M.O.637

HANDBOOK

OF

WEATHER FORECASTING

METEOROLOGICAL OFFICE

1964

U.D.C.
551.509.3(02)

PREFACE

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

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

CHAPTER 7

SOME FEATURES OF THE FREE ATMOSPHERE

CONTENTS

Chapter 7

SOME FEATURES OF THE FREE ATMOSPHERE

LIST OF DIAGRAMS

Figure *Page*

CHAPTER 7

SOME FEATURES OF THE FREE ATMOSPHERE

7.1. INTRODUCTION

In addition to a working knowledge of the behaviour of synoptic systems as analysed on the synoptic chart it is desirable that forecasters should have some information on various features of the free atmosphere not always indicated in much detail by the analyses. An attempt is made in this chapter to provide that information. Both the selection of features and the extent of the treatment are rather arbitrary but both have been made with the needs of the practical forecaster primarily in mind. No attempt has been made to present a theoretical treatment. Where this is required the forecaster can obtain some theory from the standard textbooks and a more advanced account by consulting some of the original papers referred to in the text.

Current needs for operational forecasting require that this chapter should contain sections dealing with fronts, jet streams and the tropopause. In regard to the stratosphere, current needs would probably be adequately met by restricting the treatment to the lower stratosphere. However, it appeared that it might be advantageous to give some information to levels rather above those currently reached as routine in manned aircraft.

Most of the illustrations selected for this chapter refer to areas over or near north-west Europe. This selection has been made intentionally for the following reason. If there is a geographical bias in the features illustrated then a selection of illustrations with such a bias will probably be of greatest immediate practical value to forecasters engaged in day-to-day forecasting for those same areas. Illustrations for other areas (for example, jet streams near eastern Asia) might present more startling features but would tend to be misleading if applied uncritically to north-west Europe.

7.2. FRONTS

The concept of fronts, which was introduced many years ago by Norwegian meteorologists, is widely known and, when used with justification and discretion, is of great value to forecasters engaged on short-period forecasting. Care must be taken to guard against too zealous a use of fronts and against attempting to "force" the analysis to conform on all occasions to the classical textbook patterns. Experience of day-to-day forecasting will soon provide examples when analyses which are forced to fit some preconceived and pseudo-theoretical pattern of fronts are not supported by many of the observations. If the unrealistic analysis is persisted in (that is, the forecaster attempts to straitjacket the atmosphere to fit the frontal model too closely) it will generally be found that the forecast goes wrong both in principle and in detail. The atmosphere is complex at all times. Detailed observations made on a close time and distance network indicate that it is particularly complex near frontal regions. The treatment of fronts which follows has been designed to present an overall picture (without too much detail) of fronts and should be of value to analysts and forecasters engaged on routine forecasting.

Observations made by aircraft of the Meteorological Research Flight (M.R.F.) in the vicinity of fronts have been analysed by Sawyer.[1]* The original paper

* The superscript figures refer to the bibliography at the end of this chapter.

amply demonstrates some of these complexities of the atmosphere near fronts and also the variability between one front and another in the lower and middle troposphere. The detailed observations indicated that no idealized frontal model could adequately represent all the detail which was observable on the fronts investigated.

On analysed and forecast maps it is customary to indicate a front by a line of discontinuity but it is important to remember that, in the atmosphere, the front will extend over a zone or band of finite width. It has so far not been practicable to formulate a definition of a front which can be applied rigorously and without uncertainty to routine data on the forecast bench. No formal definition is attempted here and it therefore follows that there must be some uncertainty when certain meteorological aspects are discussed in relation to fronts. Nevertheless the uncertainty will be less than that which the practical forecaster is forced to accept in the current standards of routine analysis and forecasting and it should not detract unduly from the value of the following account.

When attempting to locate fronts on actual charts temperature differences are usually closely scrutinized. The detailed flight observations in frontal regions which were examined by Sawyer showed that the average width over which a horizontal temperature gradient existed was about 600 miles and the average temperature difference about 15°F. (about 8°C.). A temperature difference of about 9°F. (5°C.), that is rather more than half the total temperature difference, was concentrated over a distance of about 130 miles. (The temperature aspects of fronts are more fully discussed in Chapter 14.) Sawyer called this inner band of concentrated temperature difference the frontal zone. From his examination of temperature data Sawyer concluded that the variation from front to front and the uncertainties of the identification of the frontal boundaries were such that these figures could be little more than a broad indication of the usual structure of a frontal zone. It was fairly clear however that, in the free atmosphere, frontal zones less than 50 miles in width were unusual.

Aircraft observations near fronts have also revealed some interesting features of the distribution of humidity. One feature, which was more or less well developed on all the cross-sections which were examined by Sawyer,[1] was the existence at both warm and cold fronts of a tongue of relatively dry air extending downwards in the vicinity of the front and tilted in the same direction as the front. Sawyer[1] found that the dry tongue was less well developed in association with cold fronts than with warm fronts; it extended down to an average pressure level of 700 millibars in cold fronts and 800 millibars in warm fronts. In about half the fronts investigated the driest air was found within the frontal transition zone itself but there were also occasions when it was found on either the cold or the warm sides of the transition zone. These general findings have been confirmed by Freeman[2] who examined a further series of observations by the Meteorological Research Flight. Freeman found that the dry tongue could be traced upwards nearly to the tropopause.

The aircraft observations showed also that strong horizontal (isobaric) gradients of humidity over very short distances (on the synoptic scale) were not uncommon near frontal regions. Freeman has quoted the case of 13 January 1955 when a patch of very dry air with a frost-point depression of 25°C. occurred at a height of 8000 feet only 10 miles ahead of warm-front cloud and only 500 feet below the frontal cloud sheet above.

Some Features of the Free Atmosphere

The following extract from the conclusions reached by Sawyer[1] sums up clearly and concisely the existence, origin and importance (for forecasters) of these areas of dry air:

> "Some very dry air is usually found in frontal regions in close juxtaposition to cloud and precipitation. Its presence can be adequately explained by large-scale movements of the air which are primarily horizontal; the air has been dried by subsidence but this may have been 24 hr. or more before it was observed in the frontal region. The presence of dry air is no indication of descending motion in the front, nor of the absence of adequate moisture for frontal rain. Thus reported humidity on individual radio-sonde soundings cannot provide a forecaster with any reliable indication of the activity of a front."

From the data available to them, both Sawyer[1] and Freeman[2] were able to obtain estimates of the slopes of fronts. As might be expected they found considerable range in the slopes for the fronts investigated. The evidence showed that the slopes ranged from 1:30 to 1:250. Freeman reported as follows for his sample:

Cold fronts	*Warm fronts*
from 1:30 to 1:140	from 1:110 to 1:200

The aircraft observations examined by Sawyer[1] and by Freeman[2] gave ample proof of the wide variation of the cloud distribution accompanying fronts in the middle and lower troposphere. Some fronts were crossed without the aircraft flying through any cloud at the selected level and, at the other extreme, a horizontal flight of 300 miles was made entirely in cloud. The duration of flight in cloud and in clear air is shown in Table 1 which also gives an indication of the variation of the amount of cloud with height.

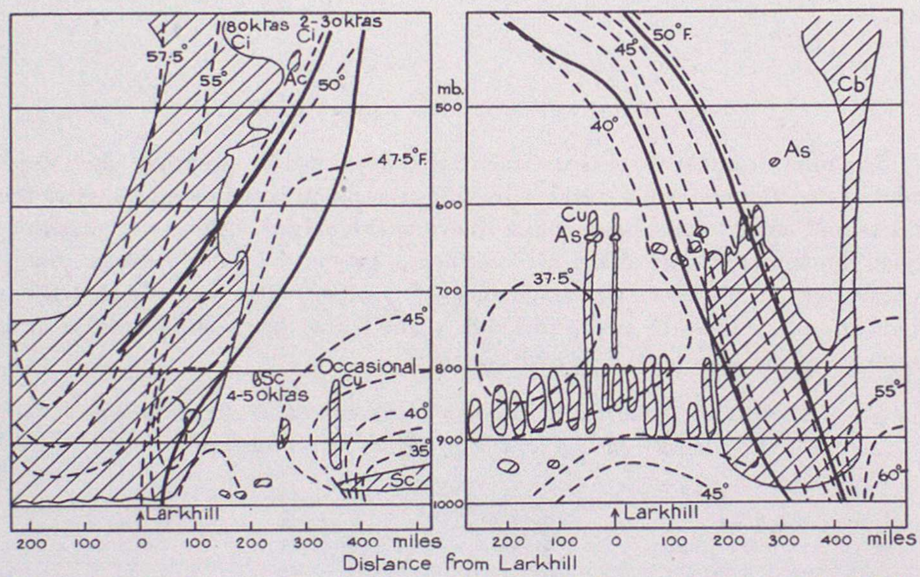
TABLE 1 *Duration of flight in cloud and clear air at various levels during series of flights (after Sawyer[1])*

Height range	*Duration of flight in cloud*	*Duration of flight in clear air*	*Proportion of flight in cloud*
ft.	*min.*	*min.*	*per cent*
Above 15,000	139	627	20
10,000 - 15,000	386	868	31
5,000 - 10,000	193	367	34
Below 5,000	197	415	32
All levels	915	2277	29

Sawyer comments as follows on Table 1: "As each front was only investigated at two levels and those not often widely separated, the figures in Table 1 are no doubt strongly influenced by the accidental choice of fronts and flight levels. However, the fairly uniform distribution of cloud at all levels up to 15,000 ft. and the relatively smaller amounts above 15,000 ft. conform with expectation and are probably real."

When interpreting Table 1 it should be noted that the majority of the flights extended over horizontal distances of between 200 and 300 nautical miles but a few were shorter or longer. Also, although the majority of flights intersected the frontal zone, some failed to do so. Thus Table 1 should be interpreted in a broad way as referring to the localities around frontal zones. Indeed, it must not be interpreted as if the cloud were symmetrically placed about the frontal zone. There is good evidence to indicate that, at most warm fronts and some cold fronts, the frontal cloud systems are mainly in the warm air mass and that the slope of the frontal cloud system is about twice as steep as the frontal zone. At distances greater than 200 miles ahead of the surface warm front the frontal cloud lies mainly in the warm air, but within 100 miles the cloud usually extends down through most of the transitional zone. This extension of the frontal cloud into the transitional zone is probably brought about by the evaporation of falling rain or snow and the subsequent recondensation of the water vapour. Freeman's analysis confirmed the general finding of the slope of frontal cloud being twice as steep as the slope of the frontal zone, both for warm fronts and also for slow-moving cold fronts. The levels at which the frontal cloud system intersects the warm boundary of the frontal zone is probably somewhat below 600 millibars.

The following figure and accompanying text are reproduced in full from Sawyer's paper.



Typical warm front, 1500 G.M.T., 6 March 1952 Typical cold front, 1500 G.M.T., 27 April 1951

FIGURE 1 *Cross-sections of cloud and wet-bulb potential temperature required for saturation at the observed temperature*

The broken lines indicate the wet-bulb potential temperature required for saturation

"The two cross-sections of cloud structure in Fig. 1 have been reproduced to illustrate some of the more common features of the cloud distribution. It would be wrong to regard these sections as typical of the flights as a whole, but some of their features occurred several times in the small sample which was available, and these diagrams may perhaps be regarded as illustrating common types of cloud structure at the more active warm and cold fronts.

The cross-section for March 6, 1952 in Fig. 1 shows a solid mass of cloud some 200 miles broad at most levels and extending upward well above the higher flight level of 650 mb. The upper limit of the cloud is uncertain, but has been inserted on the basis of the humidities reported on the radio-sonde soundings. A noteworthy feature, which also occurs at other warm fronts, is that the general inclination of the cloud mass is steeper than that of the front. The broken lines of Fig. 1 indicate the wet-bulb potential temperature required for saturation at the observed temperature. Since the wet-bulb potential temperature in the upper part of the cloud lies mainly between 52° and 58°F., the air comprising this part of the cloud may have been derived from the lower layers of the warm air mass. The lower part of the cloud lying within the frontal transition zone has a lower wet-bulb potential temperature. If it has been derived from the warm air mass it must have been subject to considerable cooling by precipitation. It is more probable that this cloud has been formed in cold or transitional air which has been moistened by precipitation.

Noteworthy features of the cloud structure at the cold front of April 27, 1951 (see Fig. 1) are the narrow area of large convection cloud in the warm air mass close to the surface front, and the absence of cloud in the frontal zone in the upper troposphere. Only 5 other cold fronts were investigated (and for one the cloud observations are incomplete). However, the frontal zone was free of cloud in the upper troposphere in all 6 fronts, and a rather narrow band of cloud with considerable vertical development occurred in the warm air above the surface front in 4 out of the 5 fronts for which this region was explored. The solid mass of cloud which appears on April 27 in the transition zone below 650 mb. was found only in this case, but this front was the most active of the series and such structure may be more common than its occurrence once in 5 fronts might suggest. On this occasion the aircraft passed through the top of the cloud mass which lay in the frontal zone and reported marked cumuliform structure. This is surprising in view of the likelihood of a stable vertical lapse rate in the frontal zone (confirmed by the Trappes radio-sonde). Possibly, however, the vertical convection currents did not extend through any considerable depth."

Most of the information in this section cannot be applied directly to forecasting but constitutes background knowledge which a forecaster should have when assessing available data and making forecasts. Two systematic statistical investigations of clouds at warm fronts by Matthewman[3] and by Sawyer and Dinsdale[4] have produced a variety of statistics relating cloud structure to various parameters of warm fronts. However, none of these seemed likely to be of direct value. Indeed Sawyer and Dinsdale reached a conclusion that "it would appear that there are no simple and reliable measurements that a forecaster can make in order to estimate the cloud structure at a warm front."

For some considerable time forecasters have recognized that the majority of cold fronts may be placed in one of two main types which have been termed ana-cold fronts and kata-cold fronts. Broadly speaking an ana-cold front may be regarded as a cold front at which the warm air is ascending and is moving forward less rapidly than the frontal surface beneath. At a kata-cold front the warm air is descending and the air just above the frontal surface is moving forward faster than the frontal surface beneath. (It should be noted, however, that there is frequently a region of ascending warm air some little distance ahead of the surface position of a kata-cold front.) In view of the different vertical motions associated with ana-cold and kata-cold fronts it would be reasonable to expect the two types of fronts to display quite different characteristics.

Sansom[5] has examined some cold fronts over the British Isles and his results contain qualitative and quantitive information of practical value on the forecast bench. That part of Sansom's results which relates to the general aspects of ana-cold and kata-cold fronts is discussed below. (Some other results of value in estimating the weather likely to be associated with cold fronts are included in Chapter 10 Section 10.3.) The characteristics of ana-cold and kata-cold fronts are summarized in Table 2.

TABLE 2(*a*) *Some characteristic surface features of ana-cold and kata-cold fronts (after Sansom*[5]*)*

Surface features	*Ana-cold front*	*Kata-cold front*
Temperature	Often a large fall which may be sudden - mean temperature drop 6°F.	Changes may be very slight and gradual - mean temperature drop 1.5°F.
Relative humidity	High and changes slight.	Decreases and the fall may be of considerable magnitude and quite sharp.
Precipitation	Generally fairly heavy rain at the frontal passage with steady lighter rain for some time (perhaps 2-3 hr.) behind the front. A comparison by Sansom showed for each frontal passage an average fall of 6.5 mm.	Amounts generally very slight and frequently nil. The precipitation amounted to only 0.8 mm. on average and fell immediately before or during the frontal passage. There was very little post-frontal rain.
Wind	Usually a sharp veer accompanied by a sharp decrease in wind speed.	Wind veer may be very gradual and speed changes usually slight.
Pressure field (frequency of occurrence of sharp, average or weak pressure troughs).	Sharp 7 Average 2 Weak 6	2 8 13

TABLE 2(*b*) *Some characteristic upper features of ana-cold and kata-cold fronts (after Sansom*[5]*)*

Upper features	*Ana-cold front*	*Kata-cold front*
Radio-sonde sounding started in cold air and passing through the frontal layer	No sharp temperature discontinuity. Relative humidities high at all levels. Ascent of warm air led to the high humidities, often saturated conditions, above the front.	A fairly sharp inversion of temperature or isothermal layer with very low humidities at higher levels. The inversion was primarily due to subsidence which was usually occurring in both the warm and cold air masses and the base of the inversion was normally in the cold air.
Wind sounding started in cold air and passing through the frontal layer	The component of wind normal to the front decreased slightly while the component parallel to the front increased rapidly with height. The total backing of the wind from 950 to 400 mb. was about 65°. At 500 mb. the mean wind direction was inclined at only 16° to the tangent to the front. The mean thermal wind between 950 and 400 mb. was almost parallel to the front (actually backed by some 3½°) and corresponded to a thermal gradient of 1.25°F. (0.7°C.) per 50 km.	The components of wind parallel and normal to the front both increased with height. The total backing of the wind from 950 to 400 mb. was only 20°. At 500 mb. the mean wind direction was at an angle of 42° to the tangent to the front. The mean thermal wind was inclined some 30° (i.e. veered) across the front and corresponded to a temperature gradient of 0.9°F. (0.5°C.) per 50 km. (This cross-front thermal wind is an essential condition for an active kata-front and can be readily recognized.)
Relation between the component of wind normal to the front and the speed of the front	The mean speed of the front exceeded at all heights the mean component of wind normal to the front, though the difference was insignificant below 800 mb. Higher up the difference was definite which implied that both warm and cold air masses were ascending.	The mean component of the wind normal to the front exceeded the speed of the front at all levels above the lowest layers, implying that both air masses were generally descending.
Angle between the tangent to the cold front and the mean wind direction in the warm sector some 50 to 150 miles ahead of the front	At 700 mb. 18° At 500 mb. 17°	35° 35°
Average upper wind veer at passage of front	At 700 mb. 23° At 500 mb. 15°	15° 5°

The data given in Table 2 should enable forecasters to recognize well marked ana-cold or kata-cold fronts on actual charts. The problem of forecasting whether a front will become or remain of "ana" or "kata" type is rather more difficult. In general cold fronts do not remain solely as a kata-front or an ana-front throughout their existence. Very occasionally they may do so and Sansom considers that this is most likely to occur when a small secondary depression has formed on a cold front within the upper circulation of the primary depression. In such a case the cold front of the secondary will generally be a kata-cold front as the upper winds will probably have a large component across the front. The more general case of cold fronts will show a transition between "ana" and "kata" types both in space and time. It is difficult to generalize but it seems likely that initially the cold front of a developing depression is of "ana" type over most of its length. Sansom found that most of the kata-fronts were occluding more actively than ana-fronts and it seemed likely that the kata-front occurred first near the centre of an occluding depression and spread outwards along its length. In the final stages of a depression when the upper circulation was well developed there might be a substantial upper flow across the cold front which would then exhibit "kata" characteristics over substantial portions of its length. However, the formation of wave disturbances might complicate the pattern and cause segments of the front to exhibit "ana" or "kata" characteristics. The subsequent variation and movement of this pattern will then depend on the synoptic development of any such wave.

Sansom found no direct relation between the speed of a front and "ana" or "kata" characteristics. There did appear to be some slight association with the curvature of the isobars behind the cold front and a greater association with the change in curvature of the isobars in the cold air towards the front. The supporting evidence is shown in Tables 3 and 4.

TABLE 3 *The curvature of the isobars behind the cold front*

	Cyclonic	*Straight*	*Anticyclonic*
		Number of cases	
Ana-cold fronts	7	3	5
Kata-cold fronts	7	3	13

TABLE 4 *The change in curvature of the isobars in the cold air towards the front*

	Increasing cyclonic or decreasing anticyclonic	*No apparent change*	*Decreasing cyclonic or increasing anticyclonic*
		Number of cases	
Ana-cold fronts	11	2	2
Kata-cold fronts	0	6	17

Synoptic examples of both types of fronts are discussed in Chapter 10.

7.3. JET STREAMS

7.3.1. *General*

The Executive Committee of the World Meteorological Organization has, at its ninth session,[6] adopted the following definition:

"A jet stream is a strong narrow current, concentrated along a quasi-horizontal axis in the upper troposphere or in the stratosphere, characterized by strong vertical and lateral wind shears and featuring one or more velocity maxima."

For operational purposes the Committee recommended the following numerical values:

"Normally a jet stream is thousands of kilometres in length, hundreds of kilometres in width and some kilometres in depth. The vertical shear of wind is of the order of 5–10ms^{-1} per kilometre and the lateral shear is of the order of 5ms^{-1} per 100 kilometres. An arbitrary lower limit of 30ms^{-1} is assigned to the speed of the wind along the axis of a jet stream."

(Note.- Vertical shear 5–10 metres per second per kilometre = 9·7 – 19·4 knots per kilometre or 3·0 – 5·9 knots per 1,000 feet; lateral shear 5 metres per second per 100 kilometres = 9·7 knots per 100 kilometres or 18·0 knots per 100 nautical miles; speed of 30 metres per second = 58·2 knots.)

The existence of strong upper winds at cirrus levels had been commented upon by a number of workers since early in the twentieth century but the subject sprang into prominence in the 1940's when aircraft operating regularly at and above cirrus levels encountered these belts of strong winds. Organized and intensive investigations of the jet stream were commenced and it soon became apparent that the jet stream was a feature of the atmosphere of great importance both for aerial navigation and for the understanding both of the general circulation of the atmosphere and of the day-to-day behaviour of the major and minor synoptic systems. A great deal of research on and investigation of the jet stream has been carried out in North America by many workers. Many investigations in other countries have also been undertaken and the published literature on the jet stream is voluminous. A useful synopsis of the jet stream and a survey of the literature are published in *World Meteorological Organization Technical Note* No. 19.[7] The following account of the jet stream is very condensed and the major emphasis has been concentrated on the features of the jet stream over or near north-west Europe. For those forecasters for whom this account is inadequate, either in the detail or manner of treatment or in geographical limitations, the *World Meteorological Organization Technical Note* No. 19 will probably form a useful starting point for a wider study of the jet stream and the available literature. An earlier annotated bibliography of the jet stream was prepared by Johnson[8] and distributed within the Meteorological Office. The monograph prepared by Riehl and collaborators[9] contains a useful account of several aspects of jet streams.

The velocity of the winds in jet streams depends almost entirely on the existence of pronounced gradients of temperature which become concentrated in horizontal bands through a substantial depth of the atmosphere. For some jet streams surface pressure gradients are important at times but, generally speaking, the velocity of a jet stream is fairly closely related to the thermal wind determined by the total thickness pattern through the layer containing the pronounced thermal contrasts. Ageostrophic winds may be substantial in various sectors of jet streams but they are not sufficiently great to invalidate the previous statement. When jet streams are present they are normally observed to occur in the upper troposphere. Jet streams will occur in those parts of the atmosphere where sufficiently strong contrasts of temperature are built up through either seasonal influences, or more transient synoptic influences which persist for periods which can be measured in days. Current work indicates that in the upper troposphere there are probably five distinct systems of jet streams which may exist but not necessarily simultaneously. There are the subtropical jet streams in the vicinity of latitudes 30°N. and 30°S. and the jet streams associated from time to time with the polar fronts in higher latitudes in both northern and southern hemispheres. These four jet streams have a westerly component when viewed on a global basis but there may be marked deviation from westerly flow from time to time with geographical location. Fifthly there is evidence for a jet stream with a predominantly easterly flow located in equatorial latitudes and mainly in longitudes where the upper easterlies are displaced 10° or so from the equator. All these jet streams usually attain their maximum speeds near the level of the tropopause. To fix ideas this means that the equatorial jet stream is at a level of about 100 millibars, the subtropical jet stream about 200 millibars and the jet stream in higher latitudes often around the 300 – 250-millibar levels. The levels of the jet streams are not of course constant either on a geographical or time basis.

The equatorial jet stream owes its existence to the fact that the middle and upper troposphere is cooler near the equator than over the subtropics of the summer hemisphere. The Eurasian land mass enhances the effect in July. This temperature distribution produces the easterly flow. When considered on a global scale, the equatorial regions over which the upper winds reach jet speeds are quite small and over extensive belts of longitude the winds may be well below jet speeds.

Gilchrist[10] has examined subtropical jet streams and shown that they occurred entirely within the tropical air mass but they appeared to be associated with the lower tropopause of the colder air mass to the north. (There was a tendency for the main vertical wind shear to be found just above the level which was isentropic with the polar tropopause.) The subtropical jet stream is usually at or somewhat above the 200-millibar level. It appears to be independent of the presence of surface fronts; there is generally no evidence of its presence at 500 millibars and it may not be apparent even at 300 millibars. The cause of the subtropical jet streams has not been firmly established. Sawyer[11] remarks "it seems difficult to regard the subtropical jet stream as set up by confluent wind fields associated with large scale eddy motion. It seems much more likely that the subtropical jet stream is a direct consequence of a slow northward drift of air in the middle and upper troposphere from near the equator to latitudes of 20 or 30 degrees. At the equator the air shares the rotation of the earth and if it retains its angular momentum it must move faster as it reaches the latitude circles of 20° or 30° which are smaller than the equator. The slow northward drift is part of a circulation first remarked by Hadley, in which air approaches the equator in the lower atmosphere in the trade-wind currents, ascends there and returns aloft. Such circulations are observed in dishpan experiments at lower rotation rates than those which give rise to a 'meandering jet stream', and it therefore appears reasonable that a simple Hadley-type cell should appear in lower latitudes on the earth where the Coriolis parameter is small. A ring of air displaced northward from the equator to latitude 30° and conserving its angular momentum would attain a velocity relative to the earth exceeding 200 kt so that the observed winds speeds are not unreasonable as arising from this cause."

Unlike the equatorial easterlies the subtropical jet streams often extend over wide belts of longitude interspersed with relatively narrower belts of lighter winds. In the northern hemisphere the subtropical jet streams, in the mean, reach their maximum speed and most southerly position (about 25°N.) in January when the area of equatorial easterlies is a minimum. The subtropical jet stream is weakest and farthest north (about 35 to 40°N.) in July at which time the area of equatorial easterlies, which are then most developed, is a maximum.

