HANDBOOK OF AVIATION METEOROLOGY

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PREFACE

THE primary object of this handbook is to provide a work suitable for the use of pilots and navigators undergoing intermediate or advanced courses of instruction at flying training schools, whether military or civilian. Within these limits the subject is covered comprehensively, but an elementary knowledge of meteorology on the part of the reader will be an advantage. The presentation has been made reasonably simply and physical ideas explained as they arise; mathematical knowledge is not required beyond an ability to handle some simple formulae, but for the benefit of readers familiar with elementary calculus, proofs of most of the formulae used are collected together in an appendix.

While the routine work of meteorology is in general performed by staff of the state meteorological services throughout the world, the maximum value is obtainable from the services provided only if the users are equipped with an adequate knowledge and understanding of the many meteorological processes and phenomena which may affect the operations of aircraft, and if they are aware of the means by which information on these matters is obtained by the meteorologist and made available to them. The range of subject matter necessary to meet these requirements is extensive; moreover it is still expanding as flying itself continues to develop. It is, however, neither possible nor desirable to confine the scope of such a book to the precise needs of students attending a variety of courses and requiring to pass one or other of the several examinations open to them. Accordingly, with but very little widening of the scope, the book has a secondary object, that of the presentation of a general account of meteorology, including its theory and practice and its applications to aviation. It is hoped, therefore, that it will be found useful to any who are in one way or another concerned with the applications of meteorology to aviation and who require a non-mathematical account of the subject; included among these potential readers will no doubt be many meteorological personnel themselves, but more especially those under training.

The book is divided into five parts which are more or less independent. Part I contains a somewhat detailed account of the physical principles of the subject, together with their immediate applications to aviation. Part II gives a brief description of the raw material of meteorology – the observations, how they are made, distributed and charted. This leads on to a discussion of synoptic meteorology including examples of synoptic charts in Part III, with an outline of the principles of weather forecasting. Part IV describes the organization of the meteorological services for aviation; and finally Part V explains and describes, very briefly, the salient features of weather over the world and on the air routes.

It is the intention that at a future date the Meteorological Office will adopt metric units completely, but for a time English units will remain in use for some purposes. In order that the reader may gain familiarity with the new units some metric equivalents have been added where considered significantly informative.

Visibilities are given in metres and kilometres, but as the foot is still widely used to express height in aviation circles, both feet and metres are freely used and conversions are given in some places. Horizontal wind speeds are in knots but vertical wind speeds are mostly in either metres, or feet, per second. It should be noted that in many instances the metric equivalents quoted are not necessarily exact conversions but represent values which will be used when observations are made in metric units; some conversion tables are given in Appendix IV.

The Celsius (also known as the Centigrade) scale of temperature has been adopted by the World Meteorological Organization and is used by the Meteorological Office. A few Fahrenheit equivalents are given (in brackets) in this book. The Kelvin scale of temperature is used in some physical equations.

The book was prepared as a successor to Dr R. C. Sutcliffe's *Meteorology for aviators* and as such naturally owes a very great deal to that earlier publication. The preparation of the manuscript was largely the work of Mr A. F. Crossley of the Meteorological Office but substantial assistance was given by other members of the Meteorological Office staff – notably by Dr A. G. Forsdyke – and by colleagues of the Ministry of Transport and Civil Aviation.

The thorough revision required to produce this second edition of the *Handbook of aviation meteorology* is largely the work of Mr E. D. Roberts of the Meteorological Office.

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PART I

PHYSICAL PRINCIPLES

CHAPTER 1

THE ATMOSPHERE

1. METEOROLOGY – A BRANCH OF PHYSICS

The meteorologist is concerned primarily with conditions within the atmosphere surrounding the earth, but the lithosphere, the earth's crust, and the hydrosphere, the oceans, must be considered also in so far as their condition interacts with that of the atmosphere. The whole surface of the earth and its atmosphere may be regarded as an enormous laboratory in which experiments are continually being performed, often on a grand scale and outside the scope of human control, but still the same in fundamentals as the experiments which the physicist performs and studies in his laboratory. The only sound foundation of an understanding of the subject is therefore a thorough grounding in the principles of physics, for example the laws of motion, of heating and cooling, of condensation and evaporation. With such knowledge it is hoped that all the complex behaviour of the atmosphere can ultimately be explained; we shall therefore take the results of observation and supply the physical explanations as far as possible, in the belief that the pilot or navigator can assimilate the facts which are important to him only if he appreciates the reasons underlying them.

Before proceeding to the account of the physical principles, it will be convenient to present a short description of the main atmospheric structure. This will show how the atmosphere may be regarded as consisting of various layers, and how those layers with which both weather and aviation are concerned stand in relation to the atmosphere as a whole. Moreover, the lower layers are for certain purposes idealized into a standard atmosphere, some particulars of which are also included in this chapter. The foundations will have then been laid for the explanation of the physical processes and their application to aviation.

2. COMPOSITION OF THE ATMOSPHERE

The atmosphere, in its dry state, is a mixture of many gases of which nitrogen and oxygen are by far the most abundant, accounting for almost 99 per cent of the whole in the proportion by weight of three parts of nitrogen to one of oxygen. Numerous analyses reveal no measurable variations in composition up to a height of at least 60 kilometres (37 miles) except in regard to ozone, but by about 70 kilometres (44 miles) gravitational separation begins to be effective. For the levels with which one is concerned in the study of weather, dry air may be taken as a uniform mixture in the proportions given in Table 1.

TABLE 1. *Percentage composition of dry air (by volume)*

		Traces only	
Nitrogen	78·09	Neon	Nitrous oxide
Oxygen	20·95	Helium	Ozone
Argon	0·93	Krypton	Sulphur dioxide
Carbon dioxide	0·03	Xenon	Nitrogen dioxide
		Hydrogen	Ammonia
		Methane	Carbon monoxide
			Iodine

Actually the atmosphere is never entirely dry, since water vapour is invariably present although in widely varying proportions. The vapour too behaves as a gas, and so long as it remains as vapour the assumption of dry air needs little modification; but frequently the vapour condenses into liquid or solid form, as in fog and mist, cloud and precipitation, so that much of the subject will be concerned with the changes of phase of the water content. Further, particles of dust, smoke and other impurities held in suspension affect the obscurity of the atmosphere and will be given consideration in the appropriate place.

