motion exerted about equal influence on the height of the tropopause. It is important to bear this in mind in assessing variations in the height of the tropopause.

In his investigation of jet streams Murray[6] mentioned that the tropopause tended to be continuous with the weaker jet streams and to be uncertain or discontinuous with the more intense ones.

8.3 THE LOWER STRATOSPHERE

The vertical temperature structure of the atmosphere up to 80 kilometres is illustrated in Figure 11: the values are those of the International Standard Atmosphere,[19] representing roughly the mean atmospheric conditions over the whole earth.*

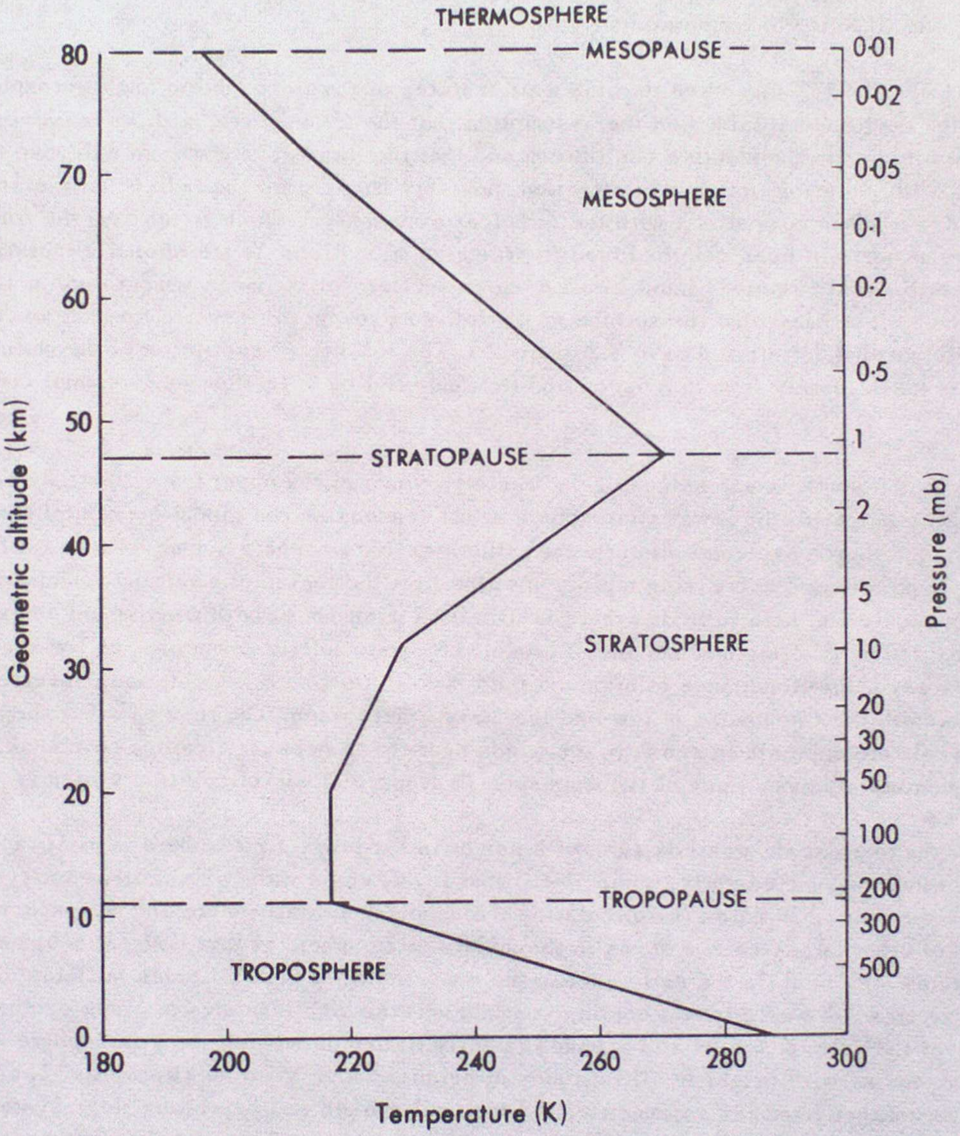

FIGURE 11. Vertical temperature profile of the International Standard Atmosphere

The major features of the mean temperature profile are determined largely by the absorption of solar radiation by various constituents of the atmosphere. Above about 100 kilometres there is strong absorption of ultra-violet radiation of wavelength less than 0.2μm by molecular oxygen. There is also radiative cooling in the infra-red portion of the spectrum, but the balance is such that temperatures are high in this region which is known as the 'thermosphere'. At 80 to 85 kilometres, on the other hand, absorption is

* Identical with the ICAO Standard Atmosphere to 32 kilometres. The ISA has been adopted by WMO and ICAO.

weak, and temperatures are low, but the temperature begins to rise with decreasing height as absorption by ultra-violet radiation of rather longer wavelengths by ozone takes place. A region of maximum temperature - the 'stratopause' - is reached at about 50 kilometres, where temperatures are, on average, around 0 °C. Between the stratopause and mesopause lies a region of temperature lapse, the 'mesosphere'. The stratosphere is the region between the tropopause and the stratopause; temperature generally increases with height, the rate of increase being greater in the upper stratosphere than in the lower layers, where there is at times a small lapse of temperature.