The jet streams of middle latitudes are caused by the concentration of temperature contrast which is set up from time to time by the confluent patterns associated with some suitable distributions of major synoptic systems. Although strong fronts are not necessary for the formation of these mid-latitude jet streams, on many occasions well marked fronts often exist in close association with these jet streams. In northern latitudes the mid-latitude jet stream is often separate from the subtropical jet, giving two distinct wind maxima on a meridional vertical section through the atmosphere. The two jet systems sometimes approach one another to form a jet stream which appears to have a single maximum of wind when the analyses are made on the scale and with the detail which can be customarily accomplished at the majority of forecasting offices. More leisurely and elaborate analyses including the construction of detailed vertical cross-sections may, however, reveal the existence of two maxima in the inner parts of such a combined jet stream. In the eastern Atlantic to the south-west of the British Isles the merging of the two jet streams may occur when a strong polar outbreak carries the mid-latitude jet to lower latitudes. Towards the other extreme the two jet streams may be quite distinct and separated by almost 30° of latitude as for example when a large blocking high to the west of the British Isles causes a split in the strong upper flow, one branch moving (perhaps to the north of 60°N.) round the northern flank of the blocking high and the other branch moving well to the south of the blocking high (perhaps near 30°N.).

Before leaving this short discussion of jet streams in general, mention should be made of the very strong winds which occur at levels above 50 millibars in the polar stratosphere in winter. At these levels in winter there is perpetual darkness and stratospheric temperatures in polar regions become very low and strong westerlies (in the mean) are believed to occur. Recent investigations indicate that at times the wind régime of the lower stratosphere contains a jet stream with a structure rather similar to that of a tropospheric jet stream. Some brief consideration of these and other winds at levels higher than 100 millibars will be included in Section 7.5.

For the remainder of this section the discussion will be limited to the mid-latitude jet stream and, for the reasons given in Section 7.1, the illustrations and numerical values will be based mainly on information obtained in the neighbourhood of the British Isles.

7.3.2. *Some features of jet streams in mid-latitudes*

At the outset it must be remarked that the mid-latitude jet streams show great diversity between one jet and another and any one jet stream displays variations from day to day and from place to place. In these respects a mid-latitude jet stream is very similar to some other features of the troposphere which, as meteorologists soon recognize, show a wealth of detail and variation in behaviour. In spite of this diversity it is useful background knowledge to be aware of some mean and probable extreme values for various features of the mid-latitude jet stream. Since wind is one of the most significant features of the jet stream, particularly for aerial navigation, the wind régime will be discussed first.

7.3.2.1. *Winds.* Murray and Johnson[12] have examined the structure of the upper westerlies in September 1950 over the eastern Atlantic and western Europe. Cross-sections along lines from Angmagssalik (Greenland) across the British Isles to near Malta and from Lagens (Azores) across southern England to near Stockholm were prepared daily on the basis of the 1400 G.M.T. upper air observations. The wind field was represented by the component of the wind perpendicular to the line of the section, being regarded as positive for flow from the Atlantic through the sections to Europe and negative for the reverse direction. Two examples of actual wind fields which commonly occurred are shown in Figures 2 and 3.

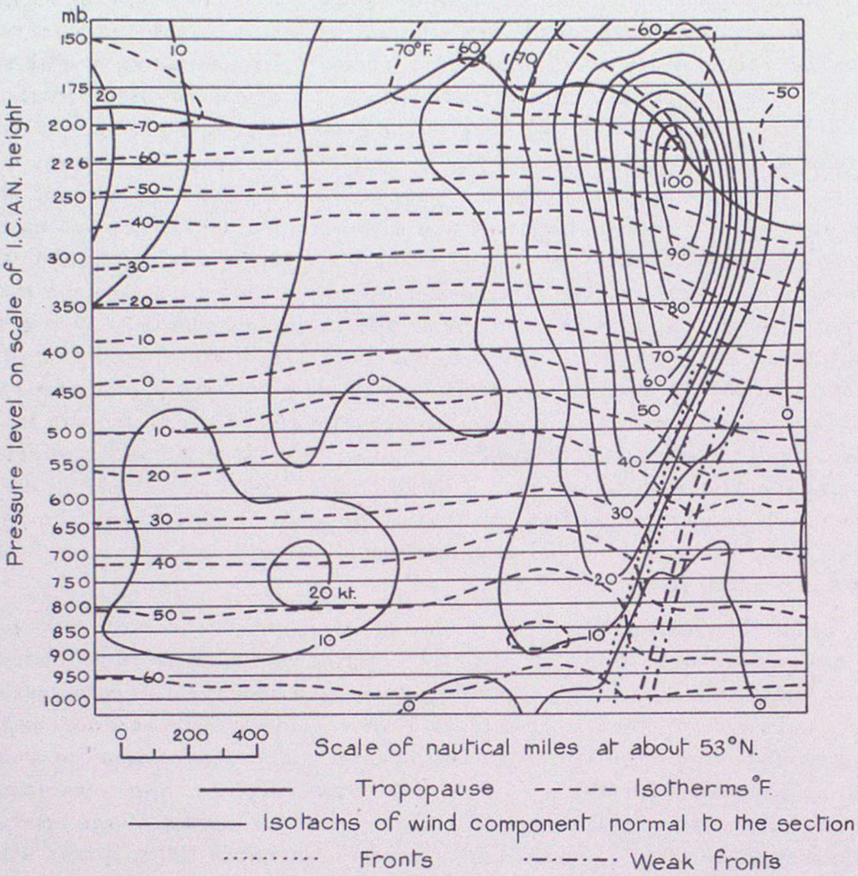


FIGURE 2 *Cross-section from the Azores to the Baltic, 1400 G.M.T., 4 September 1950*

The sections illustrate clearly the horizontal and vertical structure of the wind régime. In interpreting the figures it should be noted that the vertical scale is very greatly exaggerated in relation to the horizontal scale. Thus, although at first sight the jet appears to be a narrow filament of wind elongated in the vertical, on a true scale it is rather a pseudo-horizontal ribbon of strong wind of relatively shallow depth, that is a depth of perhaps 2 or 3 miles, a width of some 200 or 300 miles and a length of many hundreds or even a few thousands of miles.

The high wind speed can be accounted for by the accumulation of the thermal wind effect throughout the troposphere. At the lower tropospheric levels the strong horizontal thermal contrast is usually contained within narrow frontal zones but in the upper troposphere the isotherms associated with jet streams slope relatively steeply throughout a deep layer which is nearly vertical. The changes in temperature lapse on passing through the tropopause, which is at different levels in the warm and cold air masses, bring about a reversal of the thermal gradient in the lower stratosphere so that in the lower stratosphere the thermal wind detracts from the jet stream and wind speeds diminish with height. It will be noted from Figures 2 and 3 that the diminution in speed towards the colder tropospheric side of the jet is very well marked.

Murray and Johnson[12] obtained mean velocity profiles at the levels of the centres of the jet streams in September 1950. They distinguished between jet streams which were moving laterally towards the warmer tropospheric air (that is cold-front type or C jets) and those moving towards colder tropospheric air (that is warm-front type or W jets). Figure 4 shows these results.

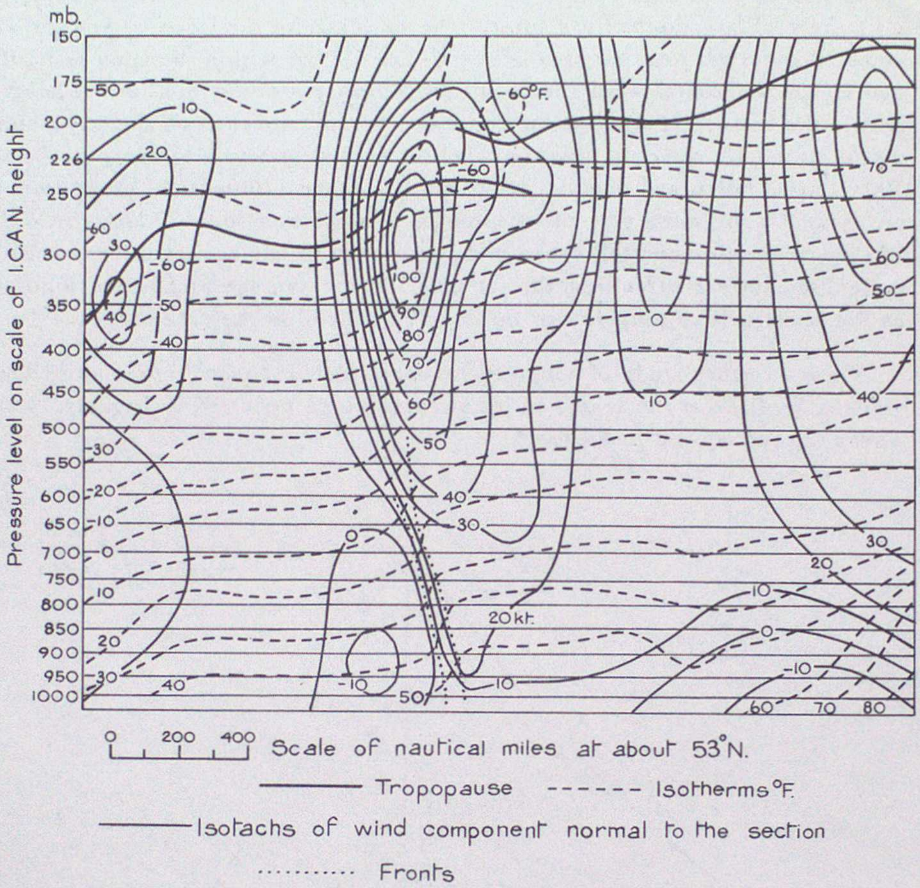


FIGURE 3 *Cross-section from Greenland to Malta, 1400 G.M.T., 30 September 1950*

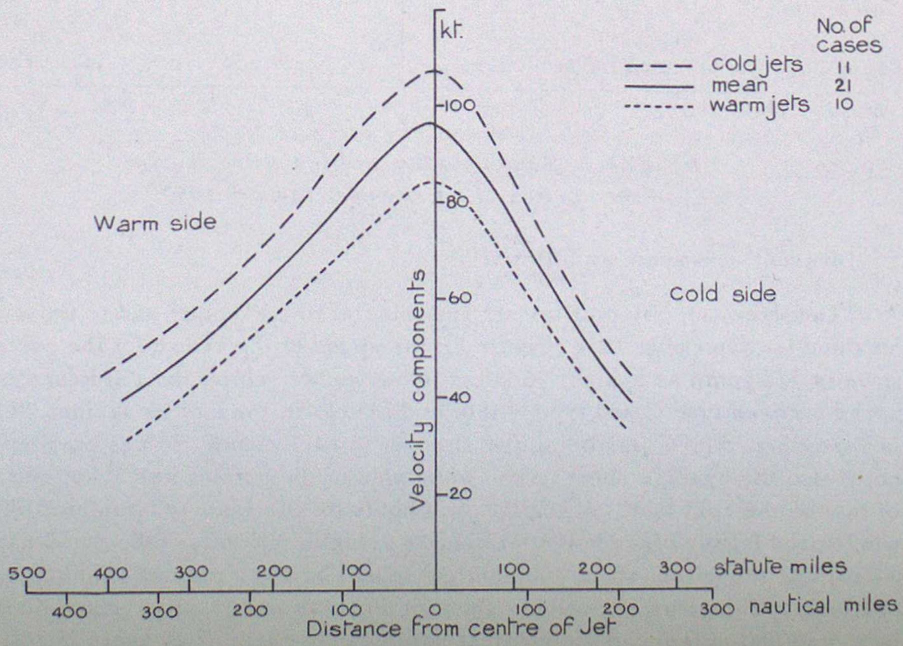


FIGURE 4 *Mean velocity profiles at the levels of the centres of jet streams, September 1950*

It will be seen from Figure 4 that the profiles for C and W jets and the overall mean profile are broadly similar. The intensity of the mean C-type jet was some 20 knots greater than that of the W-type jet but within the main body of the current the horizontal wind shears for each profile are very similar. In each profile the horizontal wind shear is greater on the cold than on the warm side of the jet axis. On the cold side the wind decreased by some 24 knots in the first 100 nautical miles and diminished to one half some 170 nautical miles from the jet axis. On the warm side the diminution in wind was some 17 knots in the first 100 nautical miles and the speed had diminished to one half of the maximum value some 280 nautical miles from the jet axis. In the average profile the wind shear on the warm side is roughly two thirds of the shear on the cold side.

From a further study of winds in January 1950, Johnson[13] has obtained mean velocity profiles at the levels of jet streams which occurred in January 1950. These are reproduced in Figure 5.

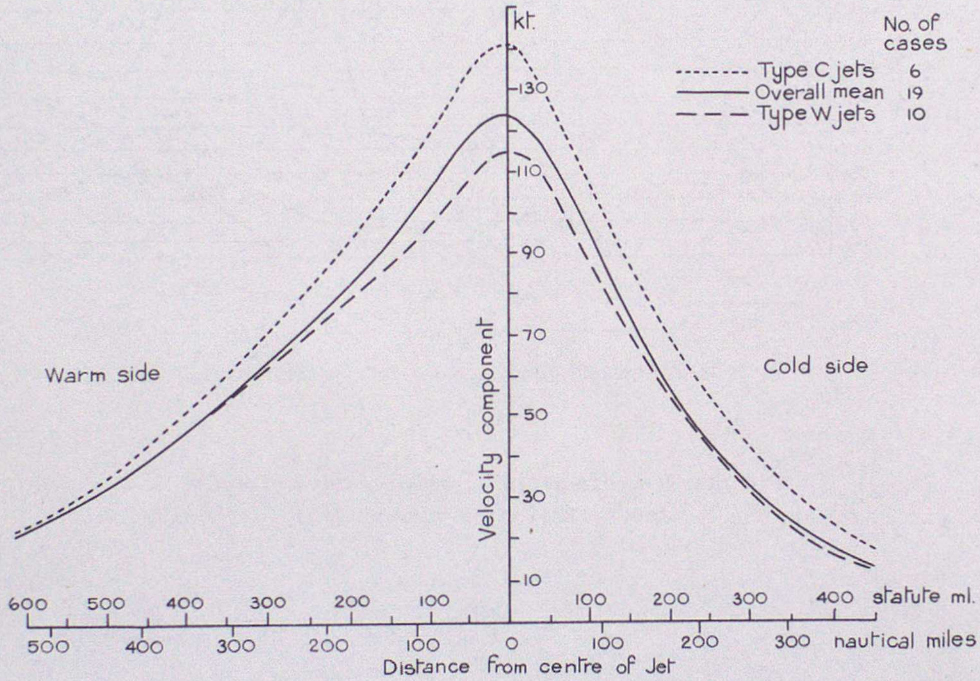


FIGURE 5 *Mean velocity profiles at the levels of the centres of jet streams, January 1950*

Johnson[13] comments as follows:

"The three velocity profiles are very similar to each other and to those obtained for September 1950 (Figure 4); the speed at the centre for the overall mean is 124 knots as against 96 knots in September, whilst the difference in speed between type C and type W jets is 26 knots in January as against 20 knots in September, type C jets being the stronger in each month. It was previously noted that the average shear on the warm side of the current was about two thirds of that on the cold side. A similar ratio holds for the January mean profile, the wind speed falling off to half its maximum value in 300 miles (about 260 nautical miles) and 180 miles (about 150 nautical miles) on the warm and cold sides respectively. The greatest cyclonic shear of 40 knots in 100 miles (about 40 knots in 85 nautical miles) and the greatest anticyclonic shear of 25 knots in 100 miles (about 25 knots in 85 nautical miles) both occur at about 100 miles (85 nautical miles) from the centre of the mean curve.

The individual velocity profiles from which the mean profiles were constructed were all of the same general form as the mean (cases of double maxima being specifically excluded), although there was considerable variation in the relative intensities of the cyclonic and anticyclonic shears. In no case did the anticyclonic shear exceed the cyclonic shear. The greatest observed cyclonic shear was again of the order of 100 knots in 100 miles (85 nautical miles), 100 miles being a lower limit to the distance over which it is possible to estimate isobaric shear with this network." (Wind speeds in jet streams deduced by Hurst[14] from photographic reconnaissance flights suggest that somewhat greater shears may occur over shorter distances.)

The profiles of some individual jet streams near the British Isles will show substantial variations from the mean curves of Figures 4 and 5. Two very different profiles illustrated by Murray and Johnson[12] are reproduced in Figure 6. These profiles should not be regarded as extremes but rather as examples of the way in which individual jets can vary from the mean.

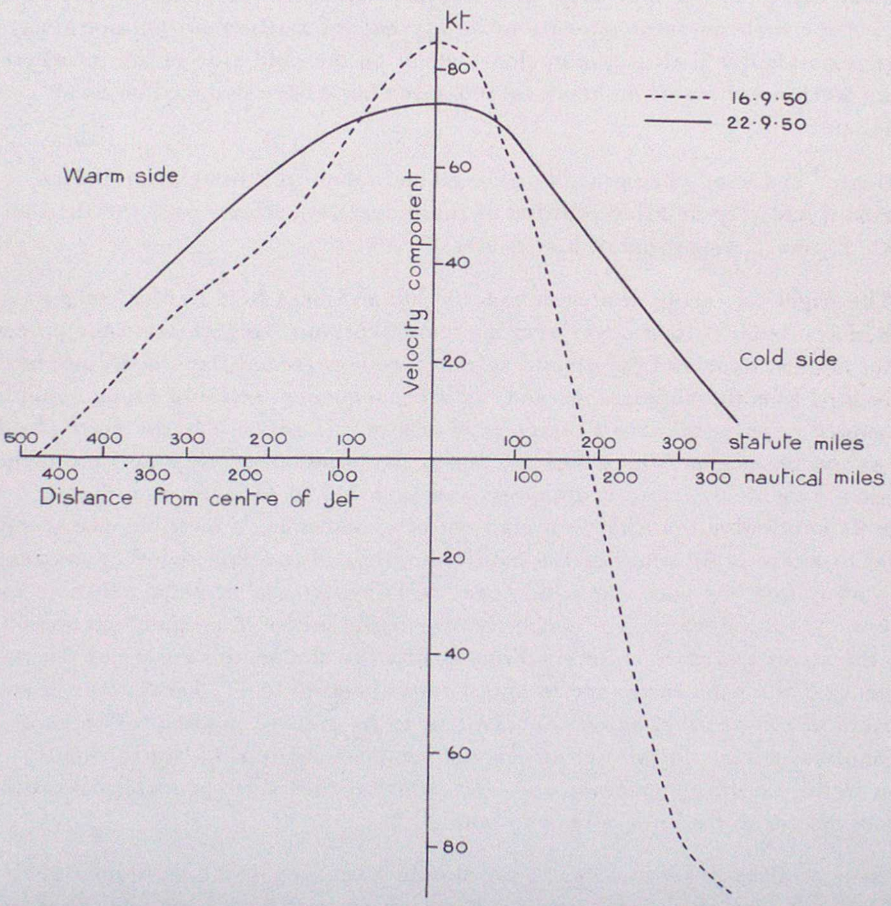


FIGURE 6 *Velocity profiles at the levels of the centres of two individual jet streams*

The profile for 16 September 1950 showed a sharp maximum and over more than 200 nautical miles the shear had a value of about 55 to 60 knots per 100 nautical miles. In contrast the profile for 22 September 1950 showed a flat maximum with shears which did not appear to exceed 20 knots per 100 nautical miles even on the cold side. This wind shear was somewhat below the average and quite small relative to that of 16 September.

The cross-sections and profiles of wind shown in Figures 4 to 6 inclusive were based on wind components normal to the sections. Murray and Johnson[12] estimated that if the orientation of sections had been adjusted to be more nearly at right-angles to individual jets, then slightly greater wind maxima and shears would have been obtained. In particular they estimated that the maxima in Figure 4 would be increased by about 10 knots and the mean shears by about 10 per cent. When interpreting the results given by Murray and Johnson[12] and also those given by Johnson,[13] it should be remembered that the data were for September and January 1950 only. The results should not therefore be regarded as true means; they must be biased towards those applicable to the winter half of the year and also by the type of synoptic weather which prevailed in the particular months examined.

In regard to horizontal shear it should be remarked that theory implies that, if the anticyclonic shear on the warm side of the jet stream exceeds numerically the Coriolis parameter, the flow will become unstable and break up into eddies. In latitude 50° the limiting shear is about 40 knots in 100 miles. No observational evidence has been found to show that greater shears are found on the anticyclonic side but the limiting value appears to be approached fairly often but not always. There is no similar limit to the cyclonic shear on the cold side of the jet where shears well in excess of 40 knots in 100 miles have been deduced on many occasions.

Hurst[15] has used photographs of the ground taken from aircraft to deduce horizontal and very detailed profiles of individual jet streams over the British Isles. Figure 7 shows one of his results.

The flight was made at almost exactly 300 millibars from Bedford to the north of Scotland. Cloud cover over southern Scotland, very close to the core of the jet stream, obscured the ground so that frequent ground fixes could not be determined from the photographs and, as a consequence, reliable winds could be determined at only (relatively) large intervals in this region. In the north (cold) side of the jet stream (which was a C jet in the terminology used by Murray and Johnson[12]) the decrease in wind speed was from 112 to 45 knots in a distance of about 75 nautical miles (that is a gradient of 67 knots in 75 nautical miles, equivalent to a rate of 89 knots in 100 nautical miles). From this point 75 nautical miles away from the core, the wind speed remained steady at about 45 knots for some way. On the warm side of the jet stream the increase in speed on approach from the south was more uniform. From Bedford to the southern edge of the jet-stream axis the wind increased in speed from about 40 to 112 knots, that is some 72 knots in 200 nautical miles – equivalent to an average gradient of 36 knots in 100 nautical miles. In this jet stream the maximum ratio of the anticyclonic shear to the Coriolis parameter was 0·83. On the cold side the maximum ratio of cyclonic shear to Coriolis parameter was 1·75.

Some profiles of jet streams in the Middle East area which were obtained also from photographs of the ground taken from aircraft have been described by Harding.[16]

The wind shears which have been discussed so far relate to horizontal shear. It will be readily seen from Figures 2 and 3 that strong vertical wind shears also occur in association with jet streams. The vertical wind shears of the jets examined by Murray and Johnson[12] were rather variable from jet to jet and mean values are not available. The greatest shears normally occurred in the troposphere below and on the cold side of the jet. If the frontal structure was well marked

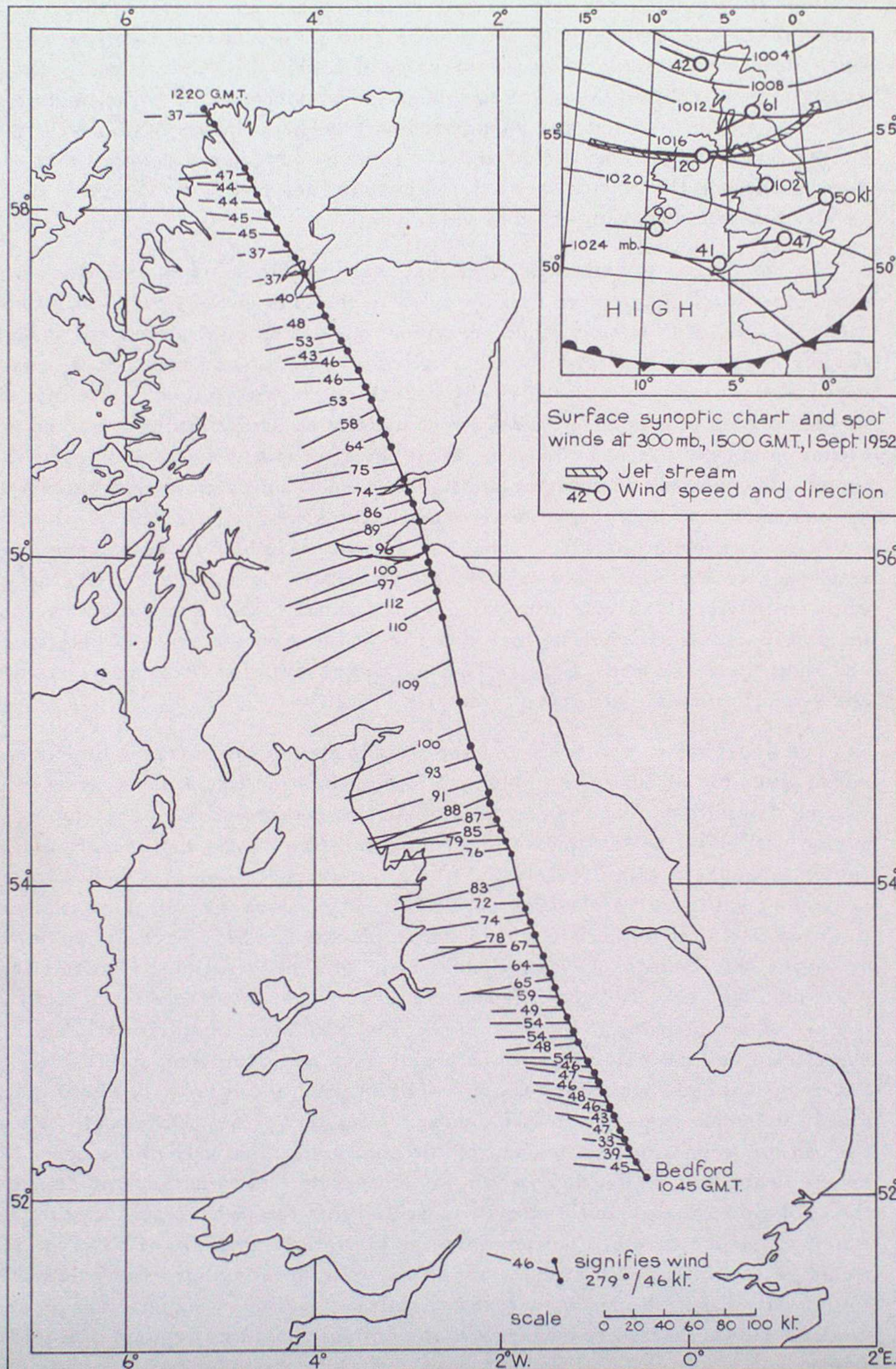


FIGURE 7 *Synoptic situation, trajectory of flight and vector winds across the jet stream of 1 September 1952*

The flight was at 30,000 feet (indicated) with altimeter sub-scale setting of 1013 millibars—almost exactly 300 millibars on this occasion.

there was usually an associated concentration of vertical shear with values of the order of 10 to 15 knots per 1,000 feet. On some occasions, however, the thermal contrast was spread over a wide zone and vertical shear was rather uniform throughout the troposphere, perhaps not exceeding 5 knots per 1,000 feet. Vertical shears above the jet also varied from case to case. On some occasions the total change of wind above the jet was spread more or less evenly over a fairly deep stratospheric layer (of the order of 12,000 feet) but quite often a very strong thermal gradient, opposed to that in the troposphere, set in immediately above the tropopause and was associated with vertical shears of about 10 to 15 knots per 1,000 feet (compare Figure 2). Johnson and Murray deduced that on some occasions the vertical shear just above the jet may be two or three times greater than the shear anywhere in the troposphere.