3. STRUCTURE OF THE ATMOSPHERE

Troposphere

Although the composition of the atmosphere remains unchanged up to great heights, there are certain changes in the physical conditions which, with modern methods of observation, enable various layers to be distinguished (see Fig. 1). The lower layers are identified in the first instance by the rate of change of temperature with height. The average rate at which temperature decreases from the surface upwards is fairly uniform over the earth; and the fall continues regularly until it ceases more or less abruptly at a height of several miles which depends

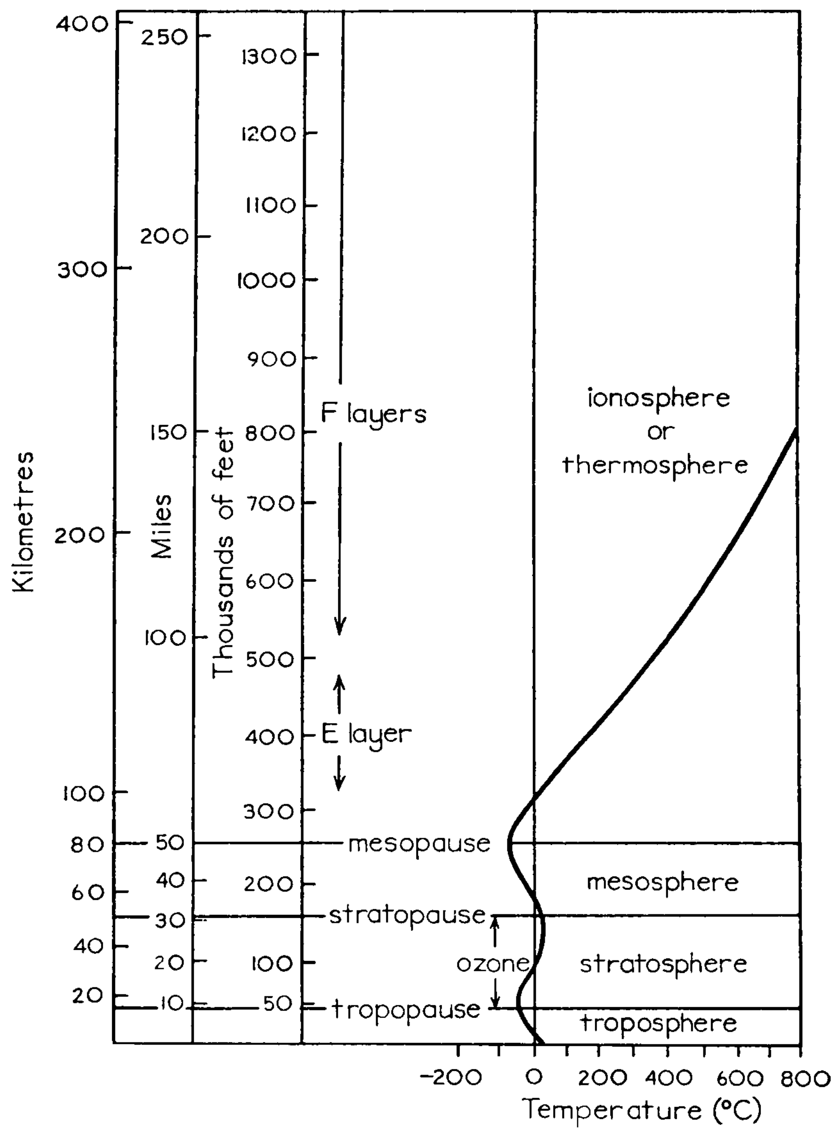


FIG. 1. *The atmosphere*

mainly on the latitude. This layer, characterized throughout by a marked fall of temperature with height, is called the troposphere and its upper boundary the tropopause.

The height of the tropopause varies with latitude, season and weather situation. In general, it is lowest (8–10 kilometres) in arctic regions in winter and highest (16–18 kilometres) in tropical and equatorial regions; over southern England it averages about 11 kilometres. Since air is compressible, the troposphere contains much the greater part (over three-quarters in middle latitudes) of the whole mass of the atmosphere, while the remaining fraction is spread out with ever increasing rarity over a height range some hundred times that of the troposphere.

Stratosphere

This layer extends from the tropopause to about 50 kilometres above mean sea level and is characterized by a temperature which is steady or increases with height; the increase is more noticeable near the top of the layer where the average temperature is only a little below 0°C. The boundary of the layer, where the temperature is at a maximum, is called the stratopause.

The relatively high temperatures at these levels are due to the presence of small quantities of ozone (O_3), an isotope of oxygen, which is a very strong absorber of ultra-violet radiation. If all this radiation were allowed to pass it would prove highly injurious to life as it has developed on earth. Thus these small quantities of ozone in the stratosphere (amounting to an equivalent thickness of only 3 millimetres if brought to sea-level pressure) are very important. It has been found that the total amount of ozone above any one place on the earth varies slightly from day to day and shows a significant correlation with certain weather phenomena. So far it has not been found possible to apply this knowledge to routine weather forecasting.

Mesosphere

This is the region of decreasing temperature above the stratopause. The mesosphere ends at the mesopause between 80 and 90 kilometres where the lowest temperatures of the atmosphere (about —90°C) are found.

Thermosphere

Above the mesosphere lies the thermosphere whose upper limit is undefined. The temperature increases rapidly with height up to about 200 kilometres and then increases more slowly or not at all. Above 200 kilometres the temperature varies widely according to solar activity; it is about 600°C when the sun is 'quiet' and possibly 2000°C during sunspot maxima.