As long ago as 1909, Gold[20] suggested that the main features of the tropospheric and stratospheric temperature profiles could be explained on the assumption that the troposphere, with its main heat source at the surface, was in broadly convective equilibrium and that the stratosphere was in radiative equilibrium. This idea, while pointing in the right direction, possibly encouraged the belief that the stratosphere was quiet and undisturbed in comparison with the turbulent troposphere, which is far from the truth. Inequalities in the amounts of heat absorbed lead to strong zonal motions, to meridional circulations and to large-scale disturbances. Figures 12 and 13 show cross-sections of the mean temperature and zonal wind fields and the variabilities from the surface to the lower mesosphere: they are from the excellent review paper by Murgatroyd,[14] referred to in 8.2 (page 7). The following description of the main stratospheric features is taken largely from that paper, and from material by C.L. Hawson (personal communication).

In middle latitudes the upwards extensions of the westerly winds of the upper troposphere decrease in intensity rapidly with height in the lower stratosphere at all seasons as the global horizontal thermal gradient reverses. For short wavelength disturbances, the lower stratosphere seems to act as a filter, the amplitudes of the wind systems decreasing rapidly upwards from the region of maximum variability near the tropopause. In middle and high latitudes there is usually a minimum of both wind speed and variability at about 70 millibars. The filtering mechanism is a result of 'stratospheric compensation': where the tropospheric air is warm, the tropopause is high and the lower stratosphere is cold, and conversely, where the troposphere is cold, the tropopause is low and the stratosphere warm. The result is that the stratospheric thermal winds are opposite in sense to the winds near the tropopause, leading to a weakening with height and eventual disappearance of the tropospheric synoptic-scale circulation systems.

The decrease of the large-scale westerly flow with height in the lower stratosphere is most rapid in summer when the winds become easterly by the 50-millibar level, and a warm symmetrical anticyclonic vortex, centred on the pole, dominates the circulation throughout the stratosphere and mesosphere. There is little evidence of other large-scale systems in the middle stratosphere at this time, although mesoscale wind variations are found. In the early autumn the temperature at high latitudes falls rapidly, and the wind flow becomes weak westerly. As cooling continues at the high latitudes a strong cyclonic vortex, centred over the pole, develops and expands in early winter throughout the stratosphere with westerly winds increasing with height in all latitudes at heights above 25 to 30 kilometres. The circulation round the vortex then becomes asymmetrical with a well-marked and persistent ridge system over the Aleutians on the 100-, 50- and 30-millibar charts. The large-scale systems slowly move and oscillate in position. Locally, temperatures may change by 20 K between the cold trough and warm ridge regimes, while the westerly winds between them at about latitude 60° may reach 100 metres/second at 30 kilometres and 200 metres/second at 40-50 kilometres in the polar-night jet. Baroclinic disturbances of the stratosphere probably occur mainly above the 20-millibar level (about 26 km), but occasionally large-scale vertical motion develops throughout the middle and lower stratosphere, associated with large-scale circulation changes which give rise to the phenomena known as 'sudden warmings'. There seem to be two main types of development leading to sudden warmings. In the 'asymmetric' type, a warm cell moves polewards from middle latitudes, in the northern hemisphere usually on the Canadian side, and displaces the cold vortex at the poles. The strong westerly winds are replaced by light easterlies, and the cold vortex moves southwards, becomes less intense, and may finally disappear. The second, the 'bipolar' type, is characterized by the poleward movement of two warm cells from opposite sides of the vortex, usually from Siberia and Canada. The cold vortex elongates and splits, and the two sections move