The variations in horizontal wind discussed so far relate to variations in directions which are nearly at right-angles to the axis of the flow. Variations along the line of flow also occur. A typical mid-latitude jet stream will usually exhibit, over a distance of a thousand miles or more, one or more zones, elongated along the jet axis, in which the speeds reach peak values. On either side of such a zone of maximum wind there is usually an area showing a marked decrease in speeds. In some of these areas the speeds may fall below jet-stream limits. These areas of stronger and lighter winds usually move downstream at speeds much less than those which occur in the strong-wind zones (the jet stream may also be moving laterally). Thus the air moves through these patterns and undergoes acceleration when approaching an area of maximum speed and deceleration on leaving it. These areas of maximum and minimum speed are sometimes intimately associated with typical synoptic features on the scale of both long- and short-wave systems. Some remarks on the association between jet streams and synoptic features are given in Section 7.3.2.7.

The acceleration which air undergoes as it approaches a region of maximum speed gives the air an ageostrophic motion towards the left of the contours looking downstream. Thus in an area of confluent contours the ageostrophic motion will be to the left toward lower heights. Conversely the ageostrophic motion associated with deceleration which the air undergoes at a diffluence will be to the right toward higher contours. These components of the wind transverse to the axis at the entrance to and exit from jet streams have been demonstrated by Murray and Daniels[17] by direct analysis of wind observations. These components average about 10 knots. It seems probable that these relatively small transverse components of wind are associated with some characteristic and systematic vertical velocities which are of much meteorological interest. Several years ago Namias and Clapp[18] showed that there were grounds to believe that in the upper troposphere in the entrance region to a jet there would be ascent on the warm side and descent on the cold side. Similarly in the upper troposphere in the exit region there would be descent on the warm side and ascent on the cold side. Murray and Daniels[17] indicated that the transverse components, whose existence they had demonstrated by statistical analysis of observations, would be consistent with circulations in opposite senses above and below the jet stream. Figure 8 indicates schematically a possible association of these vertical and transverse components in the entrance and exit regions of a jet stream.

There is little direct evidence for the vertical branches of the circulations. Some indirect support for the vertical motion in the upper troposphere is found in the distributions of humidity, clouds and rainfall about the jet stream. These are discussed in Sections 7.3.2.3., 7.3.2.4. and 7.3.2.5.

Lee[19] has examined some synoptic evidence for circulations about a jet stream. Although the results were not entirely conclusive, they were not at

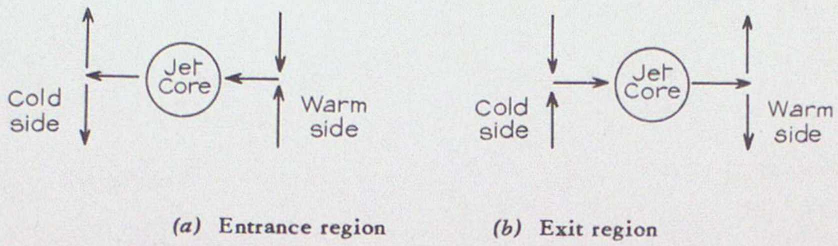


FIGURE 8 *Possible circulations, shown schematically, in sections through (a) the entrance to and (b) the exit from a jet stream (looking downstream)*

variance with Figure 8. More recently Sawyer[20] has shown that, with a number of assumptions, some deductions could be made about the patterns of convergence and divergence and associated vertical motion on the cold side of an intensifying jet stream or one which, having previously been straight, acquired cyclonic curvatures of 600 nautical miles radius at a distance 1,000 nautical miles downstream from the straight section. The approximate fields are expected to be roughly as shown in Figure 9.

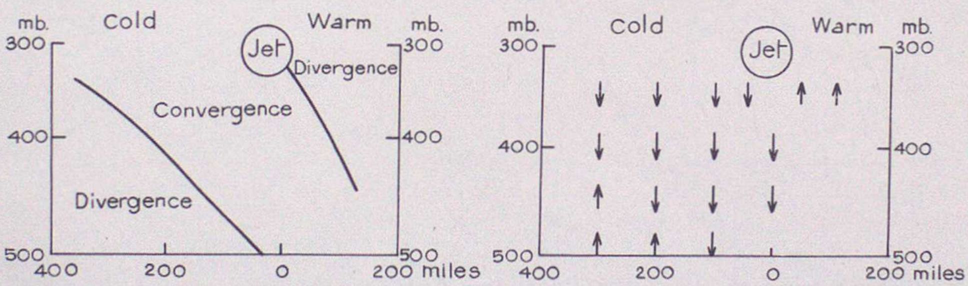


FIGURE 9 *Approximate fields of horizontal divergence and vertical motion on the cold side of an intensifying jet stream*

7.3.2.2. *Temperatures.* The broad features of the strong temperature gradients associated with jet streams can usually be determined by analysis of routine upper air data. Indeed the cross-sections shown in Figures 2 and 3 were deduced from routine observations. More detailed observations in areas close to jet streams near the British Isles have been made on a series of flights carried out by aircraft of the Meteorological Research Flight. These observations have been analysed by Murray.[21]

Although none of the flights completely traversed the broad baroclinic zone normally associated with jet streams and extending horizontally over several hundreds of miles, the temperature distributions were well established on cross-sections prepared from the aircraft observations and supplemented by the routine radio-sonde observations. Leaving aside the small-scale temperature fluctuations which are nearly always present in the atmosphere (see Chapter 14, Section 14.6) the temperature profile on the individual flights generally showed a fairly smooth trend. The individual temperature gradients per 100 nautical miles varied from 1°F. (0·6°C.) to 7°F. (3·9°C.) with a mean value of 3·7°F. (2·6°C.). A typical example of the temperature profile is shown in Figure 10. The horizontal legs of the flight were flown at 400 and 250 millibars, that is below and above the level of the jet stream.

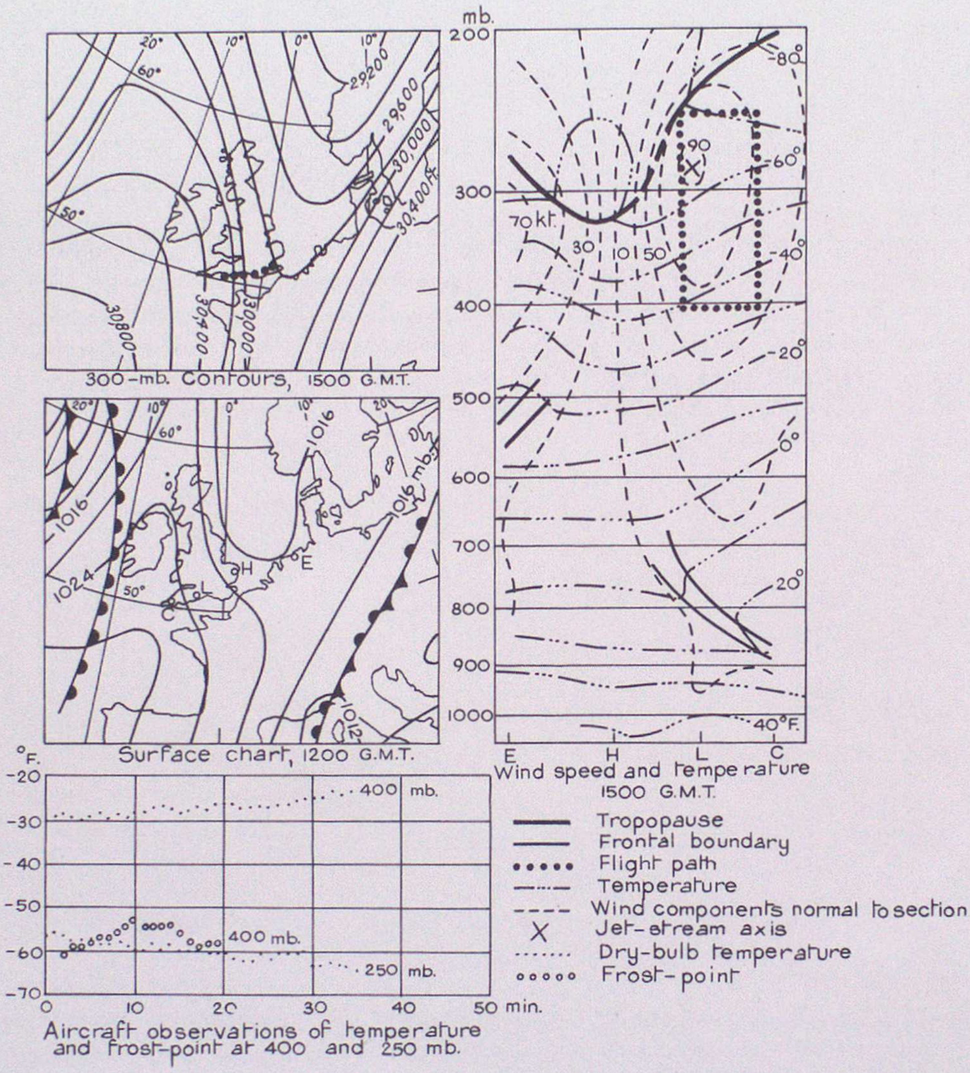


FIGURE 10 *Synoptic situation, 24 April 1952*
E = Emden, H = Hemsby, L = Larkhill, C = Camborne.

It will be noted that throughout the middle and upper troposphere there is a broad baroclinic zone extending to the level of the jet. A few thousand feet above the tropopause there is a gradient of temperature opposed to that in the troposphere.

With some jet streams the frontal zone in the upper troposphere is well marked and the temperature profile will then usually indicate a pronounced change of horizontal temperature gradient. Figure 11 shows such a profile.

The temperature profile at 400 millibars clearly shows the frontal boundary adjacent to the warm air within which the horizontal temperature gradient is

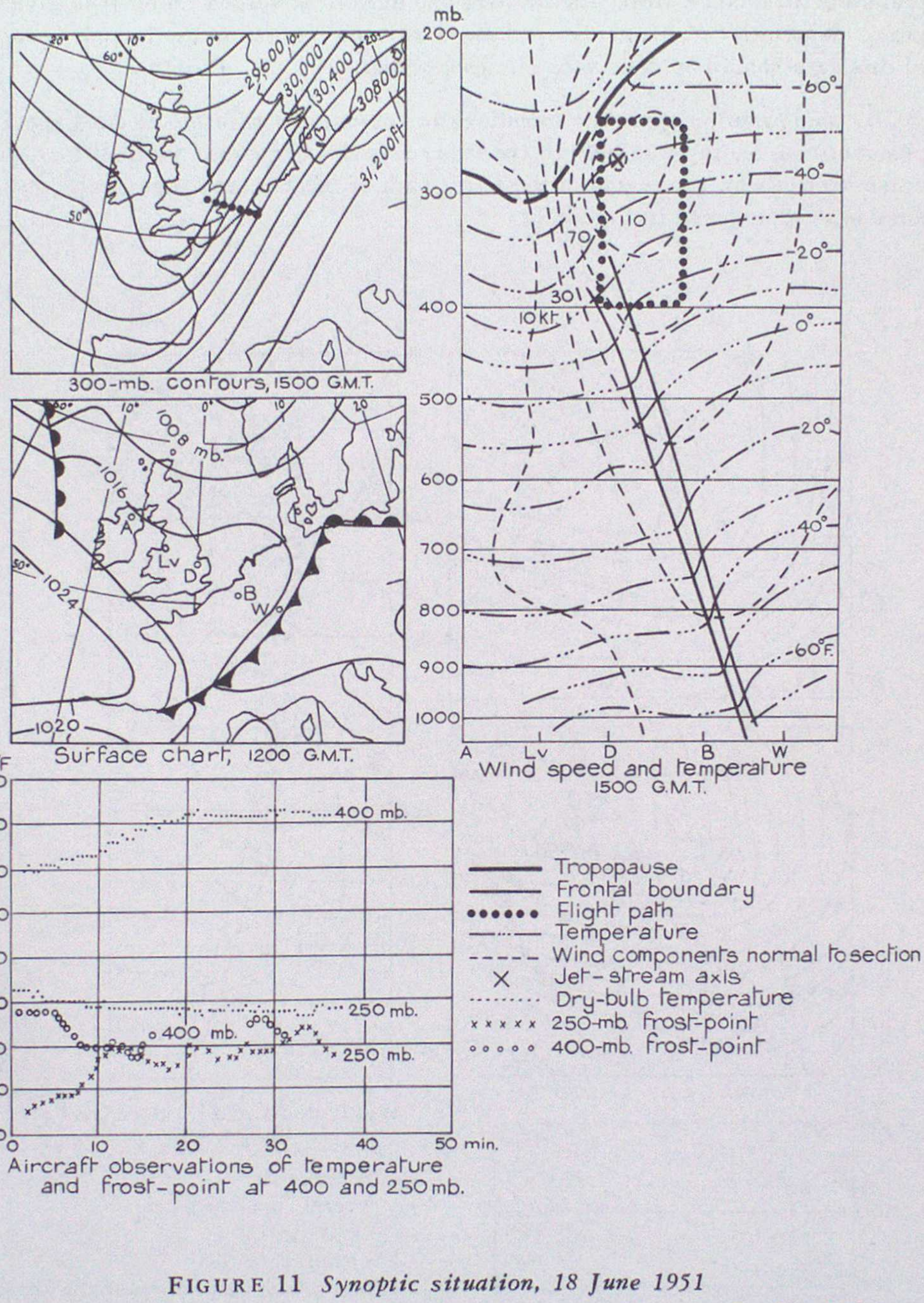


FIGURE 11 *Synoptic situation, 18 June 1951*
A = Aldergrove, Lv = Liverpool, D = Downham Market, B = Brussels, W = Wiesbaden.

smaller than that suggested on the cross-section. The reverse thermal gradient is relatively small at 250 millibars but a slight shift of the flight path for the upper leg to rather higher levels might have shown a more marked gradient.

Murray remarked that the smaller temperature gradients tended to be associated with the weaker jet streams and with flight paths nearer the level of the jet-stream axis.

7.3.2.3. *Humidities*. Murray[21] found that, in association with jet streams, humidities were more variable than temperatures. This variability detracts from the usefulness to forecasters of mean values and synoptic examples since each may be very wide of the mark in any individual case. In spite of this it seemed preferable to include some illustrations and numerical values rather than give a purely descriptive account. Nevertheless any application on the forecast bench of this data should be done with circumspection.

By using routine upper air soundings in conjunction with the aircraft special observations, Murray was able to construct for most occasions vertical frost-point cross-sections for the region of the flight path of the aircraft. One such cross-section is reproduced in Figure 12.

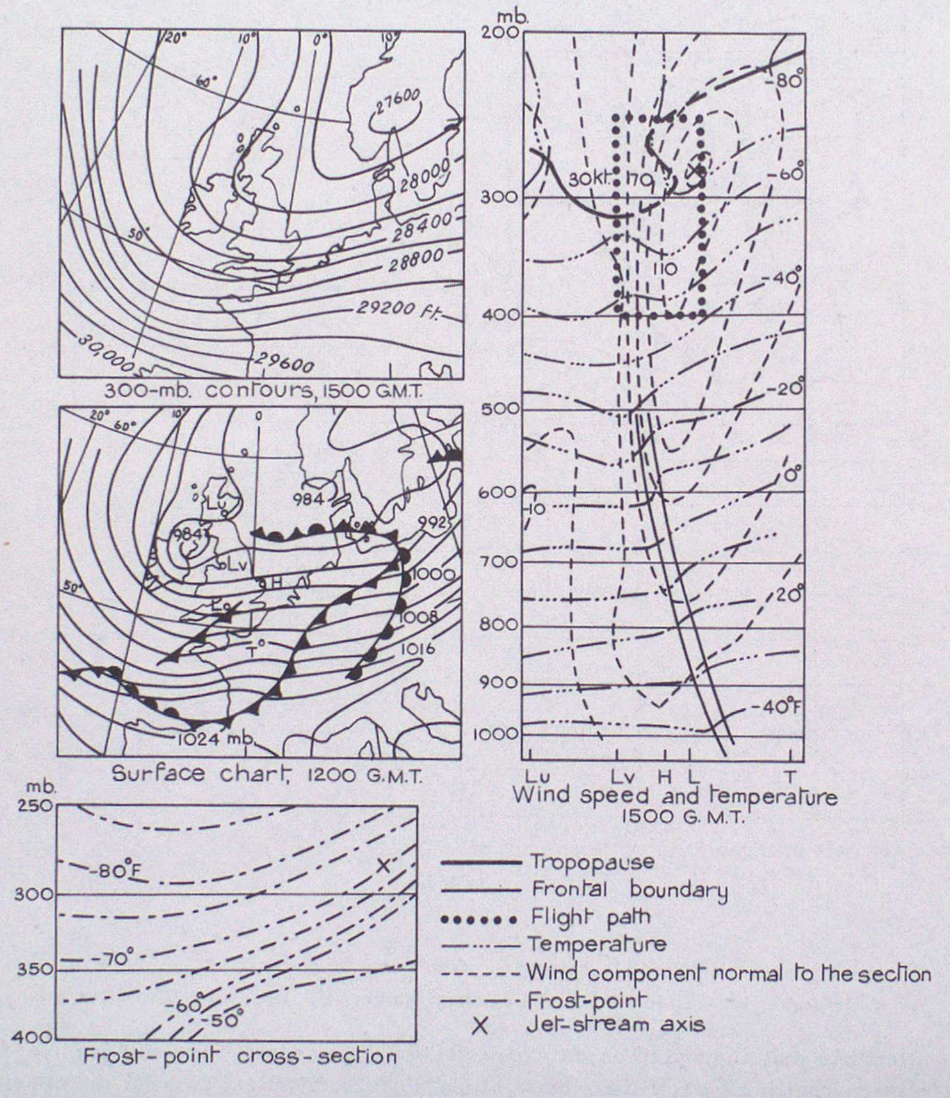


FIGURE 12 *Synoptic situation, 2 January 1952*
Lu = Leuchars, Lv = Liverpool, H = Hemsby, L = Larkhill, T = Trappes.

By using such cross-sections Murray was able to estimate the approximate values of the temperature, frost-point and frost-point depression at the jet-stream axis. The mean values were –60°F. (–51·1°C.) for temperature, –79°F. (–61·7°C.) for frost-point and 19°F. (–10·6°C.) for frost-point depression. However, the cross-sections indicated that there were great differences in humidity at the jet-stream axis from case to case, and that the air in the core might be practically saturated or quite dry. There appeared to be some association between a high value of the frost-point depression (that is low relative humidity) and high temperature. For example, out of the eleven occasions for which it was possible to estimate approximate values for temperature and frost-point at the jet axis, Murray found that the five jet streams with relatively warm air at the position of the jet axis had a mean temperature of –50°F. (–46°C.) and a mean frost-point depression of 32°F. (17·8°C.); in contradistinction the other six cases averaged –68°F. (–56°C.) for temperature and 9°F. (5°C.) for frost-point depression in the core of the current. The jet streams with a relatively moist core and those with a relatively dry core did not appear to have any particular seasonal preferences, or be related to broad synoptic features such as the forward side of a long-wave trough or ridge, or be connected with the lateral motion of the axis. However, the smallness of the sample made it uncertain whether these results were truly representative of jet streams.

The frost-point cross-section shown in Figure 12 indicates a horizontal gradient of frost-point in the same direction as the pressure gradient and this type of frost-point distribution was indicated on many of the flights (but a contrary distribution was indicated on at least one flight). Some mean values are given below. The mean horizontal variation of temperature and frost-point relative to the values at the tropopause are shown in Figure 13 which shows a considerable difference in humidity between the troposphere and stratosphere at the same level.

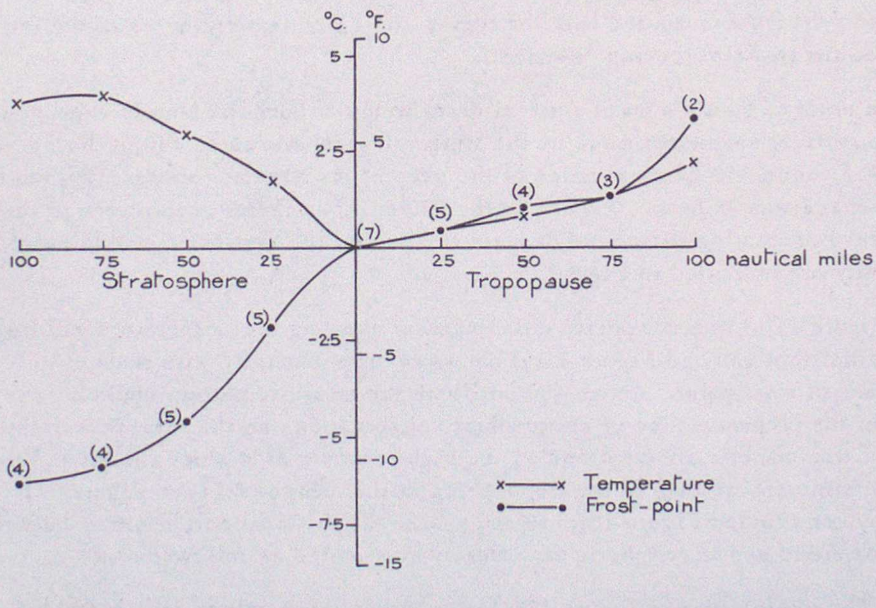


FIGURE 13 *Mean temperature and frost-point relative to the values at the tropopause based on seven horizontal traverses through the tropopause*

Mean tropopause temperature = –64.4°F. (–53.6°C.)

Mean tropopause frost-point = –75.0°F. (59.4°C.)

The number of observations at various distances are in parentheses.

Murray used the aircraft observations, taken on flights which were made wholly in the troposphere and traversed the jet stream, to construct a humidity profile across a larger extent of the jet stream than that shown in Figure 13. Separate profiles for type W and type C jets were constructed but, as these profiles did not appear to be radically different from each other, they are not reproduced here. The horizontal profile of mean frost-point depression for all jet streams investigated is given in Figure 14. This displays a gradient of depression of the frost-point in the upper troposphere across the jet stream from high to low pressure. This is consistent with a gradient of humidity from high to low pressure.

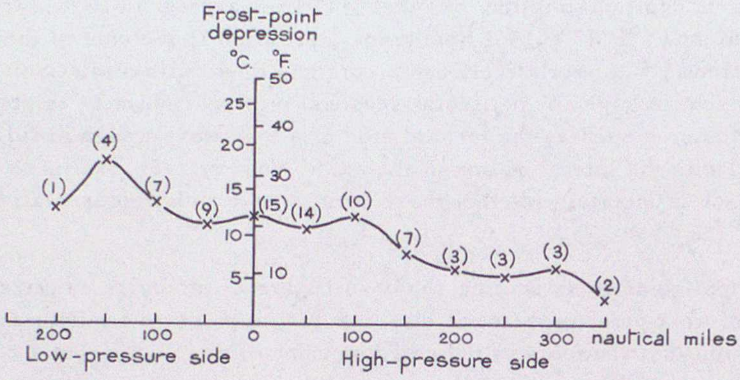


FIGURE 14 ***Profile of mean frost-point depression across jet streams in the upper troposphere***

The number of observations at various distances from the jet-stream axis are shown in parentheses.

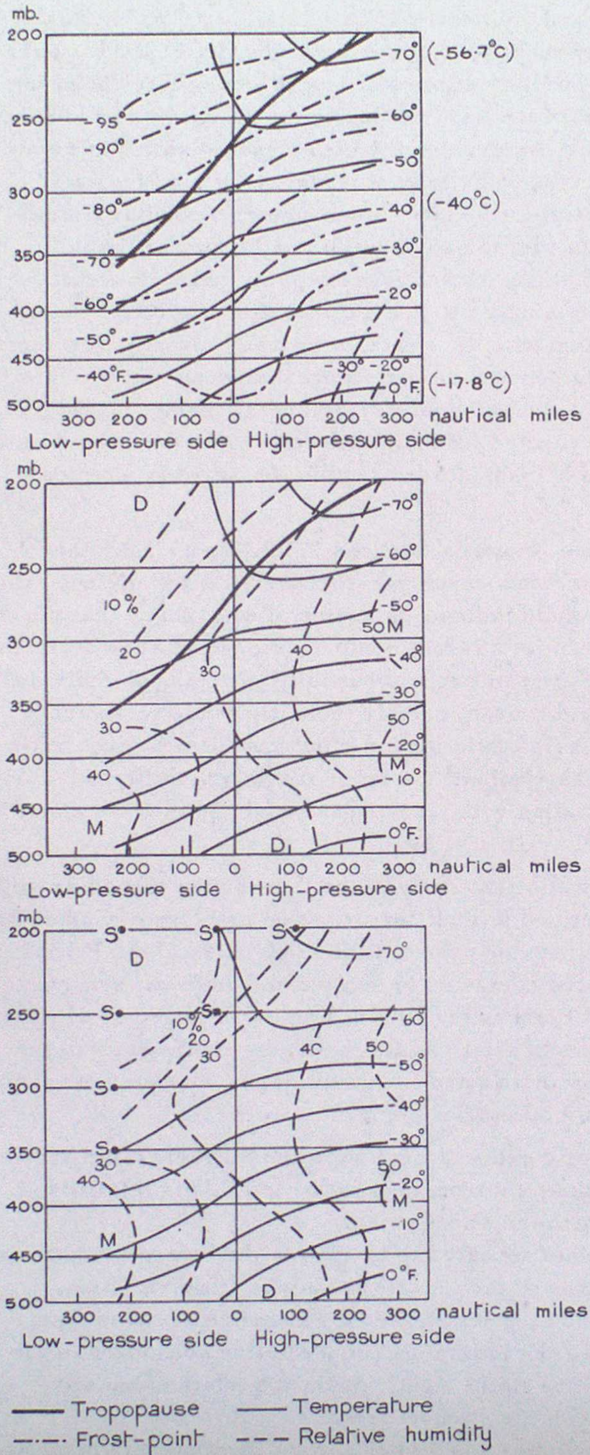
The bulk of the evidence indicated that there is a definite tendency for the air to be moister, as measured by higher frost-points, in the region 100 - 200 nautical miles to the right compared with the region 100 - 200 nautical miles to the left of the jet-stream axis, looking downwind.