The lower part of the thermosphere contains the zone known as the ionosphere because here the tenuous atmosphere is highly ionized by the action of solar ultra-violet and x-ray radiation. The resulting concentration of ions and free electrons is high enough to cause reflection of radio waves so making long-wave transmission possible over the earth's surface. The ionization reaches peak values at the levels known as the E, F and F_2 layers situated at about 110, 160 and 250 kilometres respectively, and associated with the names of Kennelly, Heaviside and Appleton. The ionosphere generally penetrates downwards into the upper part of the mesosphere.

The upper part of the thermosphere, roughly above 700 kilometres, is known as the exosphere, here the atmosphere is extremely tenuous and the mean free path of particles is so great that they can escape from the atmosphere. In this region temperature cannot be defined in the usual way.

While flight by aircraft is limited to the troposphere and stratosphere, rockets, possibly fitted with recording instruments, are not so restricted. However, in regard both to the study of weather and to flight by aircraft it is unnecessary to consider the mesosphere and thermosphere further, and with a few exceptions attention throughout subsequent chapters will be confined to the troposphere and stratosphere.

4. STANDARD ATMOSPHERES

International standard atmosphere (ISA)

For many purposes, such as the graduation of pressure altimeters and the design and testing of aircraft, the average state of the atmosphere needs to be represented in definite terms which can be used as a basis of reference. Such a representation is termed a standard atmosphere; it aims at specifying the average variation of temperature with height, from which the corresponding variations of pressure and density can also be given. It will be helpful to give details of a particular specification here although the general discussion of the variations of temperature, pressure and density with height must be left to subsequent chapters.

The standard atmosphere as defined by the International Civil Aviation Organization* comprises at mean sea level a temperature of 15°C, a pressure of

TABLE 2. *International standard atmosphere*
Surface density 1225 g/m^3

Height†	*Temperature*	*Pressure*	*Relative density*	*Height*†	*Temperature*	*Pressure*	*Relative density*
km	°C	*mb*	%	*ft*	°C	*mb*	%
32	−44·7	8·9	1·1	105 000	−44·7	8·9	1·1
30	−46·6	11·9	1·5	100 000	−46·2	11·1	1·4
27·5	−49·1	17·4	2·2	95 000	−47·7	13·9	1·8
25	−51·6	25·5	3·3	85 000	−50·7	22·2	2·8
22·5	−54·1	37·5	4·9	80 000	−52·2	28·0	3·6
20	−56·5	55·3	7·2	75 000	−53·7	35·4	4·5
17·5	−56·5	81·8	10·7	70 000	−55·2	44·9	5·8
15·0	−56·5	121·1	15·8	65 000	−56·5	56·9	7·4
12·5	−56·5	179·3	23·5	60 000	−56·5	72·3	9·5
10	−49·9	264·9	33·7	55 000	−56·5	91·8	12·0
7·5	−33·7	382·9	45·5	50 000	−56·5	116·6	15·3
5·0	−17·5	540·4	60·1	45 000	−56·5	148·2	19·5
2·5	− 1·2	746·9	78·1	40 000	−56·5	188·2	24·7
1·0	+ 8·5	898·7	90·7	35 000	−54·2	239·1	31·0
0·5	+11·7	954·6	95·3	30 000	−44·3	301·5	37·5
0	+15·0	1013·25	100·0	25 000	−34·5	376·5	44·8
−0·5	+18·2	1074·8	104·9	20 000	−24·6	466·0	53·3
				15 000	−14·7	572·0	62·9
				10 000	− 4·8	696·9	73·8
				5 000	+ 5·1	843·1	86·2
				2 000	+11·0	942·1	94·3
				0	+15·0	1013·25	100·0
				−1 000	+17·0	1050·4	103·0

†Heights are geometric heights.

*Montreal, International Civil Aviation Organization. Manual of the ICAO standard atmosphere, 2nd edn. Montreal, ICAO, 1964.

1013·25 millibars, and a density of 1225 grammes per cubic metre; the rate of fall of temperature with height is practically equivalent to 6·5 degC per kilometre, or 1·98 degC per 1000 feet, up to 11 kilometres (36 090 feet) and from there to 20 kilometres (65 617 feet) the temperature is assumed constant at —56·5°C. From 20 kilometres to 32 kilometres (104 987 feet) the rate of rise of temperature is equivalent to almost 1 degC per kilometre or 0·3 degC per 1000 feet. More details are given in Table 2.

This ICAO standard atmosphere has replaced the ICAN (International Commission for Air Navigation) standard which was previously adopted by many countries. The two standards differ only very slightly. Certain other standard atmospheres have been prepared to cover purposes such as aircraft design and testing for which the ISA specification may be inadequate (see also *Meteorological glossary*, **standard atmosphere**).

CHAPTER 2

PRESSURE

5. INTRODUCTION

The study of atmospheric pressure may be said to form the foundation of the science of meteorology. Weather is closely dependent on the distribution of pressure at the surface, and charts of sea-level pressure, supplemented by other charts for higher levels, constitute the basis of weather forecasting. Moreover, pressure differences provide the forces which are responsible for the generation of wind and the consequent changes in weather. The intimate connection between pressure and height is utilized in the pressure altimeter for the ready determination of height in the atmosphere.

Barometric pressure

The pressure of the atmosphere, as of any liquid or gaseous fluid, is the force exerted on a surface of unit area by the activity of the molecules composing the fluid. When the fluid in any region is at rest the motion of the molecules is entirely haphazard and the pressure is exerted uniformly in all directions; this is the static or barometric pressure. If the fluid were in motion, an additional pressure would be exerted on a surface of small area opposed to the direction of flow; this is termed the dynamic pressure, or more simply, in the atmosphere, the pressure of the wind. While meteorology is concerned mainly with the static pressure, the notion of dynamic pressure has some important applications including, in particular, the pitot tube by means of which the wind speed or the rate of movement of an object through the air may be measured. Generally when the word pressure is used without qualification it refers to the static or barometric pressure.