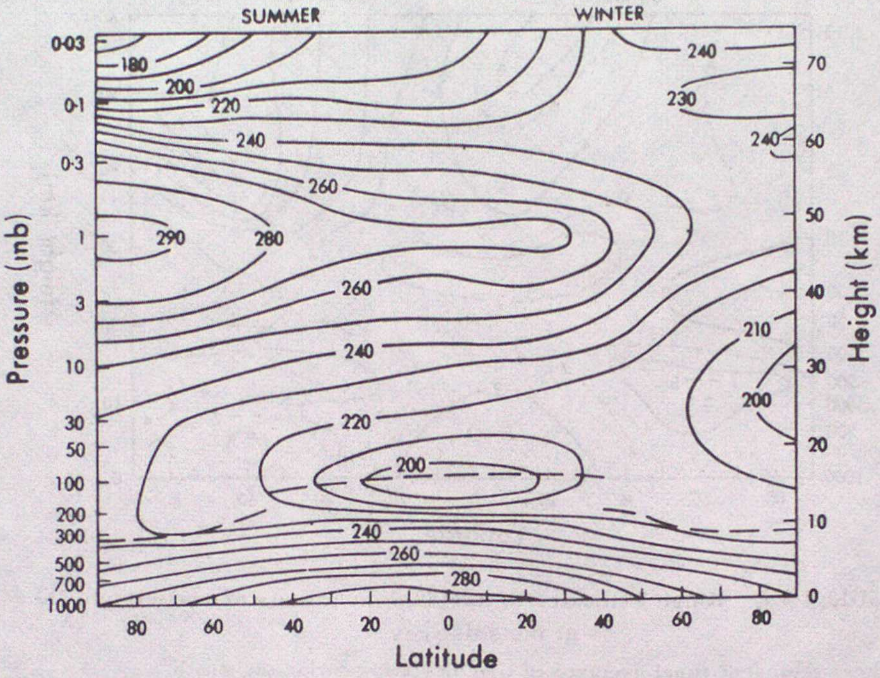

FIGURE 12(a). Latitudinal mean cross-section of temperature (K) up to 70 km

— — — the tropopause

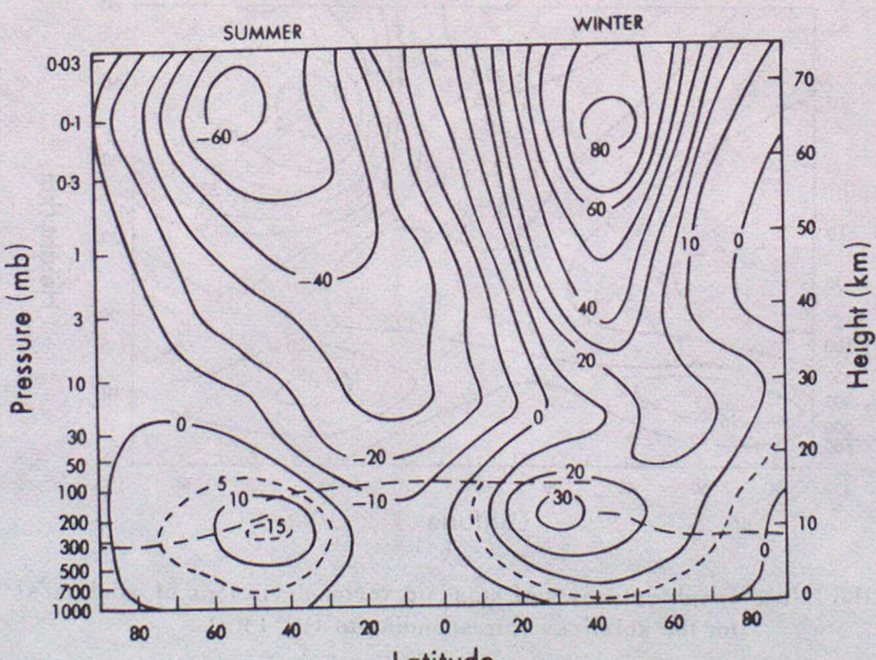

FIGURE 12(b). Latitudinal mean cross-section of zonal wind speeds (m/s) up to 70 km. (West to east components are positive)

— — — the tropopause

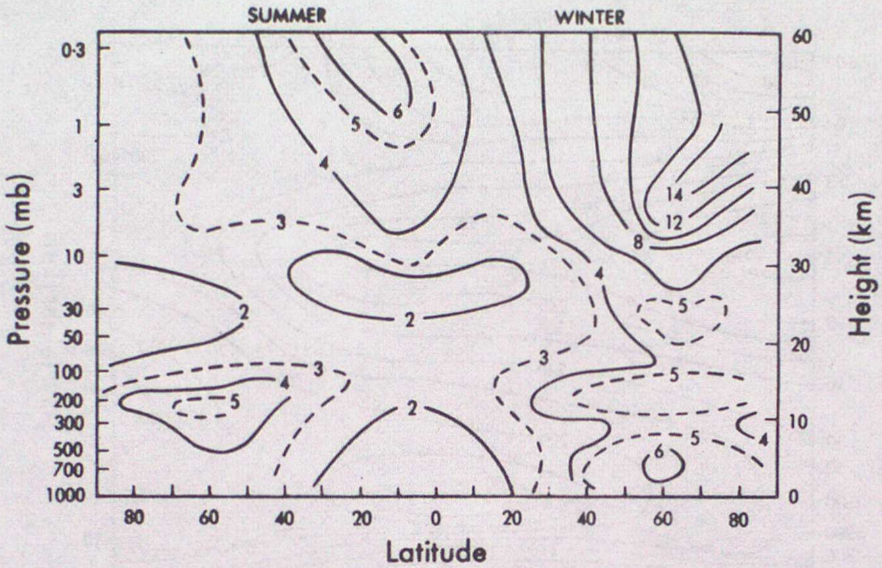

FIGURE 13(a). Rough estimates of standard deviations of temperature (K) at the solstices

These represent zonal averages of time deviation during the month.