In order to obtain a mean vertical distribution of humidity Murray examined those vertical soundings made by the Meteorological Research Flight during 1949-51 within 350 nautical miles of the axes of jet streams (occasions when the wind speed was at least 70 knots on the 300-millibar charts constructed at the Central Forecasting Office). The main features of the vertical distribution of humidity are indicated in Figure 15.

Figure 15*(a)* is a composite cross-section showing the temperature and frost-point distributions and Figure 15*(b)* shows relative humidity with respect to ice in place of frost-point. Above 400 millibars the humidity pattern could be a reflection of the preponderance of stratospheric observations on the low-pressure side and of tropospheric observations on the high-pressure side since the air is drier in the stratosphere than in the troposphere at the same level (see Figure 13). Murray constructed Figure 15*(c)* to emphasize the difference in humidity between stratospheric and tropospheric air. Murray commented as follows:

"The top left part of Figure 15*(c)* is based on mean values from individual soundings entirely in the stratosphere (marked S); the remainder of this figure refers to tropospheric conditions. The criterion for deciding that the points S were in the stratosphere was simply the existence of a majority of stratospheric observations at these points – the tropospheric observations at the same points being neglected in evaluating the mean temperature and humidity; a similar criterion was applied to the tropospheric points. Figure 15*(c)* shows that the main part of the horizontal gradient of humidity which is indicated in the upper parts of Figure 15*(a)* and 15*(b)* is concentrated near the tropopause.

Some Features of the Free Atmosphere



(a) Mean temperature and frost-point, making no distinction between stratospheric and tropospheric observations.

(b) Mean temperature and relative humidity with respect to ice, making no distinction between stratospheric and tropospheric observations.

(c) Mean temperature and relative humidity with respect to ice. Means based on stratospheric observations only are marked S, all other observations are in the troposphere.

FIGURE 15 *Mean cross-sections showing the humidity distribution near a jet stream*

D: relatively dry region

M: relatively moist region

There are clearly two main features of the average humidity distribution shown in Figure 15:

(i) Above about 400 mb. the air tends to be relatively dry on the low-pressure side and relatively moist on the high-pressure side of the jet stream. The horizontal gradient of humidity is probably, in the main, a result of the fact that, at about the level of the wind maximum, the air on the low-pressure side is usually stratospheric, whereas on the high-pressure side it is tropospheric. Nevertheless, there appears to be a tendency for a horizontal humidity gradient also to exist in the upper troposphere; a similar feature has been noted in connexion with the analysis of the 20 special flights.

(ii) At or below about 500 mb. a relatively dry patch tends to occur roughly beneath the jet-stream axis. It is clear that this relatively dry patch has a close association with the frontal zone which usually intersects the 500-mb. surface directly beneath the jet-stream maximum. Sawyer[1] has confirmed the existence of a dry patch near many frontal zones; he suggests that its presence can be accounted for by the advection into the frontal zone of air which has been dried by subsidence at an earlier stage in a region external to the frontal zone."

7.3.2.4. *Clouds.* It has been shown in Sections 7.3.2.2 and 7.2.2.3 that the mean distributions of temperature and humidity associated with jet streams exhibit characteristic patterns. This would indicate in a general sort of way that the distributions of clouds in relation to jet streams would, in the mean, show features which were consistent with the temperature and humidity patterns. In individual jet streams the deviations from the mean, notably those for humidity, indicate that there would probably be similar variations in cloud conditions. Such variations are amply confirmed by observations. In spite of this variability the following account of clouds in association with jet streams should be of some practical value.

Sawyer and Ilett[22] made a statistical study of the distribution of medium and high cloud near jet streams (defined in their investigation as being a wind of 70 knots or more at 300 millibars) passing over or within 500 miles of the British Isles during 1949. The data used consisted of surface and upper air synoptic reports received at the Central Forecasting Office, Dunstable, together with the analysed charts. For the details of their results the reader should refer to the original paper which was widely distributed to outstations. The more noteworthy results of their investigation are summarized below:

(a) Amounts of cirrus of 4 oktas or more were considerably more frequent to the right of the axis of the jet stream than to the left. This relationship was slightly more pronounced with the stronger jets.

(b) Medium-cloud amounts appeared to be less in the left entrance to the jet. Amounts in other regions of the jet stream showed little variation.

(c) The statistics of cirrus types showed a dominance of cirrus type 3 (anvil cirrus) to the left of the jet stream and a relative abundance of the frontal and layer type of cirrus to the right. Again the relationship was slightly better developed with the stronger jets.

(d) The statistics of medium-cloud types showed a remarkable predominance of type 6 (altocumulus formed by the spreading out of cumulus) to the left of the jet stream. Altostratus, frontal type of altocumulus and altocumulus castellanus were more common to the right of the jet.

Murray[21] has examined the distribution of cloud as observed on twenty flights made by the Meteorological Research Flight especially to investigate the jet stream. The traverses of the jet stream were normally made at two levels on each flight, the lower level being generally between 400 and 350 millibars (but one at 325 millibars) and the upper level between 275 and 200 millibars (the majority at 250 millibars). The following account is taken entirely from Murray's paper.

The percentage of time during which the aircraft was in cloud whilst carrying out the horizontal traverses is shown in Table 5. The total flying time on the horizontal legs was some 26½ hours. Table 5 shows that clouds were traversed quite infrequently on horizontal flights.

TABLE 5 *Duration of flight in cloud on horizontal legs*

Type of cloud	*Position of horizontal legs relative to jet-stream axis*		
	above	*below*	*above and below*
	percentage of flying time		
Cirrostratus, cirrus and anvil cirrus	0	2.9	1.8
Nimbostratus, altostratus, altocumulus	0	3.6	2.2
All cloud	0	6.5	4.0

On the sixteen legs which were located above the jet-stream axis no cloud was traversed and on only three legs were cirrus or cirrostratus patches observed above the aircraft. The cirrus type of cloud was observed at distances varying between about 300 and 400 nautical miles from the jet-stream axis on the high-pressure side in the upper troposphere; no cloud was observed at or above the level of the jet-stream axis on the low-pressure side. It seemed clear that the atmosphere above the jet-stream axis was generally cloudless, apart from occasional patches of cirrus or cirrostratus on the high-pressure side in the uppermost part of the troposphere.

Cloud was traversed on only seven legs of the twenty flights. The one report of cumulonimbus anvil cloud occurred just over 200 nautical miles from the axis on the low-pressure side of the jet stream, below which region a good deal of convection was taking place. One leg traversed delicate cirrus practically vertically below the jet-stream axis. On the remaining five of these seven legs the aircraft traversed cloud of the layer type, that is nimbostratus, altostratus, altocumulus and cirrostratus, which occurred at distances from the jet-stream axis varying from about 70 nautical miles on the low-pressure side to 400 nautical miles on the high-pressure side.

On seven flights no cloud was observed at or above the bottom leg of the flight. In each case there was usually broken stratocumulus or cumulus below the aircraft and, on one or two occasions, some local altocumulus. However, these legs were not located symmetrically about the jet axis but were biased to the low-pressure side so that cloud amounts would probably be biased to low values. Although there was this bias Murray thought it significant that in six of these cases no surface front accompanied the jet stream and in five out of these six cases the typical sloping baroclinic zone appeared linked with an inversion or very stable layer of the subsidence type in the lowest few thousand feet.

In contrast with the seven flights with no cloud at or above the lower leg (400 - 350 millibars) the remaining thirteen flights reported some cloud. Seven flights reported cloud at the lower level and six reported cloud above the lower level. Excluding the nimbostratus, altostratus and altocumulus cloud actually traversed and already discussed, all clouds observed on these legs were of the ice-crystal type, that is cirrus or cirrostratus, or on one occasion with some cirrocumulus in addition. Excluding some cumulonimbus anvil cloud traversed at just over 200 nautical miles from the axis on the low-pressure side of the jet stream, the high cloud was observed from about 70 nautical miles on the low-pressure side to just over 400 nautical miles on the high-pressure side of the jet-stream axis, that is predominantly on the high-pressure side. Although the jet streams could not readily be placed in any clear synoptic category Murray considered that "the most notable feature associated with these particular jet streams was the existence of a surface front in all but one or two cases, although more than half the fronts were quite weak in terms of rainfall activity."

From the observational evidence obtained on all the flights Murray reached the following conclusion about clouds above 400 millibars in association with jet streams:

> "(i) generally no cloud (except occasionally cumulonimbus tops or anvils) occurs on the low-pressure side within the region between about 250 and 100 nautical miles from the jet-stream axis,
>
> (ii) some cloud generally occurs on the high-pressure side between about 150 and 450 nautical miles from the jet-stream axis,
>
> (iii) from about 100 nautical miles on the low-pressure side to about 150 nautical miles on the high-pressure side there may be nil or any amount of cloud."

It should be noted that on most of the flights some cloud was generally observed below about 400 millibars. Murray stated that: "As regards low cloud, it was predominantly cumulonimbus or cumulus or stratocumulus on the low-pressure side, and stratocumulus or nimbostratus on the high-pressure side of the jet-stream axis, but amounts were variable from case to case; such a distribution of low cloud was generally supported also by the surface synoptic observations, and was not unexpected. Medium cloud tended to be much more common when the region explored was on the high-pressure side of the jet-stream axis, as with cloud above 400 mb."

The distribution of layer cloud (excluding low cloud) near jet streams appeared to fit into a rough pattern and some instructive histograms were prepared. These are reproduced in Figures 16 and 17.

Figure 16 shows the percentage frequency of occurrence of layer cloud (whatever the amount) within the various sectors. Figure 16*(a)* is based on reports of nimbostratus, altostratus and altocumulus cloud. The observational data are presented rather differently in Figure 17, which shows for each sector the observed cloud amount expressed as a percentage of the maximum amount of cloud which would have been observed if the flights had reported layer cloud all of the time, either above or below the track.

Murray comments: "Despite the smallness of the sample upon which the histograms of Figures 16 and 17 are based, all tell a consistent story. The placing of cloud observations on individual flights in their correct position relative to the jet-stream axis was uncertain in some degree; but there is no doubt of the reality of the overall distribution of layer cloud portrayed in Figures 16 and 17. It is quite clear that layer cloud is mainly a feature of the high-pressure side of jet streams, although great variations in individual cases make one hesitant to say that the average picture is the typical model in any but broad respects.

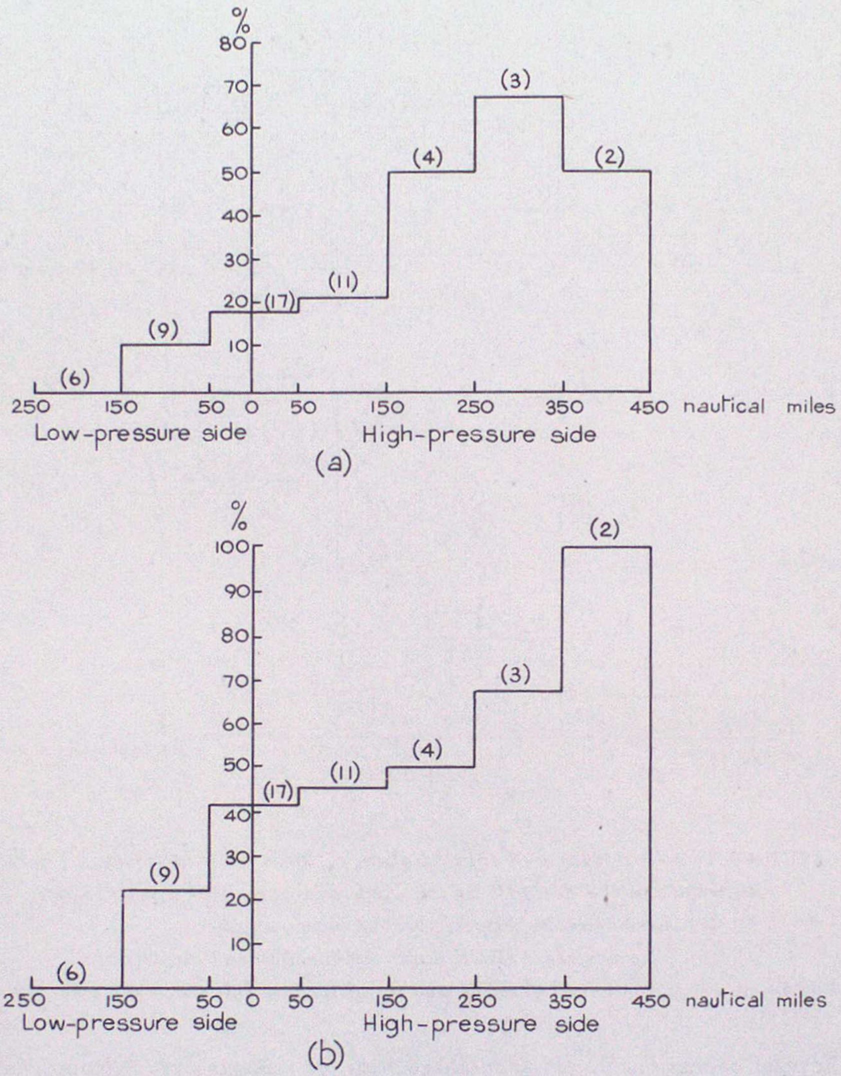


FIGURE 16 *Percentage frequency of occurrence of layer cloud in the various sectors of a jet stream*

(a) Nimbostratus, altostratus and altocumulus

(b) Nimbostratus, altostratus, altocumulus and cirrostratus

The figures in parentheses give the number of flights which investigated part or all of each sector.

Examination of the surface synoptic and the aircraft observations suggests that the jet streams with extensive layer cloud mainly on their high-pressure side are usually associated with a surface front which may be either active or rather inactive in terms of weather; such jet streams may be located on the forward side of either an upper trough or an upper ridge. Moreover, the jet streams with little or well broken layer cloud on their high-pressure side are normally either not associated with a front on the surface chart (a low-level subsidence inversion usually occurs where a front might be expected) or the front is very weak; these jet streams occur more frequently on the forward side of an upper ridge than on the forward side of an upper trough."

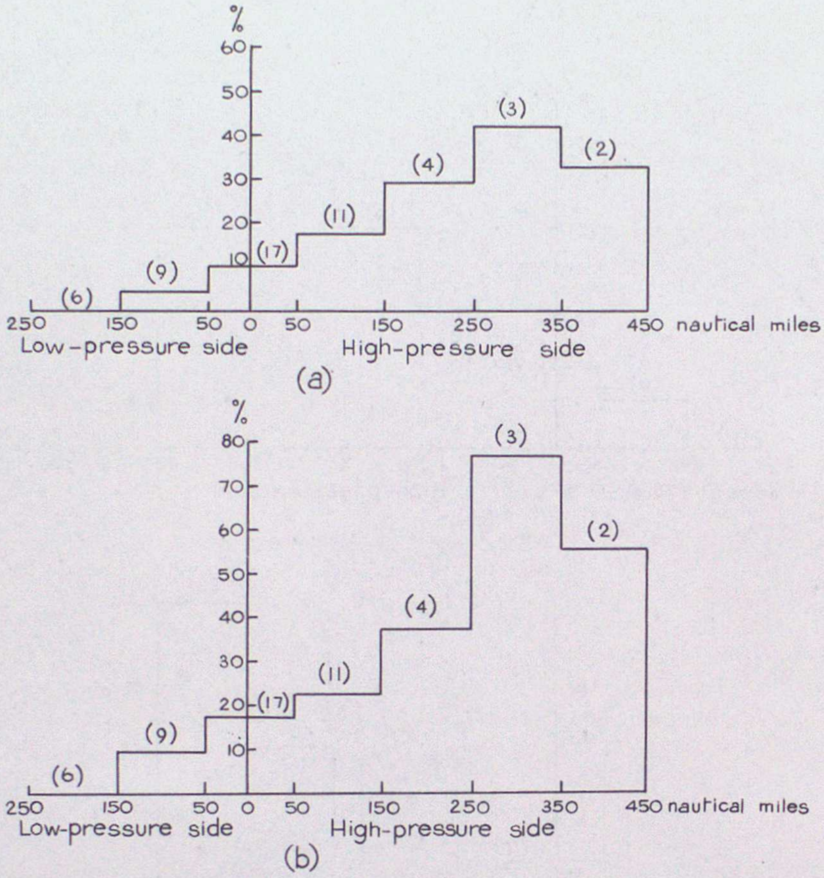


FIGURE 17 *Percentage of total distance flown with layer cloud above, at or below the aircraft in the various sectors of a jet stream*

(a) Nimbostratus, altostratus and altocumulus

(b) Nimbostratus, altostratus, altocumulus and cirrostratus

The figures in parentheses give the number of flights which investigated part or all of each sector.

Several papers on clouds associated with jet streams over the north-eastern part of the United States of America have been prepared by Schaeffer and his colleagues. References to some of the papers published in 1953 and 1955 will be found in *World Meteorological Organization Technical No. 19.*[7] Some of Schaeffer's studies are based on observations of clouds as the jet streams moved across the observer's fixed position and include some picturesque and illustrative photographs (some in colour) of jet-stream clouds.

7.3.2.5. *Rainfall.* Since there is often an association between mid-latitude jet streams and the polar front, it seems likely that some relationship between rainfall and the jet streams should also exist. Some further support for this is indicated by the fact that a jet stream associated with long-wave synoptic patterns may often act as a steering current for frontal and non-frontal short-wave synoptic patterns. However, as short-wave patterns evolve and move through the long-wave pattern the relative position of the short-wave feature to the axis of the jet stream often changes so that these changes may cause both variability and a skewness in the relation of precipitation to the jet stream. In Section 7.3.2.1 it was indicated that there are grounds to believe that in the upper troposphere there is often upward motion to the right and downward motion to the left of the confluence at the entrance to a jet stream (and conversely downward to the right and upward to the left of the diffluence at the exit). The existence of any such systematic vertical motion would probably also affect the distribution of rainfall.

Johnson and Daniels[23] have examined some rainfall data associated with jet streams for some four stations in the British Isles. The authors selected Scilly, Valentia, Tiree and Lerwick in the hope that local effects on rainfall amounts would be small but such effects were not negligible. The detailed results are not reproduced in this handbook but the general conclusions of Johnson and Daniels should be of interest to practical forecasters and are given below.

Precipitation was distributed relative to the jet stream according to a fairly definite pattern and, in particular:

(i) in the entrance, twice as much rain occurred to the right of the axis as to the left,

(ii) in the central region the amounts to the right and left of the axis were practically equal,

(iii) in the exit there was twice as much rain to the left as to the right of the axis.

The data for the central and exit regions (but not for the entrance) were sufficient to permit a further classification according to the character of the precipitation. Three classes were formed: *(a)* when only showers were reported, *(b)* when both rain and showers were reported for the same twelve-hour interval and *(c)* when rain alone was reported.

In the central region the distributions of amounts due to rain and amounts due to showers were very different. Shower-type precipitation increased towards the left of the flow but amounts due to rain-type precipitation tended to be greater to the right of the axis than to the left.

In the exit region rain-type precipitation increased markedly towards the left of the flow. In addition shower-type precipitation increased more sharply from right to left in the exit than in the centre.

7.3.2.6. *Turbulence.* Turbulence can be experienced by aircraft flying at all levels in the troposphere and in the last two decades there has been an increasing interest in turbulence encountered in clear air at upper levels near the tropopause. Several investigations of this clear-air turbulence have been made and it is evident that at least on some occasions there is an association between clear-air turbulence and the jet stream. The relationships between clear-air turbulence and jet streams will be more fully discussed in Chapter 20 and are not further discussed in this chapter.

7.3.2.7. *Association with synoptic features.* At the outset it must be stated that there is no one-to-one correspondence which can always be relied upon to exist between mid-latitude jet streams and synoptic features. Nevertheless the relationship between an individual jet stream and the associated synoptic systems often presents a coherent picture at least over a day or so. Further, the variation in such a relationship often tends to occur in a systematic and relatively steady manner in accordance with the evolution of the synoptic situation. Accordingly it should serve a useful purpose to describe a few associations between jet streams and synoptic features. This should enable the forecaster engaged on practical work to recognize the associations readily when they occur and also to pick out the important features of the particular association which exists between a jet stream and the synoptic systems in any individual situation. When such an association is reliably identified some reliance may be placed on its continuance and gradual modification when forecasting for a few hours ahead.

Firstly it is important to realize that on the broad scale the jet stream shows a similarity in orientation to the long waves. As discussed in Chapter 5 the long waves vary from the nearly straight westerly flow to the marked meandering patterns associated with a meridional type of circulation characterized by marked incursions of deep layers of warm air polewards and of cold air equatorwards over substantial bands of longitudes. As a result of these northward and southward movements of warm and cold air, extensive warm ridges and cold troughs are formed throughout the greater part of the troposphere in those longitudes. When the thermal contrast between such ridges and troughs is both well marked and concentrated, jet streams will occur. Thus jet streams are closely allied to long-wave patterns. As with the long-wave patterns, well established jets usually move relatively slowly at right-angles to their axes but the air moves through the pattern at great speed, undergoing marked acceleration as it enters a jet and marked deceleration on leaving. Large-scale changes in the jet-stream pattern usually take place only slowly in association with related changes in the major synoptic systems. Such changes may well take place over periods of days rather than hours.

Smaller-scale synoptic features affect jet streams over limited areas. It is generally found that a front with a well marked thermal contrast usually has an attendant jet stream but jet streams may occur in the absence of well marked fronts. The core of the jet streams associated with a strong front is normally located in the uppermost part of the warm troposphere almost directly above the intersection of the front with the 500-millibar surface. Murray and Johnson[12] remarked that: "This empirical rule is a fairly good approximation to reality in cases of strong jets and fronts. The rule implies that the jet-stream axis should often be 200 mi to 400 mi behind a surface cold front and 400 mi to 800 mi in advance of a surface warm front. In many cases much greater complexity exists in the relationship between surface fronts and jet streams."

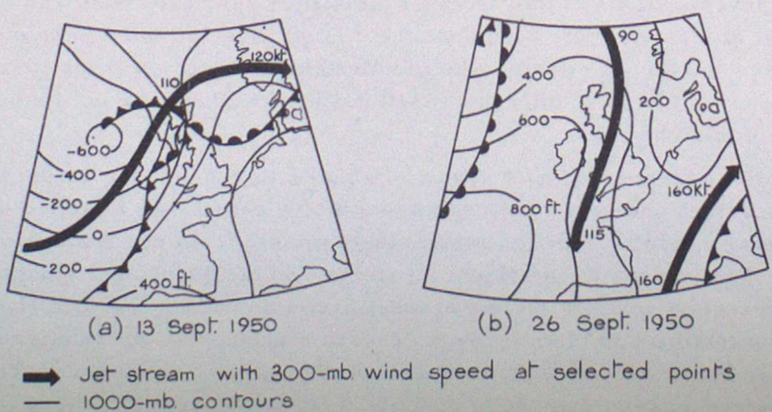


FIGURE 18 *Examples of association between jet streams and surface fronts, 13 and 26 September 1950*

Jets are parallel to the surface fronts usually over a limited distance only. In general jets tend to approach that part of the front which is near to a cyclonic centre and to recede from the other end of the front. Figure 18*(a)* (taken from Murray and Johnson's paper,[12] as are many other illustrations in this sub-section) shows the fronts and jet stream in positions which accord quite well with the preceding statements. Figure 18*(b)* shows surface fronts and jet streams almost parallel over considerable lengths of the fronts.

Murray and Johnson further commented that small running waves on a slow-moving cold front affect the associated jet hardly at all. If the wave develops, the jet becomes distorted in sympathy with the increasing amplitude of the wave-like surface disturbance. As the occluding process proceeds, the jet tends to weaken in the neighbourhood of the surface centre, perhaps lying almost perpendicular to the surface occlusion, as in Figure 19*(a)*. On other occasions the original jet separates out into two distinct jets – a type C and a type W jet (see Section 7.3.2.1 for distinction between type C and type W jets) – as in Figure 19*(b)*. Warm and cold occlusions having the character of warm and cold fronts may sometimes be associated with type W and type C jets respectively.

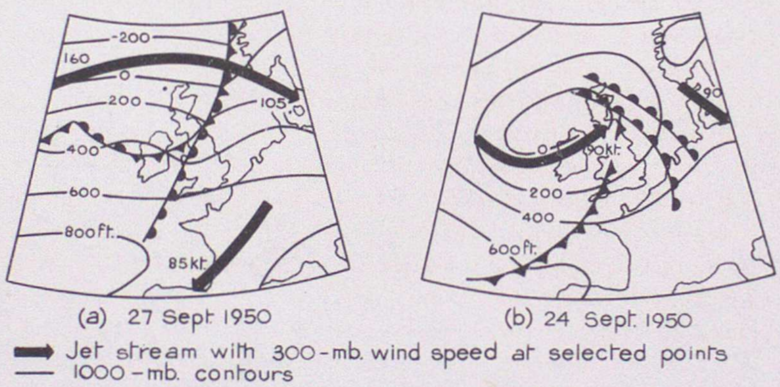


FIGURE 19 *Examples of association between jet streams and surface fronts, 27 and 24 September 1950*

These remarks are in the nature of broad generalizations. On occasion there appears to be little or no association between surface fronts and jets, as in Figure 20. This state of affairs is likely to arise where the surface fronts are thermally weak, or when a new strong jet is overtaking an older degenerating frontal system, as in Figure 20. The remnants of degenerating jets may also complicate the picture. These complexities are inevitable in an atmosphere essentially dynamic in its behaviour. However over a short period, say of the order of twelve hours, there is considerable conservatism in the relationship of the jet with its front; this fact may be of some use in short-range forecasting. For instance, different type W jets may be differently aligned in relation to warm fronts but a particular relationship, once established, is likely to persist for at least twelve hours.