Pressure as the weight of air above

The atmosphere is held to the earth by gravitational attraction but is prevented from collapse by the molecular motions referred to above; hence any area of the earth's surface may be regarded as supporting the weight of the overlying air. The pressure of the atmosphere at a point on the earth's surface is therefore equivalent to the weight of the whole column of air standing on unit area at that point. Similarly if one considers a point at some particular height above the surface, it is seen that the pressure at that height is equal to the weight of the air above. This rule is of general applicability, and it may be shown to remain true within very wide limits when the air is in motion; the only exceptions are concerned with the presence of very violent air currents such as may occur in thunderstorms, and even then any modification required is of theoretical rather than practical interest.

Units

In meteorology the unit of pressure, which is a force per unit area, is derived from the metric system. The metric unit of force is the dyne, the force required to produce an acceleration of 1 centimetre per second per second in a mass of 1 gramme (the weight of, or force of gravity on, 1 gramme is about 981 dynes). The metric unit of pressure is therefore 1 dyne per square centimetre. This is inconveniently small for the measurement of atmospheric pressure, so the bar is defined as one million of these units – the average pressure at sea level is roughly of this magnitude. The millibar (mb), one-thousandth part of a bar or 1000 dynes per square centimetre, is the unit ordinarily used in meteorology.

Before the introduction of the millibar, pressure was measured in terms of the length of the column of mercury in the barometer and was reported as so many inches or millimetres of mercury. As this practice still continues to some extent and is common among physicists, the following conversion factors may be of use:

$$1000 \text{ mb} = 750{\cdot}1 \text{ mm} = 29{\cdot}53 \text{ in} = 10^5 \text{ N/m}^2.$$

N represents the SI (Système International) unit for pressure, the newton, and is equal to 10^5 dynes. Other values should be converted proportionately.

6. PRESSURE AT MEAN SEA LEVEL

Isobars and pressure systems

In representing the distribution of pressure at a given time, it will be clear on account of the marked reduction of pressure with height that it is essential for the observations to be reduced to a common level (usually mean sea level). Isobars or lines of equal pressure, usually at 2-millibar or 4-millibar intervals, are drawn in accordance with the reduced observations to form a synoptic chart from which the pressure distribution may be perceived at a glance. Further details regarding the drawing of isobars are given in Chapter 14 while examples of synoptic charts occur in Part III. Experience with synoptic charts shows that the isobars take up various configurations, each of which has its own characteristics in regard to wind and weather. These pressure systems are described briefly below and are illustrated in Fig. 2.

From an inspection of a sequence of synoptic charts the pressure systems are seen to undergo continuous modifications and to move about from place to place; occasionally one disappears, or a new one is formed. Lows may be said to develop, deepen, intensify, dissipate, decay or fill up; highs to build up, intensify, give way, weaken or collapse.

Depression or low (or occasionally cyclone, this last word being more commonly used in its adjectival form) (L in Fig. 2). A region of relatively low pressure shown by more or less circular and concentric isobars surrounding the centre, where pressure is lowest. The size of the low may vary from a few hundred feet in a tornado to some hundreds of miles in a tropical revolving storm (when the term cyclone is appropriate) and to well over a thousand miles in the larger systems of temperate latitudes. Lows are often classed as deep or intense, shallow or weak, but these are relative terms only.

Secondary depression or low (S in Fig. 2). A small low within the area covered by a larger or primary low and appearing rather as a satellite. The isobars need not show a closed centre, and the term may be applied to any small region of relatively low pressure which maintains its identity for a time.

Trough of low pressure (T in Fig. 2). Indicated by isobars extending outwards from an area of low pressure so that the pressure is lower in the trough than on the two sides. When the isobars change suddenly in direction at the trough they may be described as V-shaped, although this name for a pressure type has fallen into disuse.

Anticyclone or high (H in Fig. 2). A region of relatively high pressure shown by more or less circular isobars similar to those of a low but with the highest pressure at the centre. The isobars are often more widely spaced than in the low, particularly near the centre.