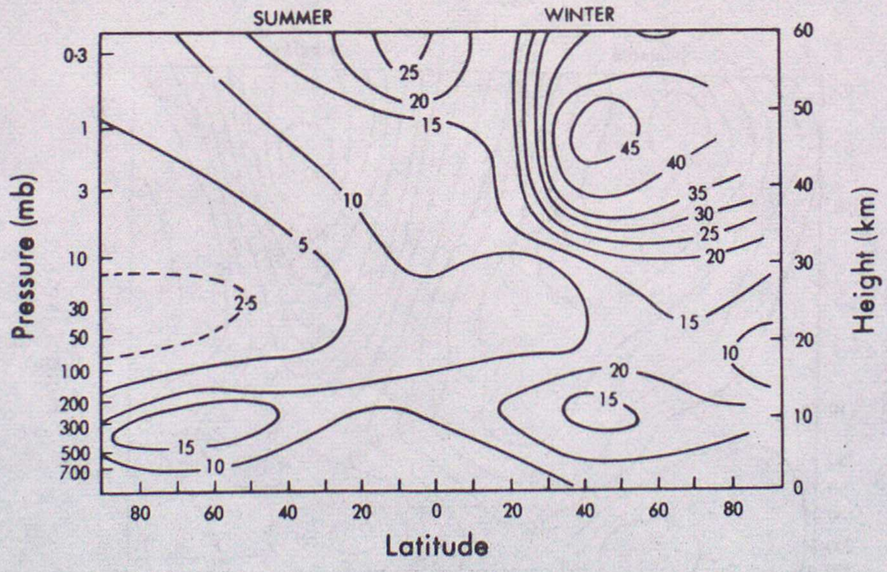

FIGURE 13(b). Rough estimates of standard vector deviations of wind (m/s) for the solstices corresponding to Fig. 13(a) southwards as the warm cells join over the pole. Again a warm region and light easterlies become established. During the history of the disturbance, the temperature may rise from –75 °C or so to somewhere near 0 °C. If the sequence occurs in late winter or early spring it is simply followed by the summer circulation regime and the anticyclonic flow spreads over the hemisphere as the polar regions are warmed by solar radiation. This is termed the 'final' warming, but if a large warming occurs in mid-winter it usually lasts over a period of about three weeks and is then followed by a re-establishment of the cold vortex at the pole. The winter regime will then be terminated by a later warming.

Major sudden warmings also occur in the southern hemisphere, but they may be less frequent than in the northern hemisphere, particularly before the spring equinox. In the northern hemisphere, topography probably has a considerable influence in setting up the asymmetric flow pattern and may also be a factor in the mechanism of the sudden warmings. In the southern hemisphere, the vortex is relatively symmetrical, but the horizontal temperature gradient is usually significantly greater than in the northern hemisphere, possibly giving rise to dynamical instability without the need for a contribution from topographically induced vertical motions.

In summer, as has already been mentioned, the dominating feature of the flow at 50 millibars and above, is a warm anticyclone centred near the pole with light easterly winds in middle latitudes. Minor troughs and ridges are sometimes evident in the circulation but, on the whole, the wind field is fairly flat and vector-wind changes with time are usually small, and almost independent of time for intervals of six hours to ten days or so. At levels above 50 millibars the easterly winds increase quite slowly with height, while the variability increases very slowly. Typically, over England in July the wind speeds are about 14 knots at 30 millibars and 20 knots at 10 millibars, from just south of east, with root-mean-square vector changes, over the periods indicated above, of about 7 knots at 30 millibars and 8 knots at 10 millibars. The stable, light easterly winds of summer enable estimates to be made of the various error components of the geopotentials derived from radiosonde data; details of the methods used are described in Chapter 11 - Upper-air charts.

It has not been possible in this section to give more than a very brief summary of some of the main features of the stratosphere, and many interesting topics, such as the 26-month oscillation of the wind field in the tropical stratosphere, have been omitted. Any reader who wishes to study further should refer to Murgatroyd's review paper,[14] and to WMO Technical Note No. 70,[21] both of which provide extensive bibliographies.

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