The changes in a jet stream which occurred over a period of 48 hours, as a wave developed on a stationary cold front and subsequently deepened and moved north-north-east, are shown in Figures 21*(a)*, *(b)* and *(c)*. On Figure 21*(a)* the

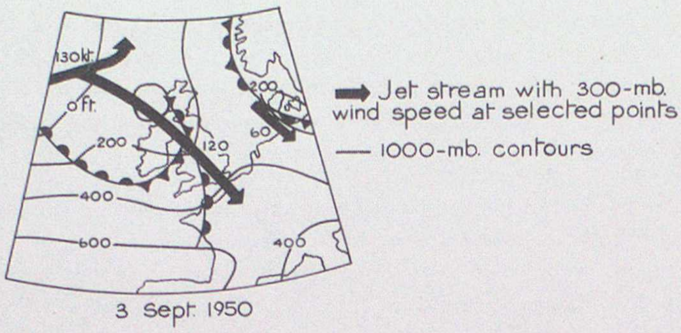


FIGURE 20 *Example of association between jet streams and surface fronts, 3 September 1950*

jet stream appears as slightly concave towards the north. Figure 21*(b)* shows that, 24 hours later, there has been little change of position over the British Isles but the jet is now convex towards the north and is also somewhat weaker. From Figure 21*(c)* it can be seen that a further 24 hours later the jet has become still weaker, has moved north in advance of the warm front and south-east in the rear of the cold front. These movements have resulted in the jet stream taking a shape which resembles part of a sine curve.

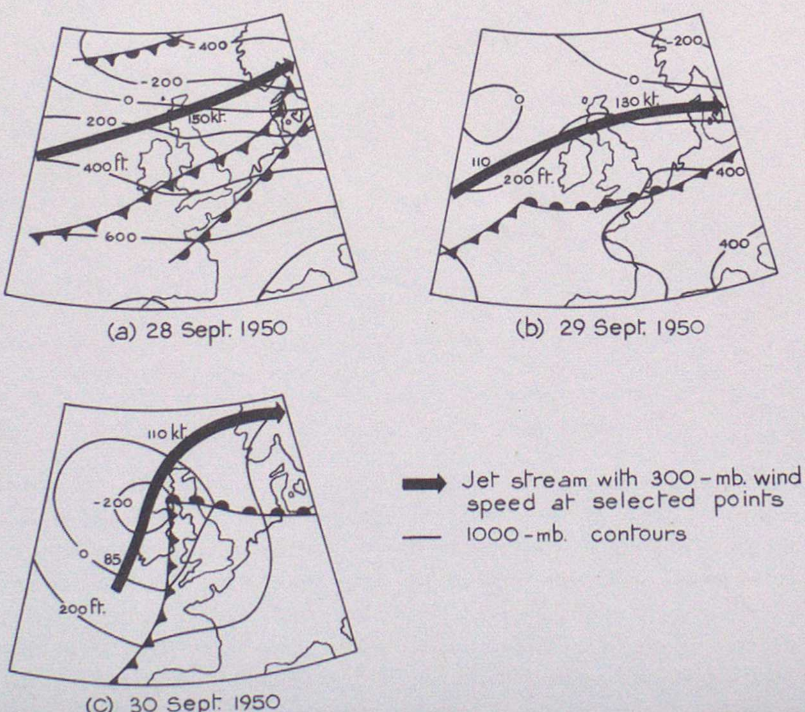


FIGURE 21 *Examples of association between jet streams and surface fronts, 28 – 30 September 1950*

A north-westerly jet stream lying across the British Isles in advance of a warm frontal system approaching from the west is shown in Figure 22. In such situations the axis of the associated thermal ridge is well to the west of the British Isles. The double warm front shown in Figure 22 is not a characteristic of this type of situation; in many cases there is only a single surface front.

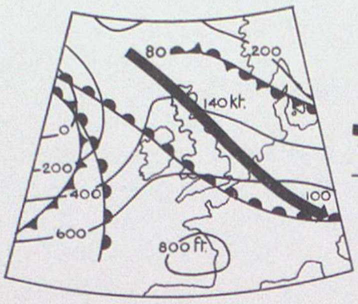


FIGURE 22 *Example of association between the jet stream and surface fronts, 23 September 1950*

Figure 18*(b)* illustrates a case where the jet stream ahead of an advancing warm front was much more northerly. The sequence of events after Figure 18*(b)* can be seen in Figures 19*(a)* and 21*(a)*, *(b)* and *(c)*. These figures illustrate that the weather was unsettled and the synoptic systems were mobile. It is worth remarking that at times a northerly or north-easterly jet stream is established over or just to the east of the British Isles in association with a blocking anticyclone to the west. The charts usually then exhibit a pronounced meridional upper flow across western Europe in association with a pronounced cold trough extending to the Mediterranean area, and in some respects resemble Figure 18*(b)*. There are, however, important differences. In the blocking situation, the axis of the cold trough is usually somewhat farther east than that shown in Figure 18*(b)*. In addition the surface isobaric ridge to the west is usually much more developed, often with its axis situated farther west but with an almost north – south orientation and there may well be a separate anticyclonic cell centred somewhat north of 50°N. The upper warm ridge to the west is also generally rather more developed and is orientated in a more north – south direction than that implied in Figure 18*(b)*.

Figures 18 to 22 have shown a few actual patterns and associations of jet streams, fronts and synoptic systems. In practice the patterns and combinations of patterns are infinitely varied and little useful practical purpose would be served by including a large number of additional illustrations. The information of a general nature which has already been given should be sufficient to enable practising forecasters to analyse occasions of jet streams with understanding and to make reasonable short-term forecasts.

Where the jet stream shows a close association with a synoptic feature it tends to move laterally with that feature, except for shallow non-developing waves on fronts and in the later stages of development of a deep cyclone when the cold front sometimes penetrates well south. In this latter case the jet stream may not move south in sympathy with the plunging cold front and the separation between jet and front thereby increases. In some cases the lower tropospheric cold air subsides and a new frontal link may be formed to the north of the old one and fairly close to the jet-stream position. It has been remarked in Chapter 5 that, during the development of a deep cyclone, the initial wave often forms first on the warm side of the zone of large-scale temperature contrast or jet stream and during the development and decaying stages moves progressively to the cold side of the current.

Little has so far been said about variations in wind speed which occur in jets as can be readily seen by isotach analysis (see Chapter 13, Section 13.11.1). The patterns of strong wind exhibit some persistence and coherence. In the short term, their movement is also systematic. Once they are reliably inferred some confidence may be placed in extrapolation for a few hours in conformity with the general long-wave pattern and the expected shorter-term synoptic developments. These variations in wind speed are due primarily to variations in thermal contrast in the troposphere and are not directly related to changes in isobaric gradients on the surface map although, in well marked cases, strong surface winds may boost a jet stream. The areas of peak winds often assume a characteristic shape, being very approximately an ellipse of large eccentricity, the major axis lying more or less along the axis of the jet stream in that area. These areas of peak winds usually move downstream, speeds being variable from case to case. The speed of movement is usually much less than the speed in the jet stream and 30 knots might be typical of many movements. In *World Meteorological Organization Technical Note* No. 19[7] it is stated that during the development of a strong trough in the upper troposphere, which is often associated with surface cyclogenesis, the strongest winds are characteristically on the western side of the developing trough. This zone of strongest winds migrates through the equatorward side of the trough as development proceeds and in the final stage is often found on the eastern side.

The length of jet streams is quite variable. Jet streams can sometimes be traced as an unbroken line on synoptic charts over many hundreds or one or two thousands of miles. On other occasions the jet streams may appear as a number of relatively short broken segments – the intervals, however, usually having fairly strong winds if not quite of jet-stream speed. The extent of a jet stream along the current is seldom a serious problem on the forecast bench.

Although there is a close association between some jet streams and fronts, the fronts do not necessarily exhibit strong features uniformly well marked throughout the troposphere. For example, with some jet streams the front at the surface and in the lower troposphere may be very well marked but the front in the upper troposphere may be quite weak and scarcely distinguishable on the routine analyses. With other jet streams, however, the fronts in the upper troposphere may be strong and well marked whilst those in the lower troposphere are weak and ill-defined.

Sawyer[20] has made some theoretical considerations of the association between jet streams and frontal zones. By taking a reasonable horizontal profile of wind and introducing dynamical considerations together with some simplifying assumptions of an approximate nature, Sawyer deduced that it was reasonable to expect the approximate patterns of horizontal divergence and vertical motion on the cold side of an intensifying jet which were reproduced in Figure 9 (Section 7.3.2.1).

Sawyer comments: "The region of horizontal divergence below and to the cold side of the jet stream will bring about an intensification of the vertical gradient of potential temperature (i.e., it will increase the vertical stability) in this region as is observed in the 500 mb frontal zone. At the same time the tilting about a horizontal axis will increase the horizontal temperature gradient, and this will be accompanied by an increase in the cyclonic vorticity about a vertical axis .... The treatment is probably adequate to show that a jet stream which has a strong shear on its cold side, and flows from a region of relatively low speed to one of relatively higher speed, must develop a region of general subsidence on its cold side which is accompanied by an intensification of both horizontal and vertical gradients of potential temperature such as is observed in 500 mb frontal zones.

"Similar conclusions would follow regarding regions where the jet stream flows into a slow-moving upper trough although the frontal zone would then be built up at a smaller horizontal distance from the jet axis.

"This dynamical process provides no explanation as to why the boundaries of frontal zones should be so sharp and well-defined as is often observed. However, these rather abrupt changes of temperature gradient may well arise from the intensification and tilting of already existing boundaries to inversions or stable layers. The horizontal divergence will intensify the changes in vertical gradients of temperature and the tilting will introduce into the horizontal temperature gradients much stronger discontinuities than are normally present."

The above arguments relate to the formation of frontal zones in intensifying jets and in the regions of upper troughs. Sawyer suggests that "the arguments may be reversed to suggest a reverse circulation and frontolysis as the air passes into weaker sections of the jet and into upper ridges. No doubt detailed observations of the jet stream and studies of upper fronts have concentrated on regions where they are intense, but I (Sawyer) think there is another reason why the structure which is built up in these intense regions of the jet and frontal zone should persist elsewhere. That is because it is unlikely that the vertical circulation as the air enters a diffluence or a ridge will be the exact reverse of that in a preceding confluence or trough. It will not be the same air which descended in the confluence that ascends again in the diffluence – sufficient small changes in the relative position of front and jet will probably have taken place to mean that the air which rises most will not be precisely that which descended most. Thus in association with weaker sections of the jet and with upper ridges, we might expect to find some remnant, albeit less intense, of the dry frontal zone built up in a preceding confluence or trough."

Among the conclusions Sawyer draws from his study are:

(i) "Theoretical considerations suggest the horizontal temperature gradient (in the frontal zone in the upper troposphere on the cold side of an intensifying jet stream) arises from the differential subsidence of air on the cold side of the jet stream which leads to a tilting of the isentropic surfaces. Dynamical considerations also explain the vertical stability and cyclonic vorticity associated with the front.

(ii) "The front-forming processes in the upper troposphere only operate where the jet stream is intensifying or becoming more cyclonically curved along its length, but the characteristic structure is carried forward into other weaker sections of the jet and probably does not degenerate completely."

7.4. THE TROPOPAUSE

The formal definitions of the tropopause used by the Meteorological Office are reproduced below. These definitions, which were introduced with effect from 1 January 1956, are consistent with, but amplify the definitions recommended by the World Meteorological Organization in Resolution No. 42 of the Fourth Session of the Executive Committee.[24]

(i) The "first tropopause" is defined as the lowest level at which the lapse rate decreases to 2°C. per kilometre or less, provided also that the average lapse rate between this level and all higher levels within two kilometres does not exceed 2°C. per kilometre.

(ii) When, above the first tropopause, the average lapse rate between any level and all higher levels within one kilometre exceeds 3°C. per kilometre then a "second tropopause" can occur and is defined by the same criteria as in (i). This tropopause can be either above or within the one-kilometre layer.

(iii) Further tropopauses are defined at higher levels by the same criteria as in (ii).

Some explanatory notes for the practical application of these definitions have also been promulgated.[25]

The levels of the earth's atmosphere with which many forecasters have to deal on a day-to-day basis extend from near the earth's surface up to a height which normally does not exceed that at which the pressure falls to about 50 millibars, that is approximately the lowest 20 kilometres. This depth of atmosphere can normally be sub-divided into two broad regions – the lower, called the troposphere, in which on average there is a decrease of temperature with increasing height and the upper, forming part of the stratosphere. This lower part of the stratosphere is generally characterized by relatively little change of temperature with height. The region separating the troposphere from the stratosphere is called the tropopause. It is normally very thin in the vertical compared with the troposphere and stratosphere and it can often be recognized as an almost continuous surface over substantial areas of dimensions of hundreds of miles. However, some breaks or discontinuities do occur. The tropopause is seldom level and sometimes slopes quite sharply – often in close proximity to the region where the tropopause becomes discontinuous. On some occasions it is possible to recognize more than one tropopause in the vertical and this gives rise to what are known as multiple tropopauses.

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°A.) 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[26] 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 two 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.

Some Features of the Free Atmosphere

Forecasters who wish to study the broad-scale distribution of tropopause pressures and temperatures should consult a paper by Moore[27] which contains maps of the world between latitude 60°S. and the north pole showing average pressures and temperatures at the tropopause for the mid-season months, January, April, July and October.

Over the British Isles the mean pressures and temperatures at the tropopause are normally between about 220 and 280 millibars and 210° and 220°A. There are wide variations beyond these limits which are usually readily recognizable on the synoptic scale from routine data. When the troposphere over the 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.*[28] 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. The examples are biased geographically to the neighbourhood of the British Isles.

Sawyer[29] 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 5°F.) 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 23.

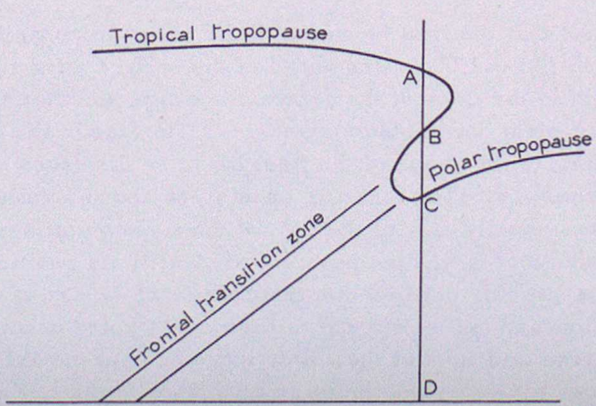


FIGURE 23 *Schematic cross-section through a warm front showing a folded tropopause*

Figure 23 shows that the level at which tropical air first replaces the polar air may be above the level of the polar tropopause. A sounding at point D would be in the polar troposphere from D to C, the polar stratosphere from C to B, the tropical troposphere from B to A and in the tropical stratosphere above A.

Sawyer found no fewer than twelve 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 is shown schematically in Figure 24. These diagrams bring out only significant features as the details varied considerably from case to case.

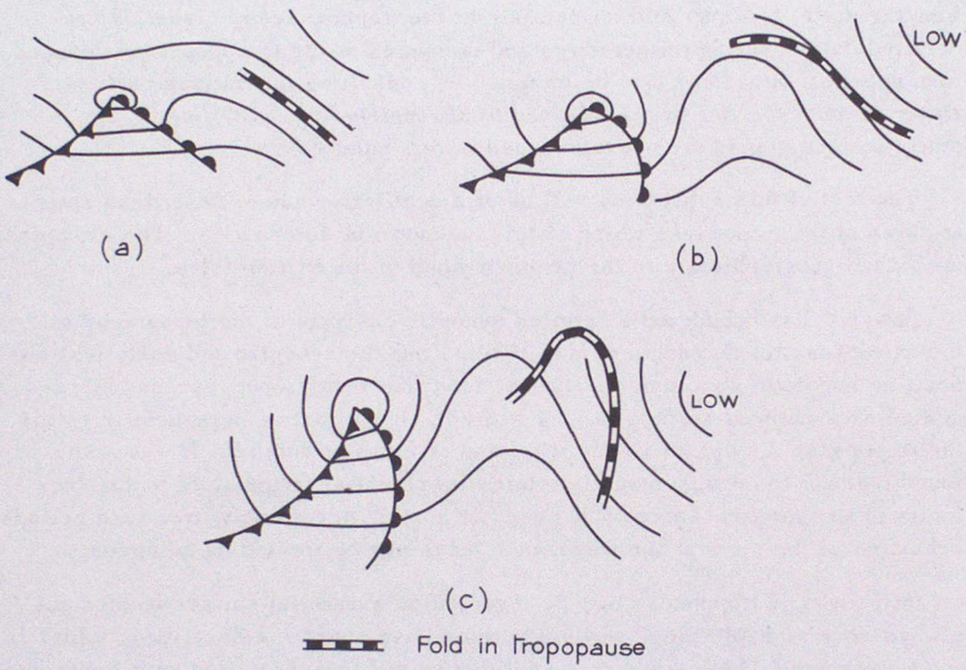


FIGURE 24 *Schematic relation of the folded tropopause to the frontal system in the horizontal*

Sawyer found that the first formation of the folded tropopause appeared to take place some 500 to 1,000 miles 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. The line of the discontinuity tended to conform to the shape of the front but to be displaced some 600 to 1,000 miles ahead of it. 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. Figure 25 shows a transverse section through a confluent entrance to a jet stream. At the confluent entrance air masses from widely different localities are brought into close proximity. The warm air would be expected to have a high tropopause and the cold air a low one and this would result in the existence of a sloping tropopause near the jet stream as shown by the dashed line in Figure 25. If the transverse wind circulation is as indicated by the arrows this sloping tropopause could be further distorted into the folded structure as shown.

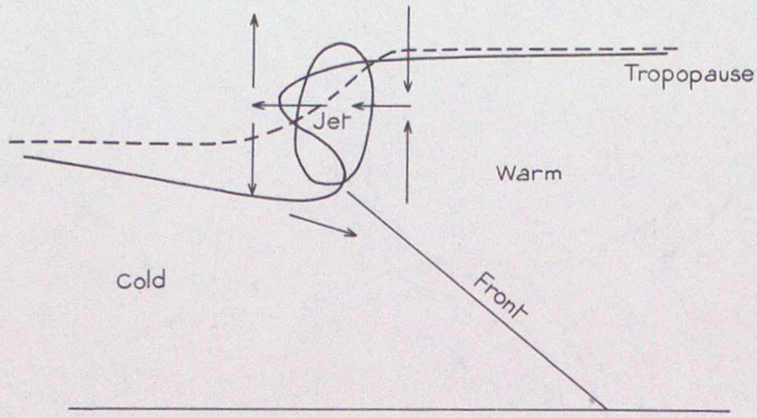


FIGURE 25 *Schematic cross-section through the confluent entrance to a jet stream, looking east (after Namias and Clapp*[18]*)*

The folded tropopause formed at the entrance to a jet stream is sometimes further distorted in the horizontal at the exit from the jet stream, as shown schematically in Figure 26, leading to a rather narrow tongue of high tropopause extending from the left exit of the jet stream. Sawyer noted eight such examples in May 1949. With one exception they were short-lived features with a life of only a little over twelve hours and were destroyed in a very similar manner to that in which warm tongues disappear during the occlusion process of depressions.

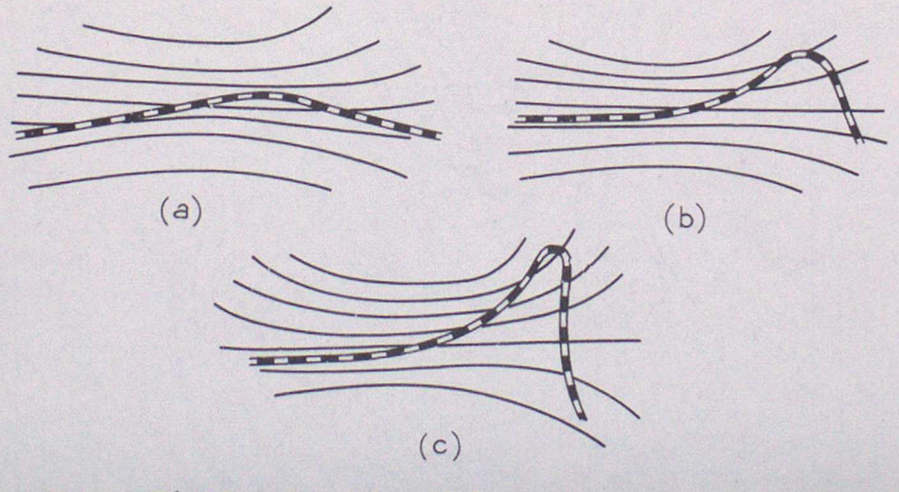


FIGURE 26 *Distortion of the folded tropopause at the exit from the jet stream illustrated by horizontal displacements at the fold*

An example of a folded tropopause is shown in Figures 27 to 30.

Discontinuities exist in the tropopause surface and are particularly well marked with very low tropopauses. To describe this phenomenon Palmén[30] has coined the expression "tropopause funnel". A "tropopause funnel" is shown schematically in Figure 31.

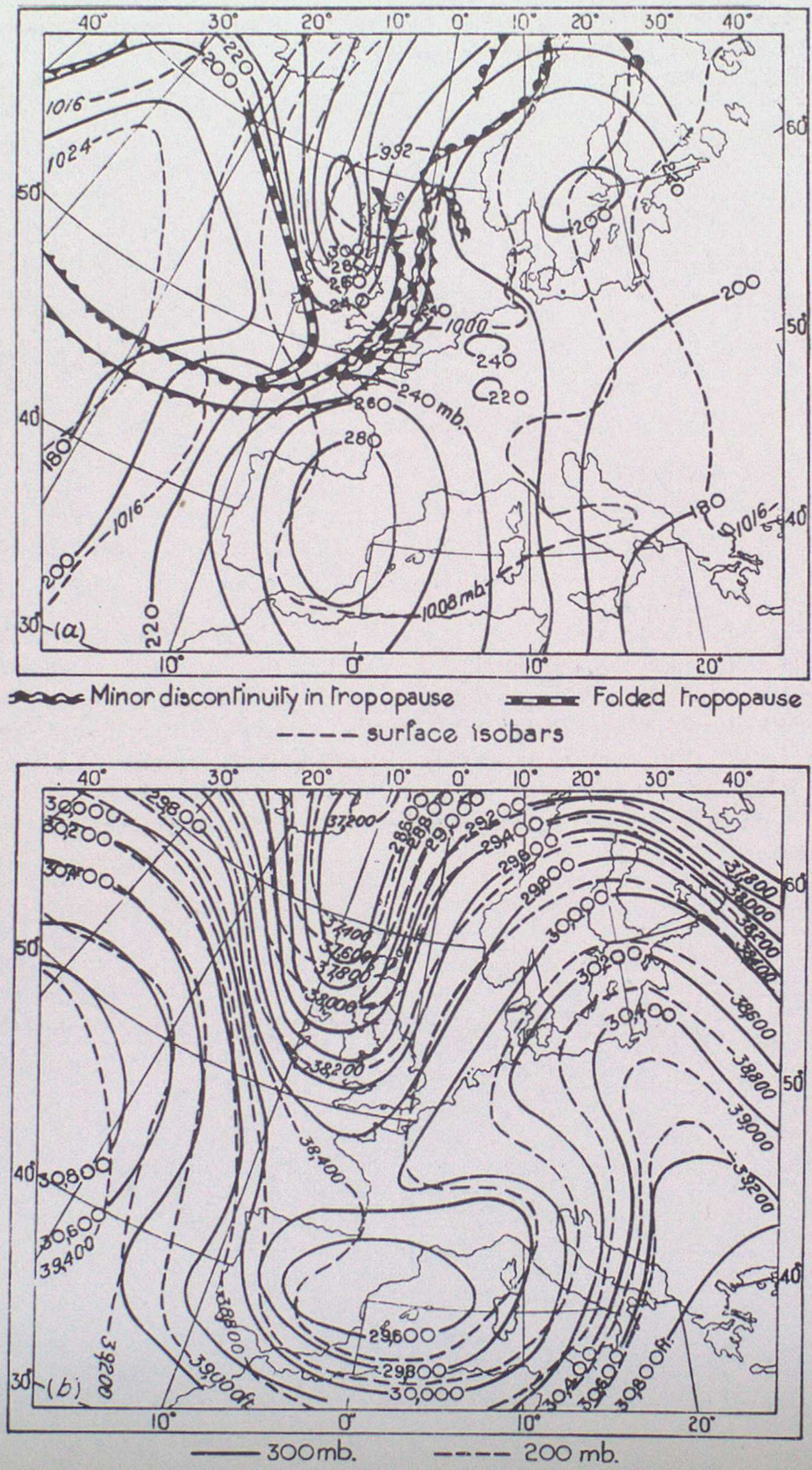


FIGURE 27 *(a) Surface chart and isobars on the tropopause and (b) contours of the 300-mb. and 200-mb. surfaces, 0300 G.M.T., 5 May 1949*