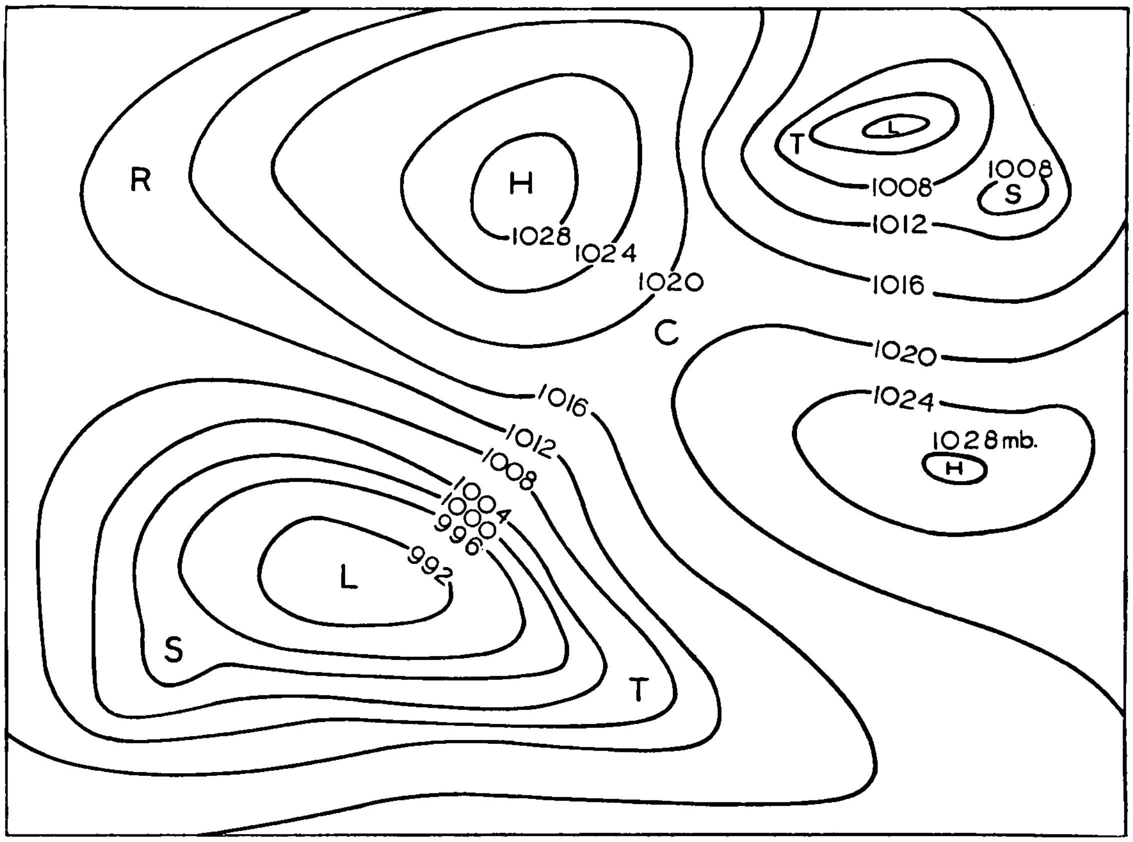


FIG. 2. *Various types of pressure distribution*

Ridge, wedge or tongue of high pressure (R in Fig. 2). Indicated by isobars extending outwards from a high and always rounded, never V-shaped as sometimes in a trough.

Col (C in Fig. 2). The region of almost level pressure between two highs and two lows.

Pressure gradient

The slope or gradient of the surface of the ground has its analogy in the field of pressure. The steepest slope on a contour map is at right angles to the contour lines, and is greater or less according as the lines are close or far apart; and in exactly the same way we speak of the gradient of pressure as being directed at right angles to the isobars. A field of pressure is said to be flat when the differences are slight and the isobars far apart; the gradient is steep when the isobars are close together.

Barometric tendency and isallobars

One of the elements plotted on the weather map is the tendency or change of pressure during the period of three hours preceding the time of observation. It is of great practical value in forecasting as it enables one to see at a glance in what areas pressure is rising or falling. Just as the field of pressure is made to stand out clearly by drawing isobars, so the distribution of the tendencies can be shown by drawing lines of equal tendency; these are known as isallobars. Regions of falling pressure may thus be shown by closed curves similar to the isobars of a depression, the greatest falls being at the centre, and it is customary to describe such an area as an isallobaric low; similarly, an isallobaric high is a region of rising pressure.

Diurnal variation of pressure

Changes of pressure occur almost continuously and are due to many causes, including the movement and development of depressions, anticyclones, or other systems; but superimposed on the irregular trends it is often possible to detect a rhythmical oscillation. In middle and high latitudes irregular changes are usually sufficient to obliterate the oscillations, but they may be recognized during quiet settled weather. In England the daily range averages less than 1 millibar, but at the equator it is about 3 millibars while in polar regions it is negligible. The oscillation may best be represented as a double wave travelling round the earth following the sun, having minima at 0400 and 1600, with intermediate maxima at 1000 and 2200 local time. The oscillations give therefore a semi-diurnal variation, but the changes are not perfectly symmetrical and vary considerably with locality. The variations have little influence on other meteorological factors but their occurrence needs to be kept in mind when interpreting the three-hourly pressure changes plotted on synoptic charts, otherwise the diurnal effect might give a false impression of the changes taking place. In the tropics and subtropics these regular variations are the predominant feature. In the regions subject to tropical revolving storms any departure from the regular oscillation is often the first warning of the approach of a storm.

The explanation of the semi-diurnal variation has been difficult, but recent research has indicated that its origin is probably a natural oscillation of the atmosphere which happens to have a period of almost exactly 12 hours. Such an oscillation would be excited and maintained by resonance produced by the 24-hour variation of temperature.

7. PRESSURE AND HEIGHT

Variation of pressure with height

From the rule given in Section 5 to the effect that the pressure at any point is equal to the weight of air above, it follows that the pressure at any height above the surface of the earth is less than the pressure at the surface itself by the weight of the column of air of unit cross-section extending from the surface up to that height. This implies that pressure invariably decreases with height in the atmosphere. The amount of the reduction up to a given height, being simply the weight per unit area of the air column up to that height, depends on the density or temperature of the air. This is illustrated in Fig. 3 which represents an imaginary vertical section through the atmosphere. The pressure at ground level is taken as uniform (p_0) but the air above A is supposed warmer than that over B (level for level). Up to some height h, the weight of the column AC is therefore less than that of BD, consequently the reduction of pressure from A to C is less than that from B to D, and pressure at C is greater than at D. Therefore at height h the pressure decreases from C to D, the higher pressure lying over the warmer air, and the lower pressure over the colder air. Further, if the pressure at C is p_1, this value of the pressure is reached at some lower level, F, in the cold air; and if the pressure at D is p_2, then this pressure is not attained in the warm air until some higher point E is reached. In this way we can indicate the lines of equal pressure CF, ED, etc. in a cross-section of the atmosphere, and find that they are in general sloping (or curved) lines, even if the surface pressure should, exceptionally, happen to be uniform. The ideas of this paragraph will be found later to have an application to the variation of wind with height (Section 26).

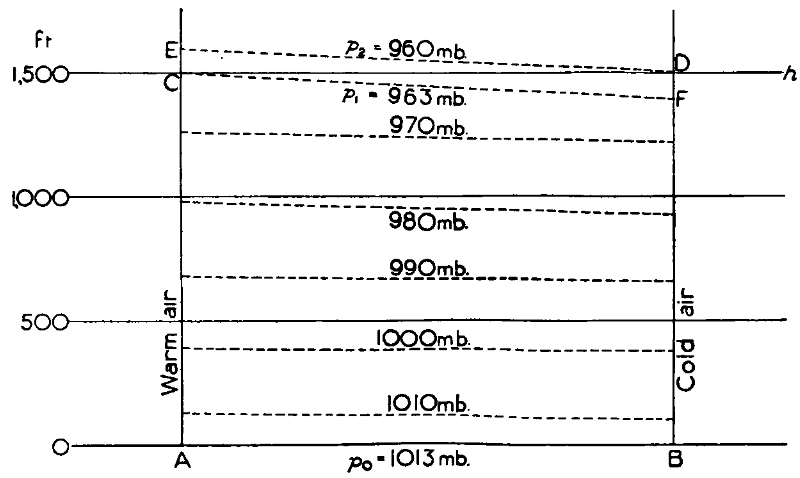


FIG 3. *Variation of pressure with height*

The general fall of pressure with height in the international standard atmosphere may be noted from Table 2 (Section 4). It is also useful to know what increase of height corresponds with a fall of pressure of 1 millibar for any particular values of pressure and temperature. This is given by the equation (Appendix I)

$$\text{height difference in feet for 1 mb change or pressure} = 96\,\frac{T}{p}, \qquad \ldots\ldots \quad (1)$$

where T is the temperature in degrees Kelvin (degrees Celsius+273) and p the pressure in millibars. Some values derived from this formula are given in Table 3 for selected temperatures and pressures. It is seen that a pressure change of 1 millibar in the international standard atmosphere is roughly equivalent to a height interval of 27 feet at 1000 millibars, 50 feet at 500 millibars and 100 feet at 200 millibars.

TABLE 3. *Height difference corresponding to a pressure difference of 1 millibar*

	Pressure (millibars)											
Temperature	1050	1000	900	800	700	600	500	400	300	200	100	50
°C						*feet*						
40	29	30	33	38	43	50	60	75	100	150	301	601
0	25	26	29	33	37	44	52	66	87	131	262	524
—40	21	22	25	28	32	37	45	56	74	112	224	448
—80	18	19	21	23	26	31	37	46	62	93	185	371

Another formula is required to express the difference between the heights h_1 and h_2 in terms of the corresponding pressures p_1 and p_2. When the temperature T is constant it may be shown that (Appendix I)

$$h_2 - h_1 = 221\cdot 1\ T\,(\log p_1 - \log p_2), \qquad \ldots\ldots \quad (2)$$

where the heights are in feet and the temperature in degrees Kelvin. When the temperature is not constant, T may be regarded as the mean temperature of the layer. This equation may be used to determine either the change of pressure between two given heights or the change of height between two given pressures. Its use is illustrated in the following example.

Example of pressure–height calculation. Given the following observations of upper air temperature: mean sea level 1016 millibars, 12°C; 1000 millibars, 14°C; 900 millibars, 9°C; 800 millibars, 6°C; 700 millibars, 2°C; what is the height of the 700-millibar level?

To a close approximation this is best done by finding the arithmetic mean temperature of the columns of air between successive observations and then using equation (2) to obtain the thickness of each column, finally adding the results to give the height of the 700-millibar level above the surface. Thus the mean temperature between $p_1 = 1016$ and $p_2 = 1000$ millibars is 13°C or 286°K; hence by the formula the thickness of this layer is 436 feet. Similarly the mean temperatures of the succeeding 100-millibar layers are 284·5°, 280·5° and 277°K and the thicknesses 2879, 3172 and 3552 feet respectively. The total of these, 10 039 feet, is the height of the 700-millibar level above mean sea level.

A quicker but less accurate method is to estimate the mean temperature of the whole column from the surface to 700 millibars and to use this figure in the formula with $p_1 = 1016$ and $p_2 = 700$ millibars. To find the mean temperature for this purpose, the mean temperatures of the layers as determined above should be added together and divided by the number of layers; this gives 282·5°K, from which the total thickness is determined as 10 088 feet. The difference from the figure obtained by the more detailed method arises because the value used for the mean temperature of the whole air column is only an approximation.

An alternative method of calculating the height of a pressure surface will be described later in connection with the tephigram (Section 15).

Reduction of pressure to mean sea level

Surface observations are necessarily made at various heights above mean sea level. As a pressure difference of as little as one millibar is of consequence in the construction of synoptic charts and may be caused by an elevation of only 30 feet, it is clear that the pressures must be reduced to a common level. With the choice of mean sea level as the standard, the corrected pressure is in most cases a hypothetical value since it usually requires surface pressure to be increased by the weight of the supposed column of air between ground level and sea level. The correction is that appropriate to a certain standard temperature with an adjustment, depending on the air temperature at the time, which allows for variation from the standard. This gives reasonably consistent results for heights up to 1500 feet, but for higher levels the reduction becomes less reliable. British practice in such cases is to report the pressure reduced to station level, but over high plateau country such as East Africa some more appropriate standard level may be used.

8. ALTIMETRY

The close connection between pressure and height is made use of in the construction of the pressure altimeter which in principle is merely an instrument for measuring pressure, that is, an aneroid barometer with its scale graduated to read height in feet instead of pressure in millibars or inches. It has been seen (Section 7) that the relationship between pressure and height is not invariable, but depends both on the surface pressure and on the weight or temperature of the air up to the height concerned, factors which are constantly changing with time and place. Accordingly the altimeter can indicate the height accurately only in certain specified conditions; whenever the temperature or pressure departs from these, either an adjustment to the setting or a correction to the reading, or both, must be made to obtain the true height. The instrument has a linear height scale so designed as to indicate the correct height when the conditions are those of the ICAO standard atmosphere which has commonly replaced the almost identical ICAN standard (see Section 4). As the instrument, of which there are various types, is in general use for the determination of height in aircraft, the effects of differences of pressure and temperature from standard will be considered at some length. For this purpose it will be assumed that the instrument is properly adjusted and exposed in an aircraft and that it is in fact responding correctly to the atmospheric pressure at the position of the aircraft. The instrumental errors are considered on page 16.