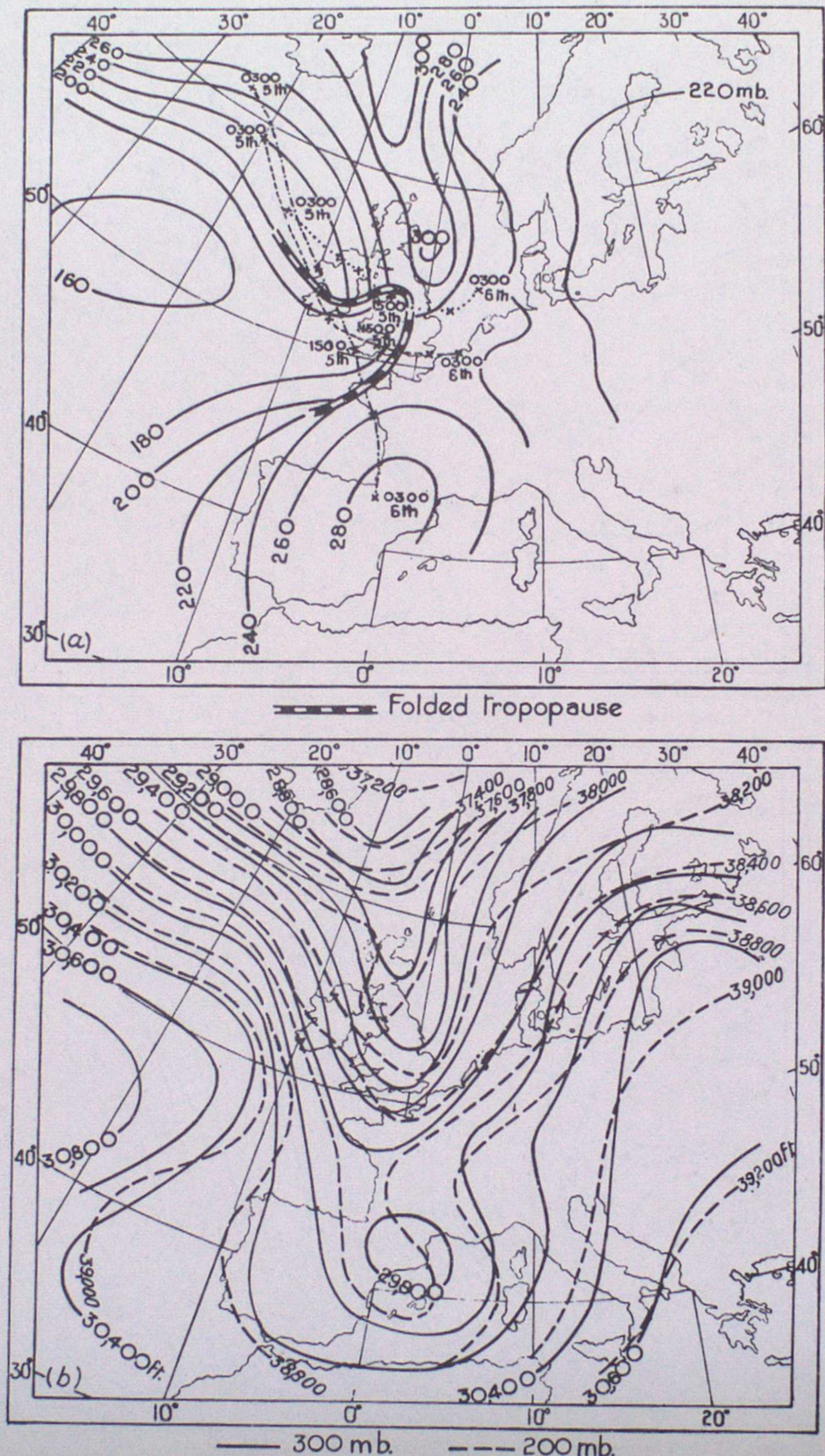


FIGURE 28 *(a) Isobars on the tropopause and estimated trajectories of air at tropopause level and (b) contours of the 300-mb. and 200-mb. surfaces, 1500 G.M.T., 5 May 1949*

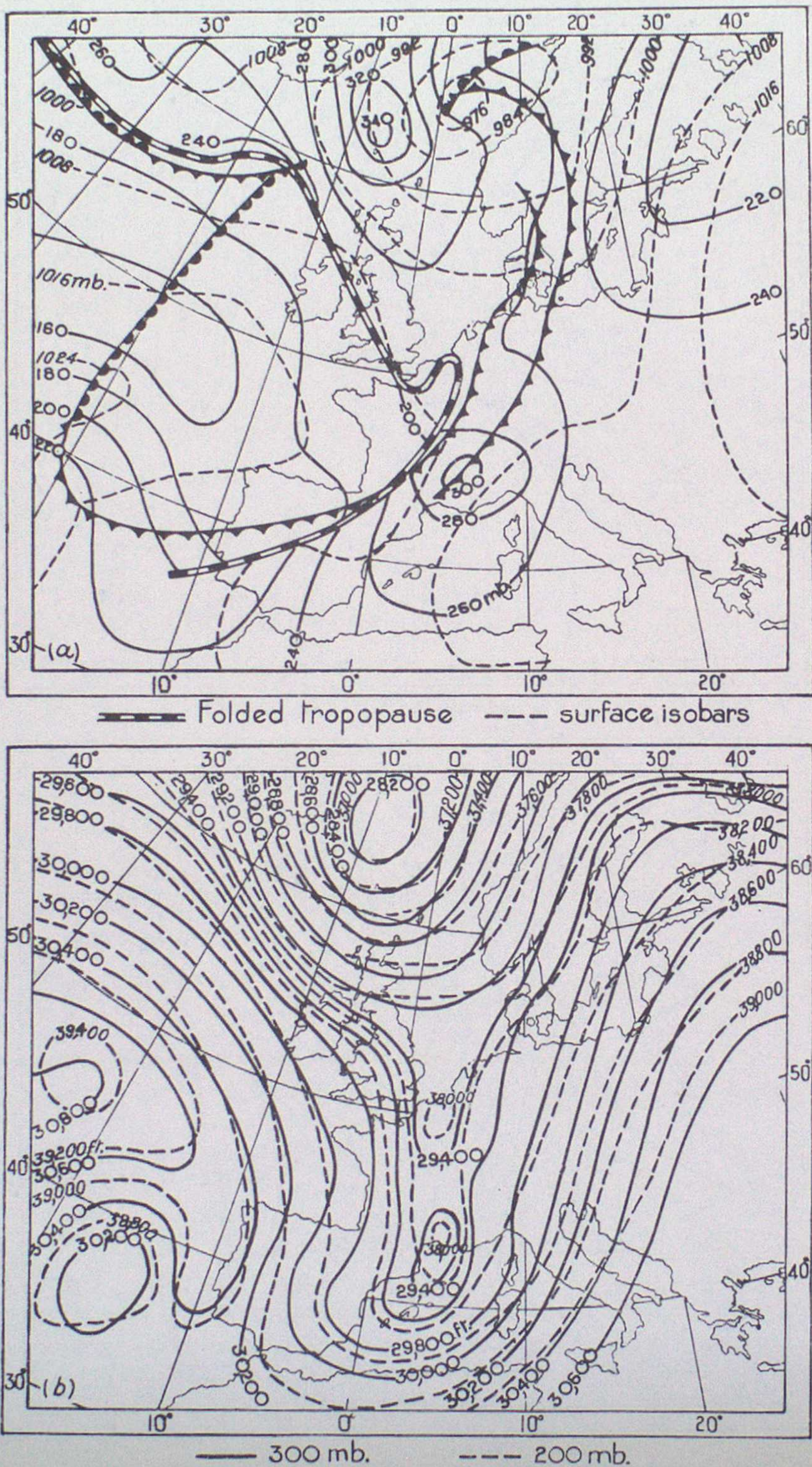


FIGURE 29 *(a) Surface chart and isobars on the tropopause and (b) contours of the 300-mb. and 200-mb. surfaces, 0300 G.M.T., 6 May 1949*

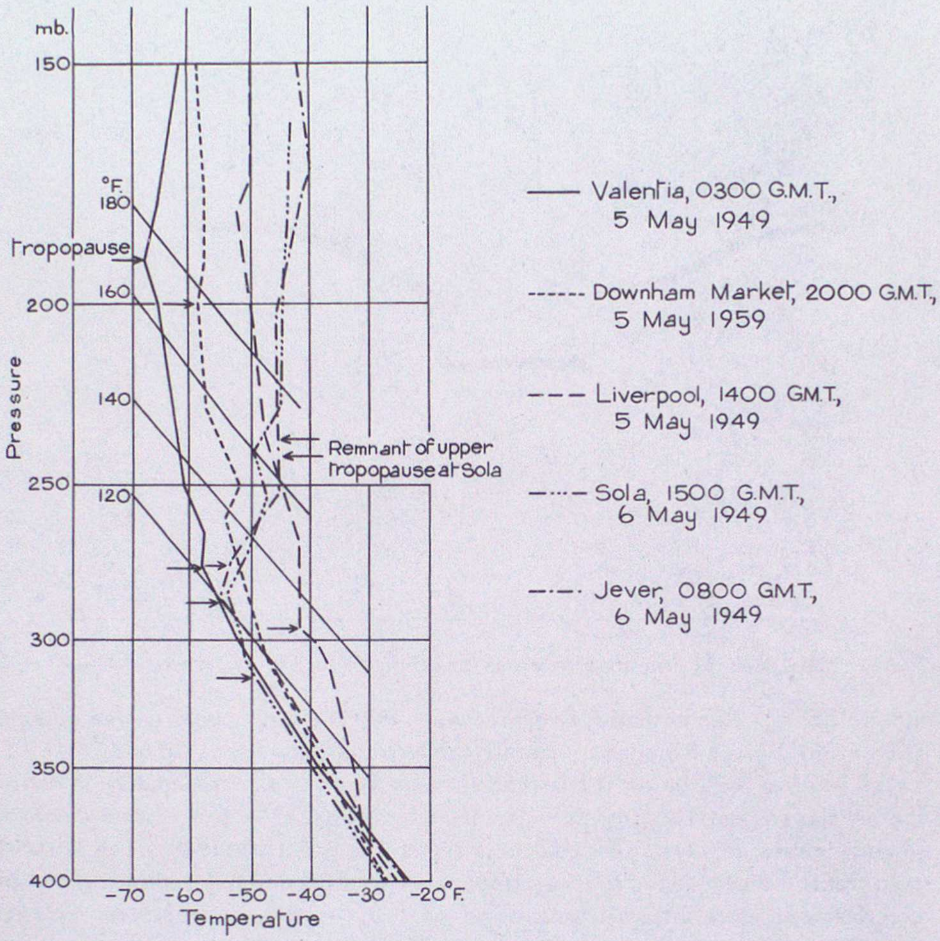


FIGURE 30 *Temperature soundings, 5–6 May 1949, illustrating the occlusion of a high-level tropopause*

Sloping lines are segments of adiabatics at potential temperatures indicated and arrows indicate the tropopause or other significant changes in lapse rate.

The centres of low tropopause shown centrally in Figure 31 develop frequently in association with vigorously deepening depressions and subsequently become centred over the old depression when it has become slow-moving and started to fill. Sawyer[29] found twelve 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 considers it probable that depressions developing close to the tip of a cold trough or to the left of a diffluence in the thermal pattern are 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 twelve 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. The tropopause is usually well below the 300-millibar level and sometimes below

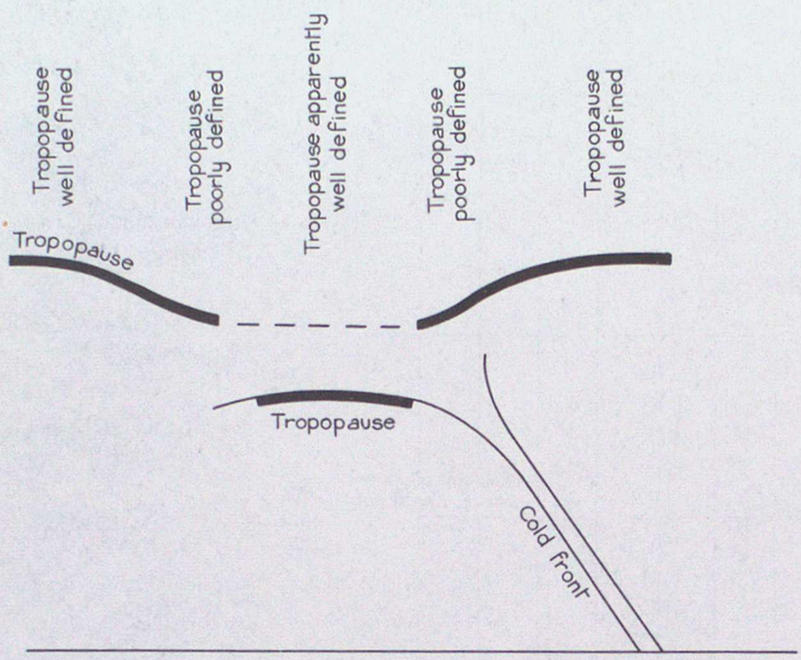


FIGURE 31 *Schematic cross-section through a tropopause funnel*

400 millibars. The potential temperature at this low tropopause in the neighbourhood of the United Kingdom is usually between 80° and 95°F. (about 27°–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 31.

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 believes the decaying process to be associated with the rapid descent of air in the tropopause funnel and its ultimate absorption into the troposphere.

The structure and behaviour of a tropopause funnel is illustrated in Figures 32–38. Charts of the pressure at the tropopause are reproduced at 24-hour intervals (Figures 32–36) together with charts of the 300-millibar contours. Surface isobars and fronts are superimposed on the tropopause charts. The 300-millibar trough in this example is not so well marked as is usual. Figure 37 shows a sequence of soundings from Larkhill which are typical for various regions in a tropopause funnel. Arrows indicate the positions of the tropopause and the discontinuity in lapse in the upper troposphere. Continuous arrows indicate the level which was accepted as the tropopause in the construction of the contour charts. Figure 38 shows two cross-sections through the funnel in perpendicular directions and illustrates its structure. The isokinetics (lines of equal normal wind speed) show a jet stream to the north and east of the tropopause funnels and a less well defined maximum to the south and west.

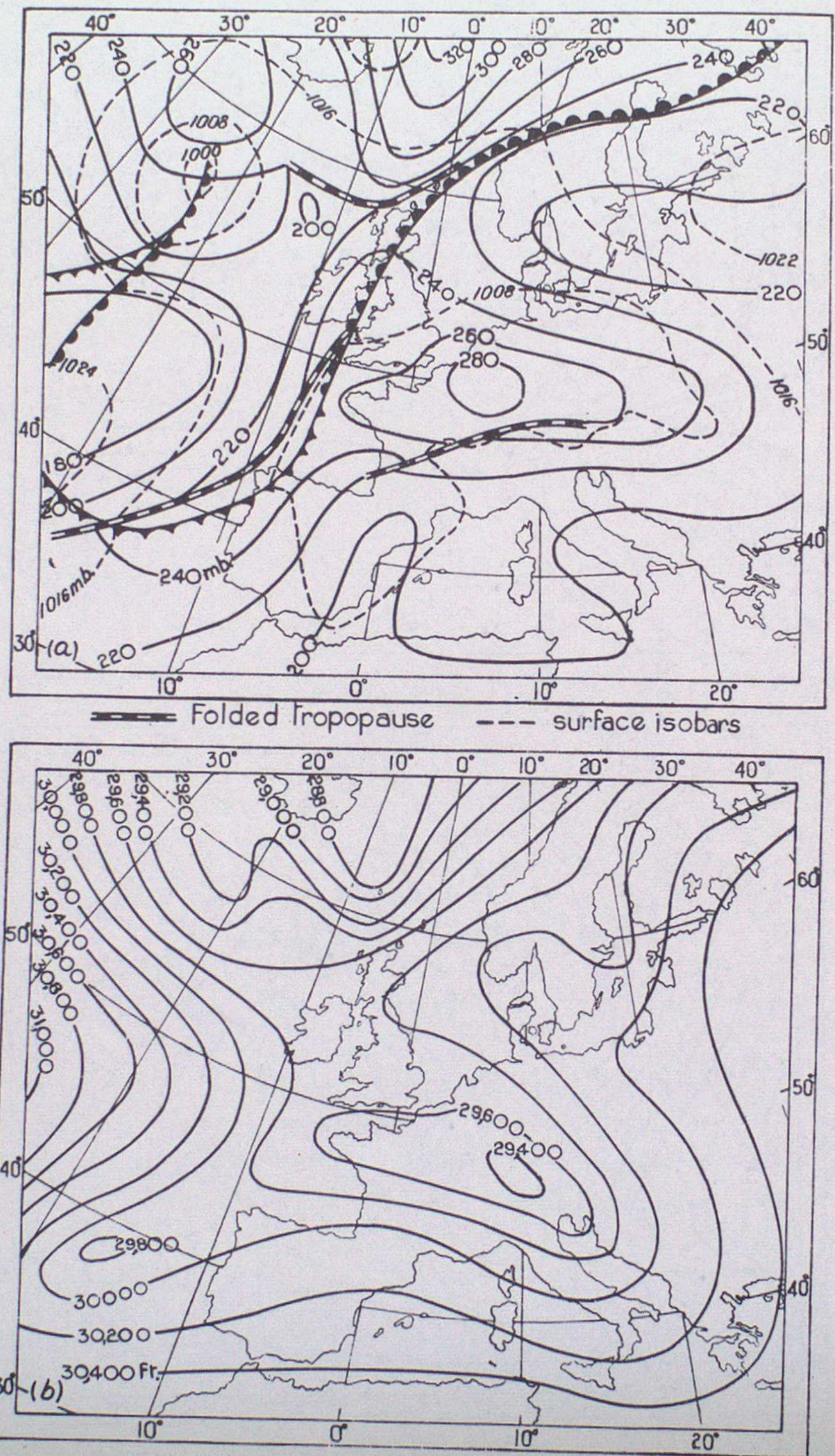


FIGURE 32 *(a) Surface chart and isobars on the tropopause and (b) contours of the 300-mb. surface, 0300 G.M.T., 15 May 1949*

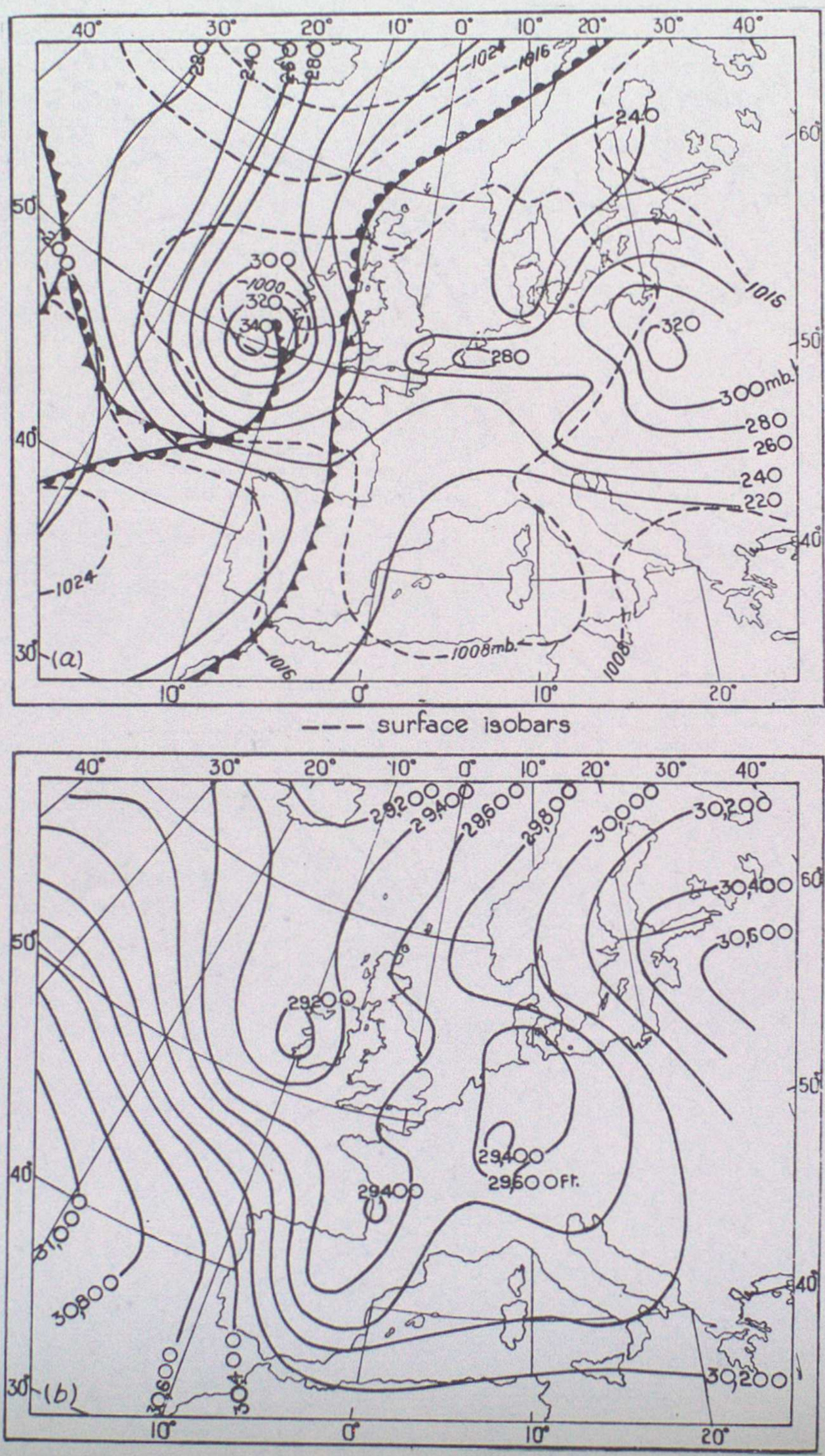


FIGURE 33 *(a) Surface chart and isobars on the tropopause and (b) contours of the 300-mb. surface, 0300 G.M.T., 16 May 1949*

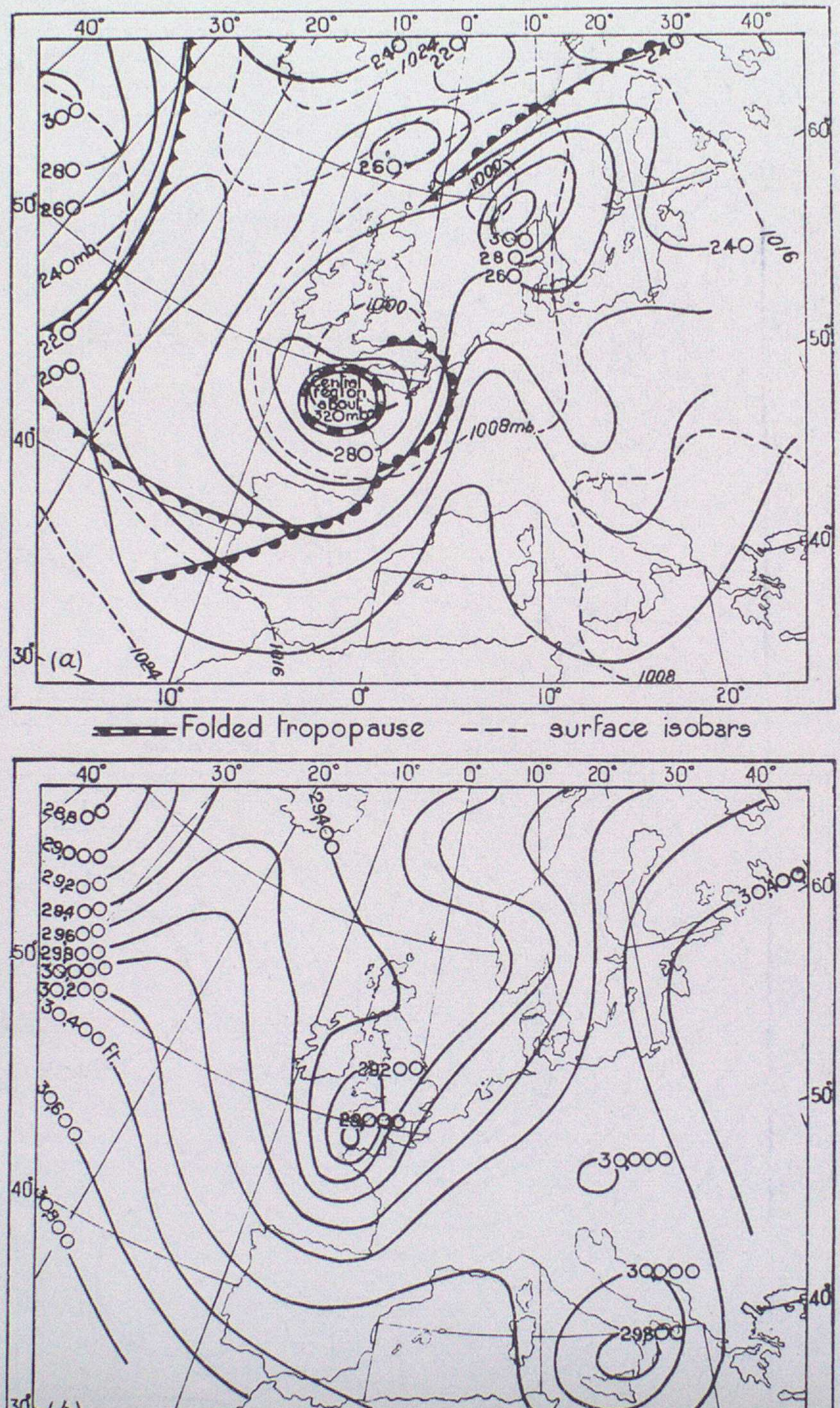


FIGURE 34 *(a) Surface chart and isobars on the tropopause and (b) contours of the 300-mb. surface, 0300 G.M.T., 17 May 1949*