Zero correction for pressure variations

Altimeters used in aircraft are fitted with a subscale in millibars (or sometimes inches) which, when the indicated height is zero, reads the barometric pressure at the position of the instrument. This subscale can be adjusted so that the indicated height (for example, above ground, or above sea level) is correct at time of take-off and can if required be readjusted during flight or in preparation for landing. Apart from a possible correction for temperature, the indicated height remains approximately correct after being 'set for zero' only so long as the surface or sea-level pressure directly beneath the aircraft remains unchanged. In practice this is seldom the case. Although set correctly when commencing a flight, the altimeter may be noticeably in error on landing either at the same or at some other aerodrome. These 'zero errors' are due to changes in the barometric pressure with time and place. The altimeter will over-read if the barometric pressure falls during the flight and under-read if it rises, the amount of the error being roughly 27 feet for each millibar change in surface pressure. To take an example, suppose the altimeter were set to a sea-level pressure of 1010 millibars; then if pressure falls to 1000 millibars on landing a height of some 270 feet above sea level would be recorded, while if pressure rises to 1020 millibars the indicated height will be some 270 feet below sea level. As a general rule it should be remembered that if the flight is towards an area of higher pressure the altimeter under-reads. On the other hand, when flying towards an area of lower pressure the altimeter over-reads; this case needs careful attention as the apparent clearance over ground or other obstacle may be more by some hundreds of feet than the actual clearance.

Again, suppose a flight is made at a constant indicated height without any change of subscale setting. Since the altimeter responds only to changes of pressure, it follows that the flight takes place at a constant pressure level and that there is a gain or loss of height above sea level according as the aircraft is flying towards an area of higher or lower barometric pressure. This is illustrated in Fig. 4 for an aircraft which leaves London with altimeter set to a sea-level pressure of 1019 millibars and flies at an indicated height of 4000 feet. Pressure is low to the north, and on arrival over Prestwick the sea-level pressure there has fallen to 991 millibars. This is 28 millibars less than at take-off, and the indicated height (ignoring differences of temperature from standard) over-reads by 770 feet.

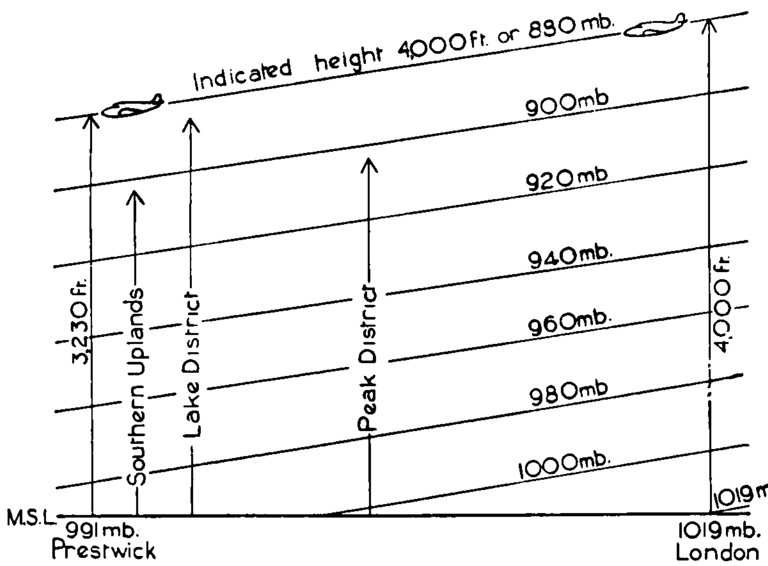


FIG. 4. *Diagram showing need for zero correction to altimeter for surface pressure variations*

When the aircraft arrives at Prestwick the indicated height over-reads by 770 feet.

Results of this type can be related with the wind which is experienced. In the case discussed, the wind throughout the flight will have been blowing from port in accordance with Buys Ballot's law (Section 24). The associated over-reading of the altimeter constitutes an example of the following practical rule: when flying with winds from port in the northern hemisphere, the altimeter tends to over-read, and with winds from starboard to under-read. The error increases with the strength of the wind and with the length of route. In the southern hemisphere the sign of the error is reversed.

Corrections for temperature variations

While the application of a correction for change of surface pressure ensures a correct indication of height at sea or ground level according to the setting, the reading at other levels is still subject to an error, although in many cases a comparatively small one, depending on the difference between the actual and standard temperatures. Thus in the conditions illustrated in Fig. 3, the indicated height will remain constant for a flight from B to A at 960 millibars (represented by the line DE) while the true height will increase, the difference corresponding to CE when the aircraft is over the point A where the air is warmer than at B (it is assumed that the surface pressure is uniform).

If the zero height corresponds to pressure p_0, and height h to pressure p, then from equation (2) in Section 7 the true height at pressure p is given by

$$h = 221 \cdot 1\ T (\log p_0 - \log p)$$

and the indicated height by

$$h' = 221 \cdot 1\ T_i (\log p_0 - \log p),$$

where T is the mean temperature in degrees Kelvin of the air column from p_0 to p, and T_i is the mean standard temperature within this range of pressure. The true height is therefore given by

$$h = \frac{T}{T_i} h' = \left(1 + \frac{T - T_i}{T_i}\right) h'. \qquad \ldots\ldots (3)$$

In practice, $T—T_i$ in this expression is replaced (as an estimate) by the difference between the observed and standard temperatures at the indicated height; if then the observed air temperature is set against the indicated height on a navigational computer the corrected height may be read off on another scale.

For example, if the temperature exceeds the standard temperature by 10 degC, the indicated height requires to be increased by roughly 4 per cent. Generally the correction is comparatively small for heights within a few hundred feet of the setting level, at least in comparison with the errors produced by variations in surface pressure, but since the temperature correction increases in proportion to the indicated height, the error at higher levels may be considerable, for example, 5 per cent at 15 000 feet is 750 feet. For this reason alone it is desirable to pay close attention to the temperature correction at altitude and to maintain ample clearance when flying over mountainous country in bad weather.

Instrumental errors of the pressure altimeter

Apart from the errors of meteorological origin already described, the instrument is subject to other errors which may be summarized as follows:

Scale error. This is due to faulty calibration. It can be eliminated in the instrument workshops.