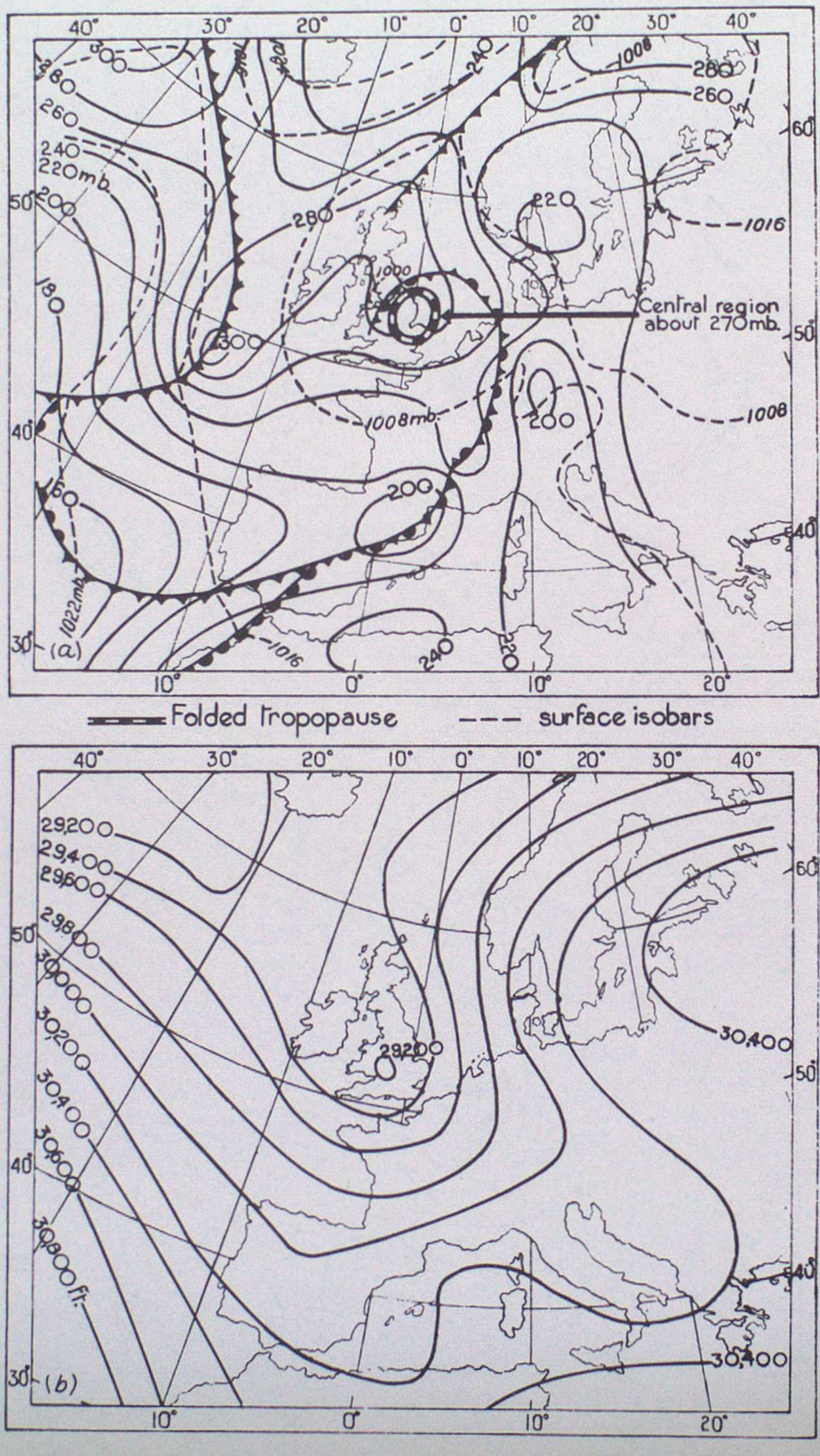


FIGURE 35 *(a) Surface chart and isobars on the tropopause and (b) contours of the 300-mb. surface, 0300 G.M.T., 18 May 1949*

Some Features of the Free Atmosphere

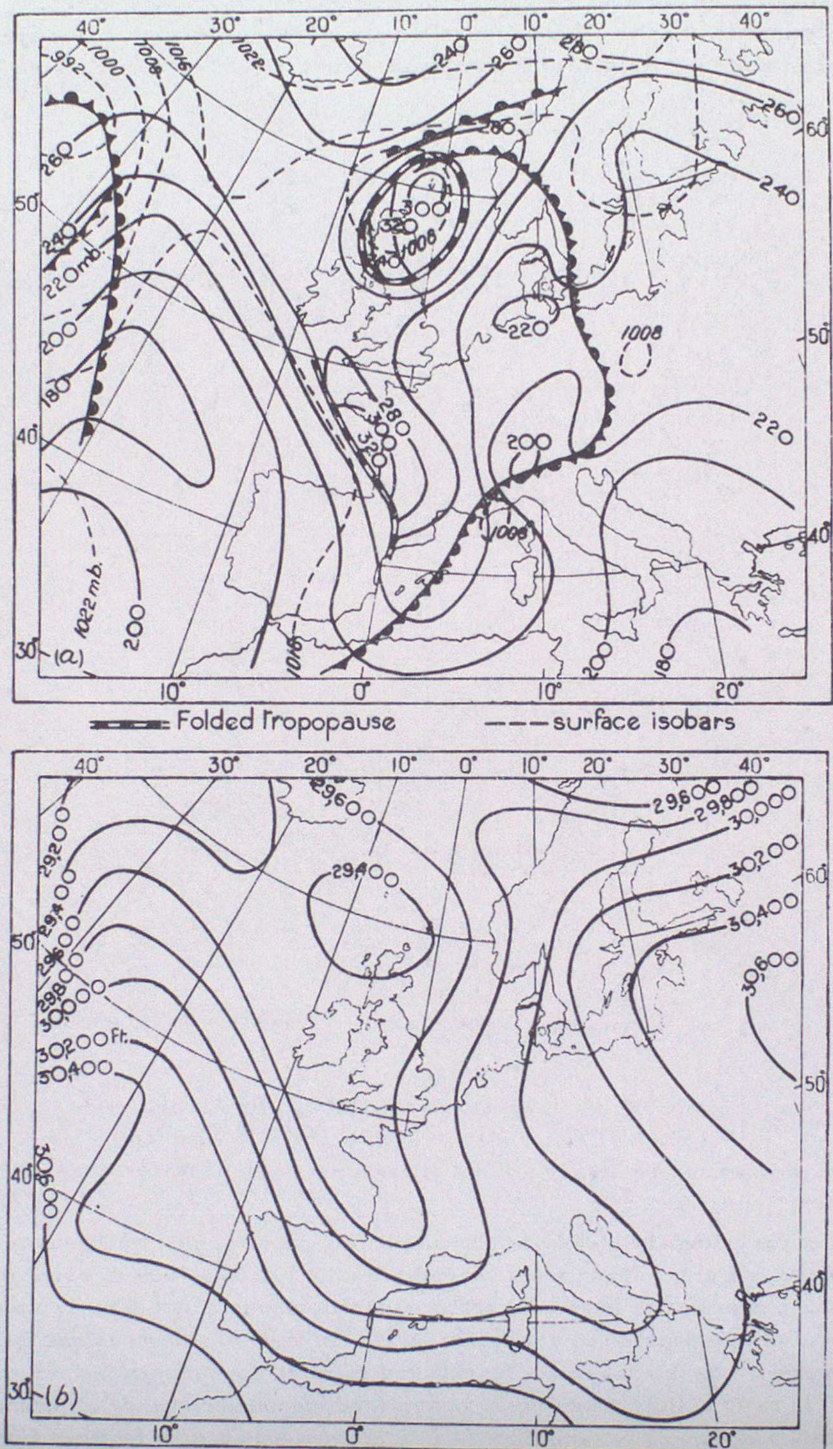


FIGURE 36 *(a) Surface chart and isobars on the tropopause and (b) contours of the 300-mb. surface, 0300 G.M.T., 19 May 1949*

Figure 36 also illustrates the overrunning of a tropopause funnel over the west coast of France where a funnel, which moved in from the Atlantic, is collapsing before the high tropopause encroaching in association with the Atlantic frontal system. There is insufficient evidence to indicate whether a typical central core existed in the tropopause funnel over western France.

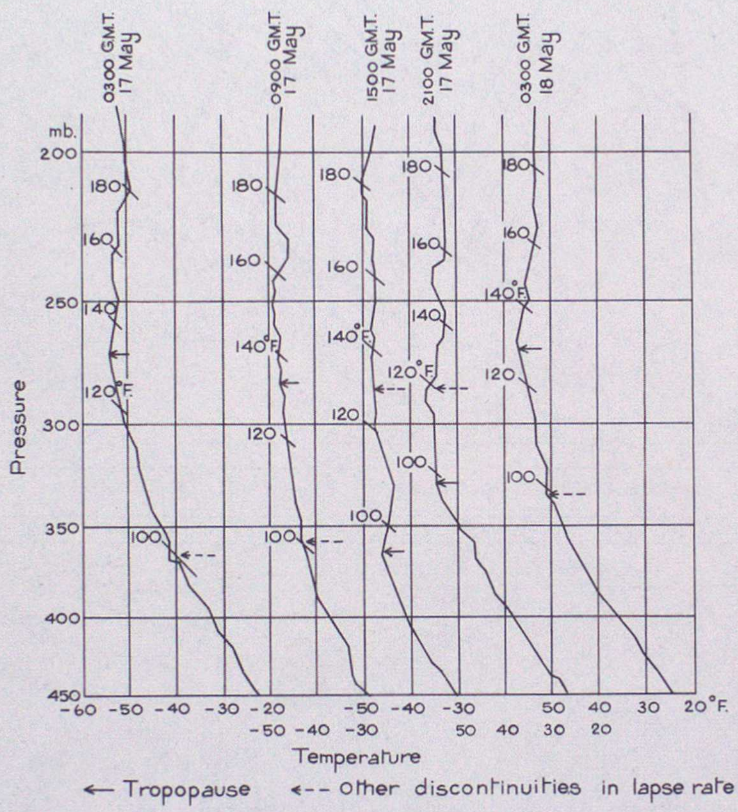
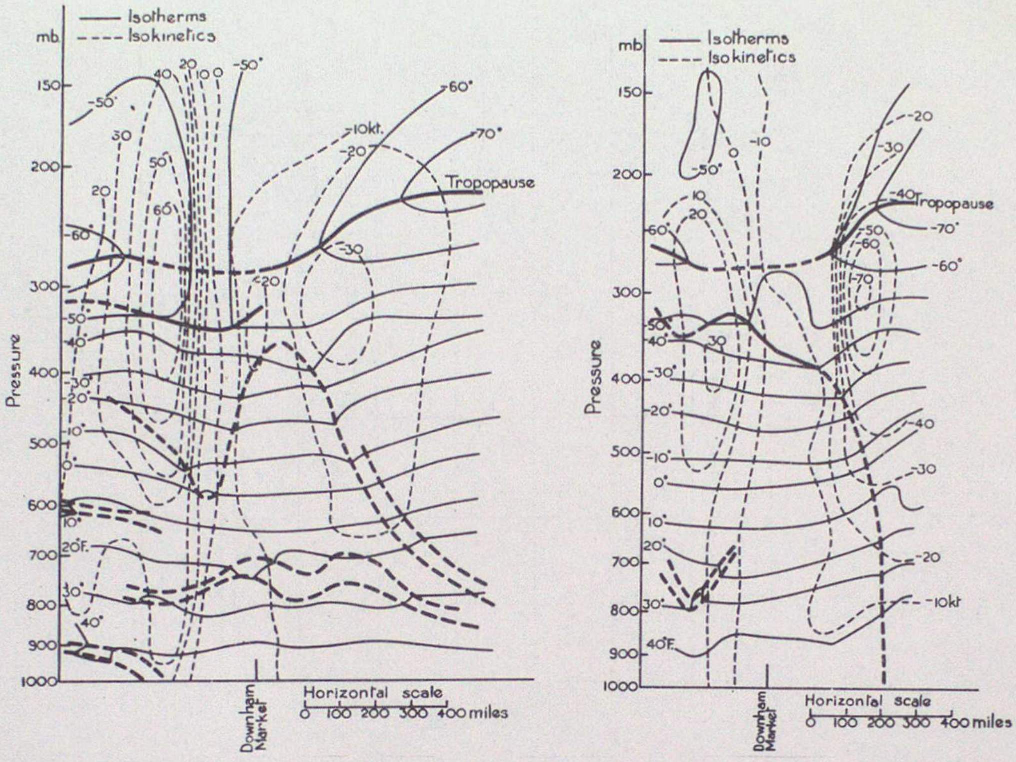


FIGURE 37 *Temperature soundings for Larkhill, 0300 G.M.T., 17 May to 0300 G.M.T., 18 May 1949*

Short sloping lines are segments of dry adiabatics at potential temperatures indicated

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 10°–20°F. (about 6°–11°C.). 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.



North–South cross-section **East–West cross-section**

FIGURE 38 *Cross-sections through the tropopause funnel*
0300 G.M.T., 18 May 1949

Heavy continuous lines indicate the tropopause and heavy broken lines indicate other significant changes in lapse rate.

An example is given in Figures 39 and 40. Figure 39 shows three successive soundings at Larkhill from 0900 to 2100 G.M.T., 21 May 1949. The original high-level tropopause at about 180 millibars at 0900 G.M.T. with a potential temperature about 160°F. disappeared while the new lower-level tropopause at about 220 millibars and a potential temperature just below 140°F., which is identifiable at 0900 G.M.T., had become dominant by 1500 G.M.T. The synoptic situation and the topography of the tropopause are illustrated in Figure 40.

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[21] mentioned that the tropopause tended to be continuous with the weaker jet streams and to be uncertain or discontinuous with the more intense ones.

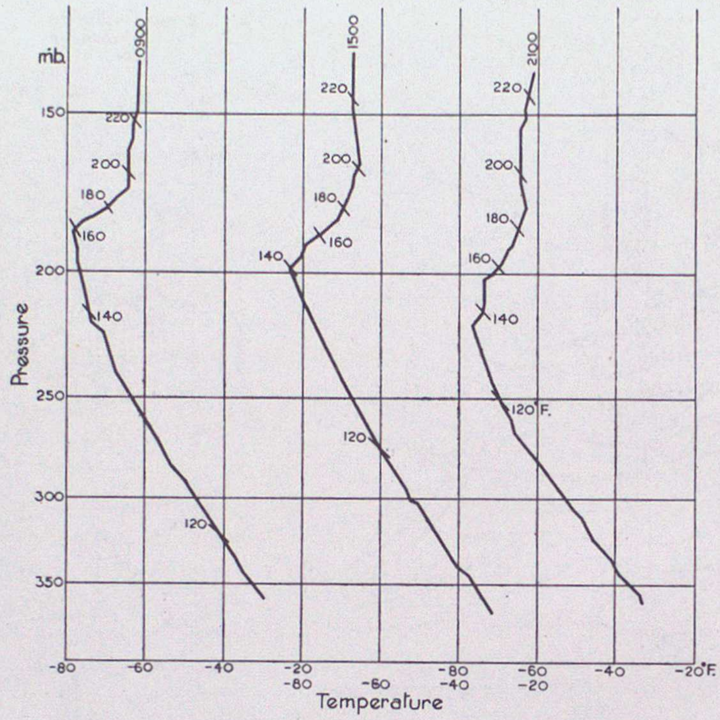


FIGURE 39 *Temperature soundings for Larkhill, 0900 to 2100 G.M.T., 21 May 1949*

Short sloping lines are segments of dry adiabatics at potential temperatures indicated

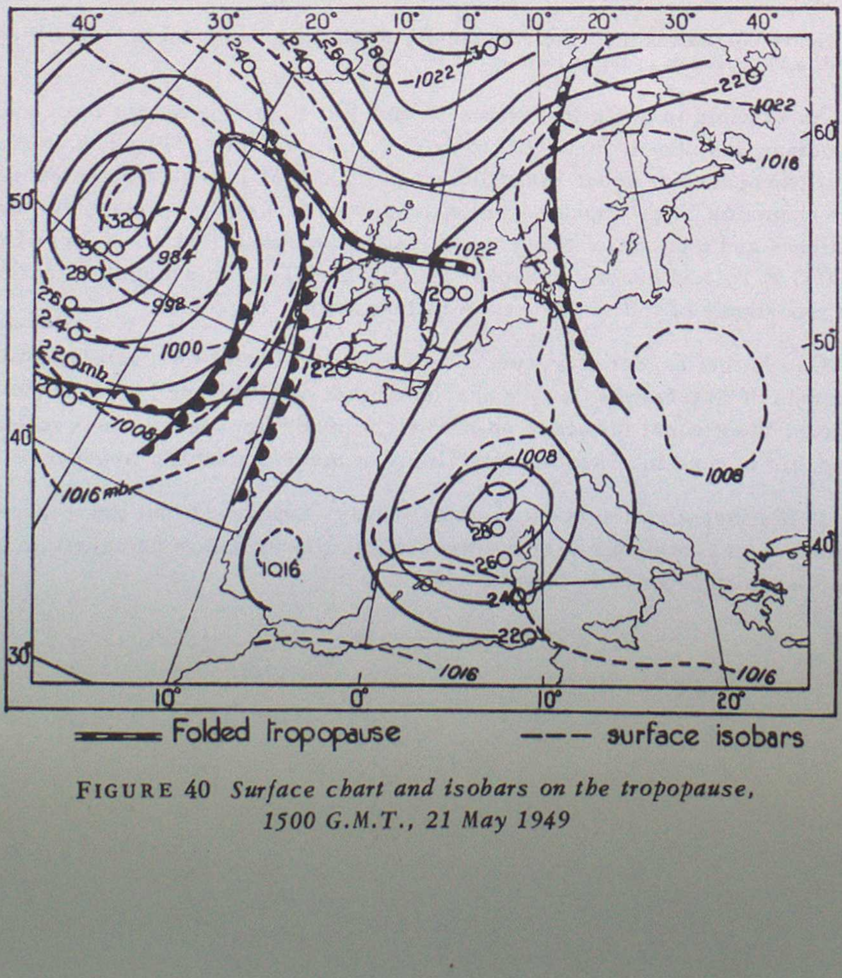


FIGURE 40 *Surface chart and isobars on the tropopause, 1500 G.M.T., 21 May 1949*

7.5. THE STRATOSPHERE

Although forecasts are not currently required as routine for altitude much in excess of about 55,000 feet (about 15 kilometres, typical pressure 100 millibars or a little less), it seemed desirable that a brief account should be given of some features of the atmosphere to rather higher levels. For levels from 15 to about 30 kilometres (pressures approximately 100 to 10 millibars) routine observational data, although less plentiful than at lower levels, are not completely absent. At levels above 30 kilometres routine data are almost completely absent. For the lower layers of the stratosphere some detail can be incorporated into the treatment. For upper layers little detail can be given; further, current knowledge is not particularly firmly established and it is possible that this account may need modification as knowledge and understanding of the upper stratospheric levels are advanced in the future, either through more data becoming available or through greater attention being paid to those levels.

It is convenient to commence this section with a description of the broad-scale features of winds and temperatures in the stratosphere from about 20 to 100 kilometres. Murgatroyd[31] has compiled an account of these features. His paper is an extensive survey and review of the available relevant literature published in the last few decades. The meteorologist who wishes to make a detailed study of winds and temperatures at these levels would find the paper an invaluable starting point. The conclusions reached by Murgatroyd should be of general interest and provide a framework in which to view the more detailed account contained in subsequent parts of this section. The main conclusions are reproduced verbatim:

"The meteorologist dealing with the atmosphere above the tropopause to 100 km or so must consider conditions in four distinct regions, i.e., the lower stratosphere, say below 30 km, the region where the temperature increases with height (30 – 60 km), the altitude band where the temperature decreases with height (60 – 80 km), and the second inversion layer above (80 km upwards). The principal features of these regions may be summarized:

"(i) The lower stratosphere has some coupling to the troposphere in that ozone amounts, temperature and winds can be correlated with tropopause conditions, but the coupling decreases with altitude (Bannon and Jones[32]). Wind systems in the region are similar to those of lower levels with sizeable meridional as well as zonal components, small-scale variations (Kochanski and Wasko[33]) and considerable longitudinal variation (Bannon and Gilchrist[34]).

"(ii) The inversion region from 30 – 60 km appears to have very little coupling with the troposphere. Friction effects and turbulence are probably small, tidal effects are inconsiderable and pressure, temperature, and wind relationships are similar to those at lower levels. Familiar atmospheric phenomena such as long waves in the general flow and possibly persistent jet streams may be expected but as yet there appears to be little evidence of smaller systems on the scale of the cyclones and anticyclones found at lower levels. Diurnal changes of temperature of a few degrees are likely depending on the solar irradiation. There is evidence also of short-period wind variations. No systematic variation of meridional components of wind has emerged from the various measurements, and the zonal components are usually stronger and more persistent in character.

"(iii) From 60 to 80 km the lapse rate appears to be large enough for convection to take place, particularly over high latitudes in summer, although it is doubtful whether the dry adiabatic lapse rate is exceeded. The evidence that the mesopause* is at its coldest above the hottest part of the ozonosphere parallels conditions at the tropopause which is also at its coldest (and highest) above the hottest region below. Pant[35] has recently attempted to explain this qualitatively in terms of the infra-red radiation balance following the lines of the discussion of Dobson, Brewer and Cwilong[36] of the radiation temperature of the tropopause. The winds in this region have regular seasonal features with the zonal components larger and more steady than the meridional component. Tidal components start to be easily measurable at the top of this region, their importance increasing with altitude. Little can be inferred about the size of the pressure systems here but considerable turbulence and large wind shears are a notable feature of all reported observations.

"(iv) Turbulence is also found in regions from 80 km upwards although the lapse rate is very stable there. The evidence for this arises mainly from observations of meteors and it is not altogether clear to what extent these merely reflect the meteor's disturbance of the atmosphere or whether they are representative of atmospheric conditions. This could be tested by observing whether the turbulence measured decreases with time in given portions of long-enduring trails. The problem of wind measurement in this altitude band is complicated by the large oscillatory components but the radar-meteor technique of wind finding has made it possible to extract the prevailing winds. Zonal components have again been shown to be predominant although their seasonal changes of direction require explanation. The variation of the oscillatory components with altitude and season, although complex, is in reasonable agreement with the predictions of the resonance theory."

In another paper Murgatroyd[37] has considered wind and temperature to 50 kilometres over England. This paper is geographically biased to the area with which many forecasters in the Meteorological Office have to deal and some of the levels considered by Murgatroyd in this paper are not vastly higher than those at which some jet aircraft currently operate. The discussion of the results obtained by Murgatroyd is reproduced verbatim:

"The main conclusions are as follows:–

Temperature.–

"(i) There is a large inversion of temperature in the 40–50-km. levels in these latitudes. At 50 km. values of temperature roughly equal to surface values are suggested.

"(ii) From about 20 to 30 km. the temperature is roughly isothermal, but there is some evidence, particularly in winter, that there may be a slight lapse of temperature there.

"(iii) At any given level at 40 km. or above there is a large annual variation of temperature. Approximately 40°A. is suggested for the amount of this variation at a height of 50 km.

Wind.–

"(i) The wind at 30–50 km. in these latitudes is predominantly westerly in winter and easterly in summer. The winter westerly components may attain values of 50–60 m./sec. (97–116 knots) but the summer easterly components have maximum values of 20–30 m./sec. (39–58 knots). There is some evidence that at levels greater than 50 km. the wind may be westerly during both winter and summer although the speeds are likely to be considerably greater in winter.

---

*The shallow region in the high atmosphere near 80 kilometres where the lapse of temperature (from about 60 to 80 kilometres) changes to an inversion is called the mesopause (see W.M.O. – No. 121.RC 21, Res. 15, p. 73, for Terminology in the High Atmosphere).

"(ii) North–south components of wind are usually less than the east–west components at these high levels. Northerly components appear to be more frequent in winter and southerly components more frequent in summer, but the amount of data is too small to draw any more definite conclusions on the meridional circulation.

"*General.*– Most of the above conclusions have been suggested in other work. Previous calculations based on anomalous sound receptions, meteor observations, calculation of direct absorption of solar radiation by ozone, work on atmospheric oscillations, and measurements in American rocket trials all lead to the conclusion that there must be a region of high temperature at about 50 km. above the earth's surface, as found in this work.

"As far as is known all other evidence so far available suggests that the stratosphere remains roughly isothermal from the tropopause until an inversion starts at about 30 km. Recent radio-sonde measurements in Great Britain[38] show that in summer the inversion starts below 30 km. but that in winter temperature falls continuously, though slightly, from near the tropopause to the limit of the ascents at about 30 km.

"The large annual variation in temperature at 40 km. or above is in agreement with results given by Wexler[39] and Crary.[40] It is reasonable to suppose that if these high temperatures are caused by direct absorption of solar radiation the values attained would be less in winter than in summer. This would however be complicated by the annual variation of ozone. If there is more ozone at higher levels in winter all the absorption could take place at higher levels in winter than in summer. In this case the maximum of temperature might occur at greater heights in winter than in summer. This result was also obtained by Crary.[40]

"The annual change of easterly to westerly wind components in these latitudes at 30–50 km. is now well known. It was discussed by Whipple[41] in connexion with his pioneer work on anomalous sound reception, and has since been supported by the work described by Murgatroyd and Clews[42], Scrase[38], Brasefield[43], Crary[40], and Richardson and Kennedy[44]. The reason for this annual change must be the differential heating between the poles and the equator at high levels, with the poles colder during the polar night but becoming warmer than the equator in the summer on account of increased absorption of solar radiation in high latitudes. The tendency for a return to westerly winds at all seasons about 50 km. would indicate that 50–60 km. may be the level of the effective top of the ozone layer."

The two preceding extracts from Murgatroyd's work will have given the reader a general idea of conditions in the stratosphere to heights of about 100 kilometres. This level is well above those currently reached in routine flights in manned vehicles. The remainder of this section will be restricted to a rather more detailed discussion of some aspects of the lower stratosphere below about 100,000 feet (30 kilometres or 10 millibars approximately).

A paper by Bannon and Jones[32] contains information on the mean wind at 60 millibars (approximately 65,000 feet or 20 kilometres) over the world between latitudes 60°S. and 75°N. The mean winds for the mid-season months of January, April, July and October are depicted on charts by streamlines and isotachs. The charts show that in January there is a general westerly circulation north of about 25°N. Two areas of maximum wind occur; one is located over the Canadian Arctic and southern Greenland with mean speeds of over 70 knots and the other is situated to the north-east of Japan with mean speeds of over 90 knots. For the British Isles and the Mediterranean area the mean winds are between westerly and west-north-westerly with speeds in the order of 30 knots, although individual wind speeds of 70 knots have been reported at most stations in these areas. In April the westerly flow is much weaker everywhere in the northern hemisphere where the southernmost limit of mean westerly flow is somewhat north of 25°N. Over the British Isles mean winds in April are between west and west-north-west with speeds in the order of 10–20 knots. In July mean winds between 40°N. and 75°N. are generally light and vary in direction. Over the British Isles the vector mean winds in July are almost calm but the individual maximum wind tends to be from an easterly direction with speeds in the order of 15–25 knots. The October mean charts show that a general westerly flow is re-established around the northern hemisphere between about 30°N. and 75°N.; mean speeds are somewhat greater generally than in April but still in the order of 10–20 knots over the British Isles. Although the mean currents are of moderate strength, individual winds of considerably greater speed, in the order of 50 knots, have been recorded in October over the British Isles.

Scrase[38] has examined a series of high-level radio-sonde ascents which were made at night from Downham Market and Lerwick during 1949-50. The ascents reached levels of about 100,000 feet (30 kilometres or 10 millibars approximately). The mean values of temperatures and wind speeds for the winter half-year are shown in Figure 41. The temperature curves show a slight but continuous decrease of temperature with height above the tropopause and the lapse rate is about 0·2°F. per 1,000 feet (approximately 0·4°C. per kilometre). It will be noted that compared with Downham Market the temperatures over Lerwick are slightly higher in the lower stratosphere and lower by about the same amount (4°F. or 2·2°C. approximately) at about 100,000 feet. At this height, temperatures were about –80°F. (–62°C.) and individual observations varied up to about 20°F. (11°C.) on either side.

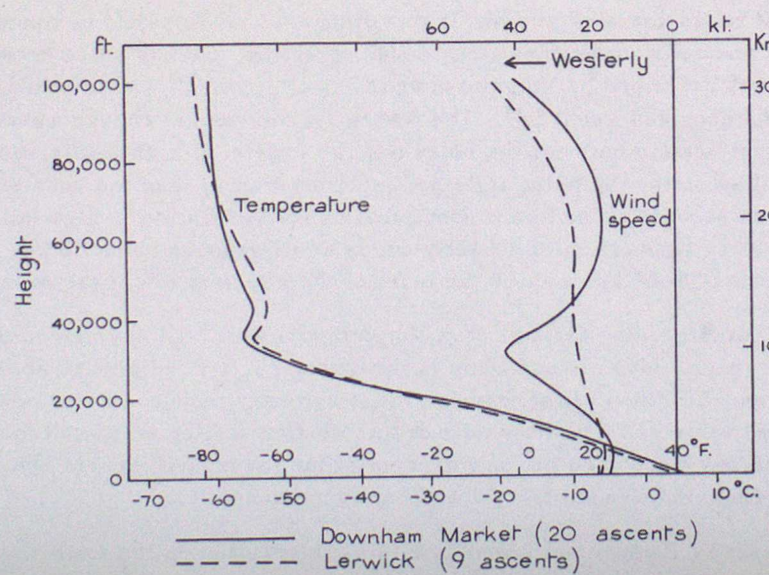


FIGURE 41 *Means of high-altitude soundings at 2200 G.M.T. for the winter half-year, October to March*

Results for the summer half-year are shown in Figure 42. The isothermal region of the stratosphere extends from just above the tropopause to about 80,000 feet (about 24 kilometres). Above 80,000 feet the temperature rose steadily at about 1°F. per 1,000 feet (1·8°C. per kilometre) over Downham Market and about 0·6°F. per 1,000 feet (1°C. per kilometre) over Lerwick. In the lower stratosphere, temperatures over Lerwick were greater than those over Downham Market by about 6°F. (3·3°C.) and lower by about the same amount at 100,000 feet. At this height temperatures averaged about −46°F. (−43°C.) and individual observations varied up to about 12°F. (about 7°C.) on either side.

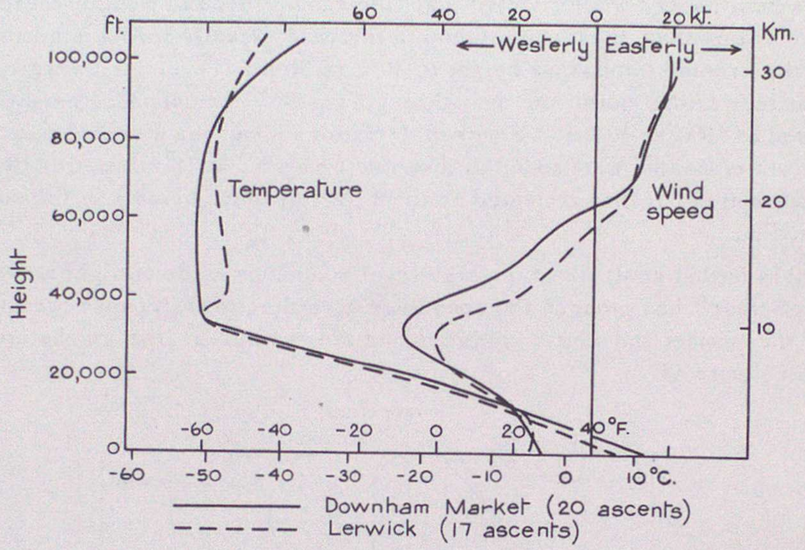


FIGURE 42 *Means of high-altitude soundings at 2200 G.M.T. for the summer half-year, April to September*

Although the number of observations is limited, they are probably sufficient to indicate the trend of annual variation of temperatures at high levels. Scrase obtained the graphs reproduced in Figure 43. The figures against the curves show the number of observations used to compute the graphs. These are quite small but the curve for 60,000 feet is reasonably consistent with that obtained from routine soundings over relatively long periods. It will be noted that the

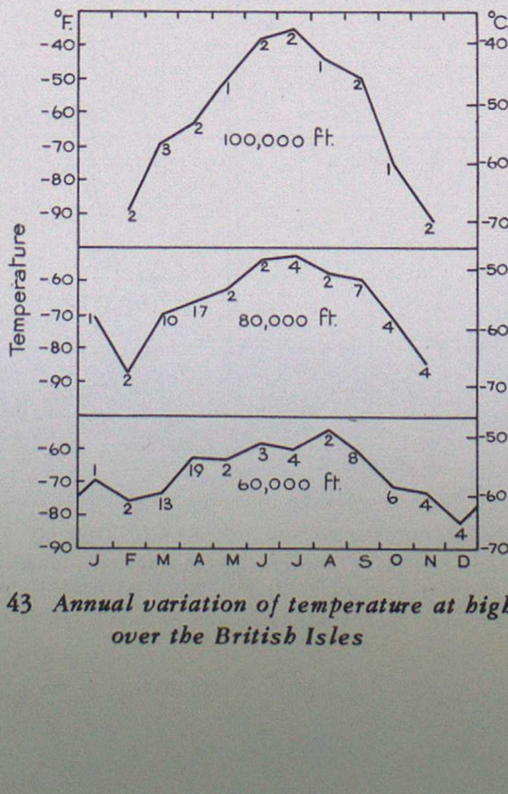


FIGURE 43 *Annual variation of temperature at high levels over the British Isles*

time of maximum temperature is earlier in the year at 100,000 feet than it is at 60,000 feet. Dobson, Brewer and Cwilong[36] have explained this type of variation as being due to the presence of greater quantities of ozone at these levels in spring. As ozone absorbs solar radiation strongly for wavelengths below about $0{\cdot}3\mu$ this will cause maximum temperatures at these levels to occur rather earlier in the year.

In a later paper Scrase[45] stated that "the annual range of monthly mean temperature at Downham Market was found to increase steadily from a minimum of 6°C at the average tropopause height to 30°C at 30 km ...... At Lerwick, about 8° of latitude further north, the annual range was 8°C at tropopause height and increased to 38°C at 30 km." Standard deviations about the mean temperatures for the winter months were found to increase from 5°C. at 18 kilometres to 10°C. at 30 kilometres but they remained at about 3°C. at these heights in the summer months.

In this further analysis of the high-level soundings made at night at Downham Market, Scrase[45] has grouped the soundings according to tropopause pressures and around the summer and winter solstices and the equinoxes. His graphs are reproduced in Figure 44.

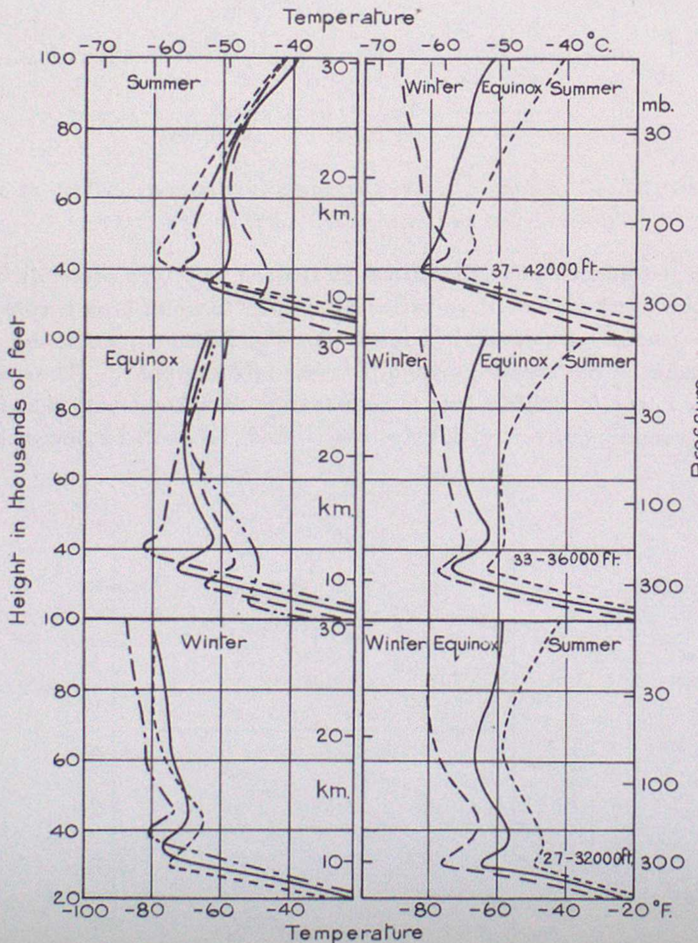


FIGURE 44 *Mean temperature–height curves at Downham Market grouped according to seasons and tropopause heights*

Summer = June and July; Winter = December and January;
Equinox = mid-March to mid-April and mid-September to mid-October

Each of the curves represents the mean of from three to six soundings

Some Features of the Free Atmosphere

Scrase remarked that: "These graphs clearly confirm the close correlation between tropopause height and temperatures in the lower stratosphere up to about 20 km, these temperatures becoming steadily warmer with decreasing tropopause height, but they indicate little or no correlation above 20 km. Correlation between tropopause height and lapse rate also becomes small about that height, whereas in the lower stratosphere decreasing tropopause height is associated with increasing lapse rate.

"Above 20 km it is the seasonal variations in temperature and in lapse rate that predominate over the effects of tropopause variations. Thus at 25 km in winter months the lapse rates average about 0·2°C/km, irrespective of tropopause height, whereas in the equinoctial and summer months there are inversions averaging about 0·3°C and 1·2°C/km respectively."

The improved performance at high levels of the apparatus currently used as routine in radio-sonde ascents has resulted, during the late 1950's, in increased amounts of routine data to levels of about 25 millibars or somewhat above. Data are still relatively sparse at these upper levels but several attempts, notably in North America, have been made to analyse the observations on what approximates to a synoptic basis. Kochanski[46] has discussed the distribution of wind and temperature in the stratosphere by considering a vertical meridional section of the atmosphere along 80°W. from the equator to the pole. Zonal wind components (that is, westerly or easterly) up to heights of about 25 millibars (26 kilometres or so) were shown by isotachs on the cross-section. The profiles of mean temperature extended to the 10-millibar level. Kochanski and Wasko[33] have prepared several daily series of charts for the 100-, 50- and 25-millibar levels over North America and have used them to discuss the wind flow and thermal structure of the atmosphere in those regions. Lee and Godson[47] have analysed high-level data from the Canadian Arctic for the winter 1955-56. Running means of temperature at 100 millibars averaged over ten days for several stations were computed and deductions made about the change of temperatures during the Arctic winter. Their results showed that a cold polar stratosphere formed over Canada in high latitudes (about 80°N.) from about November onwards. The strong temperature contrast between this cold stratosphere and the higher temperatures at comparable levels somewhat farther south in Canada set up a strongly baroclinic zone with an associated strong wind belt (approximately north-westerly in direction over northern Canada). As winter progressed this strong wind belt increased in intensity and moved slowly southwards and by January reached latitudes about 65°–70°N. over Canada. During February and March the stratospheric jet stream moved northwards and reached its maximum intensity near the end of February. During March it decreased rapidly in intensity. Selected vertical cross-sections from 400 millibars to pressures below 25 millibars showed the wind and temperature structure. The cross-section for 26 February 1957 showed a well marked jet stream with speeds of 160 knots in the core which was located at 25 millibars. The stratospheric jet stream did not appear to be a steady current moving steadily southwards and northwards with the seasons. It seemed to be subject to baroclinic disturbances which caused the jet to oscillate over a few days from its mean position – a feature which was consistent with the variance of running mean temperatures at several stations. It is clear that if a strong jet stream moves laterally over a station, upper air temperatures above that station at levels near the jet stream will vary considerably depending on whether the radiosonde ascends on the cold or the warm side of the jet.

In addition to these distortions of the polar-night stratospheric jet there is evidence to show that sometime between about January and April there is a sudden and large-scale upheaval of the wind and temperature fields and a sudden warming of the stratosphere occurs over substantial areas. Scherhag[48] was the first to report a sudden warming. Other accounts of similar occurrences have been given by Lee and Godson,[47] Craig and Hering,[49] Teweles and Finger,[50] and Teweles.[51] Scrase[52] has reported a sudden warming over the United Kingdom which is probably the same phenomenon. For more than a week in February 1951, temperatures from 18 to 30 kilometres over the United Kingdom were well above the mean for periods immediately before and after. At the peak of the warm period, temperatures at 18, 24 and 30 kilometres were about 11°C., 14°C. and 22°C. higher than those outside the period. The published accounts of these sudden warmings indicate that they are intimately connected with the break-up of the stratospheric winter jet stream. The break-up seems to occur on dates, in regions and in a manner which differ somewhat from year to year but the break-up is generally accompanied by spectacular changes in the wind and temperature fields. Wexler[53] has remarked that the stratospheric warming over the Arctic seems to occur before the return of sunlight at those levels whereas the stratospheric warming over the Antarctic in the spring of 1957 and 1958 occurred after the return of sunlight. These stratospheric phenomena are not yet fully understood and a complete account must await further research.

Flohn, Holzapfel and Oeckel[54] have shown that in summer stratospheric easterlies in middle latitudes lasted from May to August and their lower boundary was at 18 to 20 kilometres, decreasing towards the equator and the pole (the upper boundary was near 80 kilometres). Above Europe (and eastern Asia) Flohn *et alii* found that practically no relation existed between the easterlies above 20 to 22 kilometres and the varying winds near the tropopause. The occurrence of "mesometeorological" oscillations in the easterlies was demonstrated by the examination of tracks of constant-level balloons.

Knowledge of meteorological events in the stratosphere is not yet very firmly established and the above account may need modification as further progress is made. Nevertheless the account should be a useful introductory text and contribute to a better understanding by forecasters of the day-to-day observations in the lower stratosphere obtained from routine radio-sonde ascents. From stations in north-west Europe there is usually a sufficient number of ascents which reach heights of the order of 15 to 20 kilometres (about 100 to 50 millibars) so that the horizontal and vertical distributions of temperature and wind can be determined with fair accuracy. Examination of these data from the tropopause to about 50 millibars shows that day-to-day variations of temperature exist and that it is possible to distinguish areas of relatively warm and cold stratospheric air. These areas of warm and cold air behave differently from those in the troposphere where synoptic experience soon shows that these tropospheric areas of cold and warm air tend to move with the wind. In the stratosphere the areas of warm and cold air seem quite slow-moving. Sawyer[55] found that these areas tended to retain a constant position in relation to the troughs and the ridges of the flow as the wind blew through them and considered that the vertical motion of the air at these levels was probably dominent in controlling the temperatures.

Some Features of the Free Atmosphere

BIBLIOGRAPHY

1. SAWYER, J.S.; The free atmosphere in the vicinity of fronts. *Geophys. Mem., London*, 12, No. 96, 1955.

2. FREEMAN, M.H.; Fronts investigated by the Meteorological Research Flight. *Tech. Note*, No. 4, unpublished.

3. MATTHEWMAN, A.G.; A study of warm fronts. *Prof. Notes met. Off., London*, 7, No. 114, 1955.

4. SAWYER, J.S. and DINSDALE, F. E.; Cloud in relation to active warm fronts near Bircham Newton, 1942–46. *Prof. Notes met. Off., London*, 7, No. 115, 1955.

5. SANSOM, H.W.; A study of cold fronts over the British Isles. *Quart. J.R. met. Soc., London*, 77, 1951, p. 96.

6. Geneva, World Meteorological Organization; Ninth session of the executive committee, Geneva, 24th September – 14th October, 1957. Abridged report with resolutions. W.M.O. Publication No. 67, RC.14, Geneva, 1958, p. 64.

7. BERGGREN, R., GIBBS, W. J. and NEWTON, C.W.; Observational characteristics of the jet stream. (A survey of the literature). Report of a working group of the Commission for Aerology. W.M.O. Technical Note No. 19, Geneva, 1958.

8. JOHNSON, D.H.; Annotated bibliography of the jet stream. Unpublished.

9. RIEHL, H., ALAKA, M. A., JORDAN, C. L. and RENARD, R.J.; The jet stream. *Met. Monogr., Boston, Mass.*, 2, No. 7, 1954.

10. GILCHRIST, A.; Winds between 300 and 100 mb. in the tropics and subtropics. *Met. Rep., London*, 2, No. 16, 1955.

11. SAWYER, J.S.; Jet stream features of the earth's atmosphere. *Weather, London*, 12, 1957, p. 333.

12. MURRAY, R. and JOHNSON, D.H.; Structure of the upper westerlies; a study of the wind field in the eastern Atlantic and western Europe in September 1950. *Quart. J.R. met. Soc., London*, 78, 1952, p. 186.

13. 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. *Quart. J.R. met. Soc., London*, 79, 1953, p. 402.

14. HURST, G.W.; The profile of a jet stream observed 18 January 1952. *Quart. J.R. met. Soc., London*, 78, 1952, p. 613.

15. HURST, G.W. The profile of a jet stream observed 1 September 1952. *Quart. J.R. met. Soc., London*, 79, 1953, p. 407.

16. HARDING, J.; The profile of jet streams in the Middle East. *Met. Res. Pap., London*, No. 932, 1955.

17. MURRAY, R. and DANIELS, S.M.; Transverse flow at entrance and exit to jet streams. *Quart. J.R. met. Soc., London*, 79, 1953, p. 236.

18. NAMIAS, J. and CLAPP, P.F.; Confluence theory of the high tropospheric jet stream. *J. Met., Lancaster, Pa.*, 6, 1949, p. 330.

19. LEE, R.; Synoptic evidence for a direct circulation about a jet stream. *Quart. J.R. met. Soc., London*, 81, 1955, p. 462.

20. SAWYER, J.S.; Temperature, humidity and cloud near fronts in the middle and upper troposphere. *Quart. J.R. met. Soc., London*, 84, 1958, p. 375.

21. MURRAY, R.; Some features of jet streams as shown by aircraft observations. *Geophys. Mem., London*, 12, No. 97, 1956.

22. SAWYER, J.S. and ILETT, B.; The distribution of medium and high cloud near the jet streams. *Met. Mag., London*, 80, 1951, p. 277.

23. JOHNSON, D.H. and DANIELS, S. M.; Rainfall in relation to the jet stream. *Quart. J.R. met. Soc., London*, 80, 1954, p. 212.

24. Geneva, World Meteorological Organization; Fourth session of the executive committee, Geneva, 6th – 26th October, 1953. Abridged report with resolutions. W.M.O. Publication No. 20, RC.5, Geneva, 1953, p. 82.

25. London, Meteorological Office; Definitions of the tropopause. Meteorological Office Orders, Series A, No. 4, London, 1959.

26. SCHUMACHER, N.J.; The Maudheim expedition, 1949–52. *Weather, London,* 7, 1952, p. 327.

27. MOORE, J.G.; Average pressure and temperature of the tropopause. *Met. Res. Pap., London,* No. 874, 1954.

28. London, Meteorological Office; Upper air data for stations maintained by the Meteorological Office. London, 1952–57.

29. SAWYER, J. S.; Day-to-day variations in the tropopause. *Geophys. Mem., London,* 11, No. 92, 1954.

30. PALMÉN, E.; Aerologische Untersuchungen der atmosphärischen Störungen. *Mitt. met. Inst. Univ., Helsingfors,* No. 25, 1933.

31. MURGATROYD, R.J.; Winds and temperatures between 20 km and 100 km – a review. *Quart. J.R. met. Soc., London,* **83**, 1957, p. 417.

32. BANNON, J.K. and JONES, R. A.; The mean wind at 60 mb. *Met. Res. Pap., London,* No. 925, 1955.

33. KOCHANSKI, A. and WASKO, P. E.; Daily wind flow at the 50- and 25-mb levels. *Bull. Amer. met. Soc., Lancaster, Pa.,* 37, 1956, p. 8.

34. BANNON, J.K. and GILCHRIST, A.; Variation of temperature in the troposphere and lower stratosphere. *Quart. J.R. met. Soc., London,* 82, 1956, p. 58.

35. PANT, P.S.; Circulation in the upper atmosphere. *J. Geophys. Res., Richmond, Va.,* 61, 1956, p. 459.

36. DOBSON, G.M.B., BREWER, A.W. and CWILONG, B.M.; Meteorology of the lower stratosphere. *Proc. Roy. Soc., London,* **185**, Series A, 1946, p. 144.

37. MURGATROYD, R.J.; Wind and temperature to 50 km. over England. *Geophys. Mem., London,* 12, No. 95, 1956.

38. SCRASE, F.J.; Radiosonde and radar wind measurements in the stratosphere over the British Isles. *Quart. J.R. met. Soc., London,* 77, 1951, p. 483.

39. WEXLER, H.; Annual and diurnal temperature variations in the upper atmosphere. *Tellus, Stockholm,* 2, 1950, p. 262.

40. CRARY, A.P.; Stratosphere winds and temperatures from acoustical propagation studies. *J. Met., Lancaster, Pa.,* 7, 1950, p. 233.

41. WHIPPLE, F.J.W.; The propagation of sound to great distances. *Quart. J.R. met. Soc., London,* 61, 1935, p. 285.

42. MURGATROYD, R.J. and CLEWS, C.J.B.; Wind at 100,000 ft. over south-east England. *Geophys. Mem., London,* 10, No. 83, 1949.

43. BRASEFIELD, C.J.; Winds and temperatures in the lower stratosphere. *J. Met., Lancaster, Pa.,* 7, 1950, p. 66.

44. RICHARDSON, J.M. and KENNEDY, W.B.; Atmospheric winds and temperatures at 50-kilometers altitude as determined by acoustical propagation studies. *J. Acoust. Soc. amer., Menasha, Wis.,* 24, 1952, p. 731.

45. SCRASE, F.J.; Variations of temperature in the stratosphere up to 30 km over the British Isles. 10th general assembly, Rome 1954. *Sci. Proc. int. Ass. Met., U.G.G.I., Rome,* 1956, p. 514.

46. KOCHANSKI, A.; Cross sections of the mean zonal flow and temperature along 80°W. *J. Met., Lancaster, Pa.,* 12, 1955, p. 95.

47. LEE, R. and GODSON, W.L.; The arctic stratospheric jet stream during the winter of 1955–1956. *J. Met., Lancaster, Pa.,* 14, 1957, p. 126.

48. SCHERHAG, R.; Explosionsartigen Stratosphärenerwärmungen des Späswinters 1951/52. *Ber. dtsch. Wetterdienstes U.S. Zone, Bad Kissingen,* No. 38, 1952, p. 51.

49. CRAIG, R. A. and HERING, W.S.; The stratospheric warming of January – February 1957. *J. Met., Lancaster, Pa.,* 16, 1959, p. 91.

50. TEWELES, S. and FINGER, F.G.; An abrupt change in stratospheric circulation beginning in mid-January 1958. *Mon. Weath. Rev., Washington, D.C.* 86, 1958, p. 23.

51. TEWELES, S.; Anomalous warming of the stratosphere over North America in early 1957. *Mon. Weath. Rev., Washington D.C.,* 86, 1958, p. 377.

52. SCRASE, F.J.; Relatively high stratosphere temperature of February 1951. *Met. Mag., London,* 82, 1953, p. 15.

53. WEXLER, H.; Seasonal and other temperature changes in the Antarctic atmosphere. *Quart. J.R. met. Soc., London,* 85, 1959, p. 196.

54. FLOHN, H., HOLZAPFEL, R. and OECKEL, H.; Untersuchungen über die Stratosphärische Oströmung auf der Sommerhalbkugel. *Beitr. Phys. frei. Atmos., Leipzig,* 31, 1959, p. 217.

55. SAWYER, J.S.; The dynamical systems of the lower stratosphere. *Quart. J.R. met. Soc., London,* 77, 1951, p. 480.

BOOKS FOR FURTHER READING

PETTERSSEN, S.; Weather analysis and forecasting. New York, 1956. (2nd edition, Chapters 6 and 11).

GOODY, R.M.; The physics of the stratosphere. Cambridge monographs on physics. Cambridge, 1954.