Friction error. This causes irregular and jumpy movements of the pointer even when pressure is changing uniformly. It may be overcome by gentle tapping or by the normal vibration of the instrument panel.

Hysteresis error. This shows as a lag in the indicated reading after a rapid climb or descent, and is due to imperfect elasticity of the aneroid capsule. The error increases both with rate of change of pressure (or height) and with the magnitude of the change. It assumes great importance for military aircraft when they are executing low-level attacks, particularly at night.

Temperature error. This is the change in reading due to a change in temperature of the instrument – not to be confused with a change in air temperature. It is usually counteracted by the inclusion of a compensating bimetallic strip.

Position error. The instrument is connected to a static vent fitted to the surface of the aircraft where the pressure may differ from the true static pressure of the atmosphere.

Radio altimeters

Although these instruments record height independently of the meteorological conditions, it is convenient to refer to them here. There are two types in common use, one depending on the pulse radar system and the other on frequency modulation. With both types the height above the earth's surface is indicated in flights over the sea or level ground. The pulse radar type, often referred to simply as the radar altimeter, measures height over a range of the order of 1000 to 50 000 feet. The error in the absolute measurement is within ± 100 feet, but changes of height at an approximately constant level are measurable to within about ± 30 feet. The frequency-modulation type, sometimes referred to as the radio altimeter, covers the range from zero to about 5000 feet; the error of this instrument is proportional to height and may amount to ± 3 per cent. Both types are needed to cover the whole range of height. The pulse radar altimeter may be used in conjunction with the pressure altimeter in pressure-pattern flying (Section 30) and for other purposes.

Pressure altitude

This is defined as a pressure expressed in terms of the height to which it corresponds in the international standard atmosphere. It is therefore the height indicated by a pressure altimeter set to 1013·2 millibars.

Altimeter correction

This is defined as the amount which must be added to the indicated height to obtain the true height above sea level, h. When the subscale is set to 1013·2 millibars, the indicated height is the pressure altitude, h', and the altimeter correction is then the difference between the true height above sea level and the pressure altitude, or $h—h'$. Apart from observational error, the altimeter correction is given when flying over the sea by the difference between simultaneous readings of the radio and pressure altimeters, the subscale of the latter being set to 1013·2 millibars; it is sometimes denoted by D.

9. ALTIMETER SETTINGS

There are several methods of setting the subscale of an altimeter in order to eliminate the effect of the difference between a surface or sea-level pressure and the value according to the international standard atmosphere. The pressure which is set on the subscale is usually referred to by its abbreviation in the international Q code (see Section 151); short definitions of the relevant groups are given below:

QFE: Barometric pressure at the level of an aerodrome.

QFF: Barometric pressure at a stated place, reduced to mean sea level.

QNH: The pressure setting which causes the altimeter to read the height above mean sea level of the touchdown on landing, plus the height of the altimeter above the ground.

Forecast QNH: Forecast, valid for one hour, of lowest QNH expected in any part of the altimeter setting region.

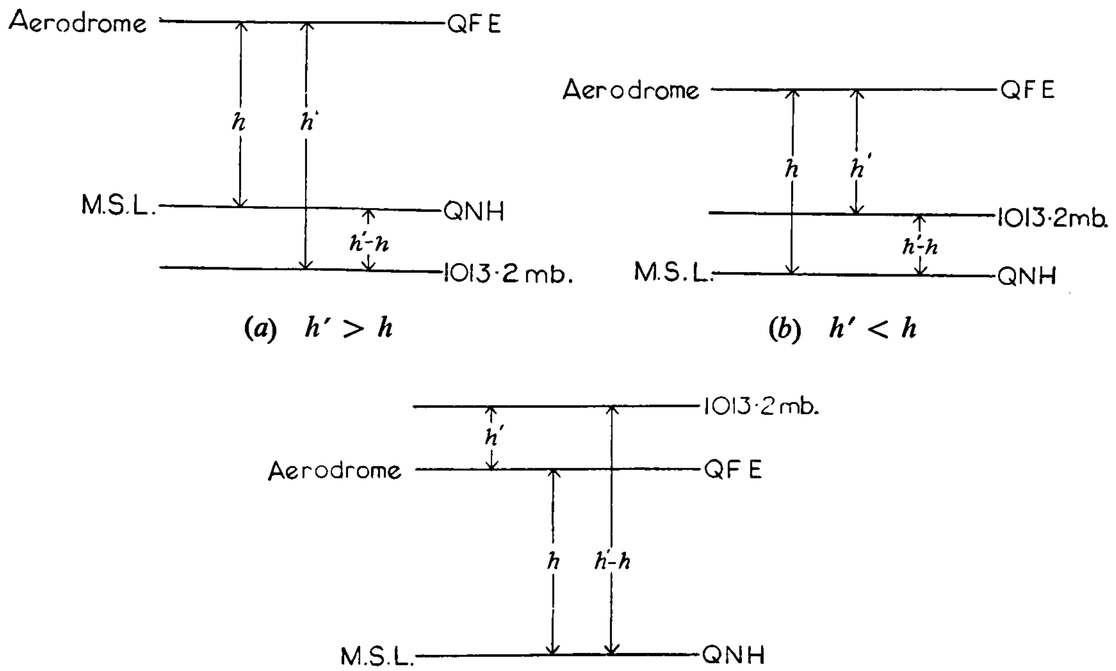
QNE: The height indicated on landing at an aerodrome when the altimeter subscale is set to 1013·2 millibars (29·92 inches).

In these definitions it is assumed that the altimeter is free from instrumental errors.

The first two of the above settings are obtained by reducing the observed pressure at the level of the barometer cistern to the recognized level of the aerodrome (QFE) or to mean sea level (QFF). The reduction is made in accordance with normal meteorological practice as described in Section 7.

The determination of QNH requires the observed barometric pressure to be reduced in accordance with the conditions of the international or ICAO standard atmosphere. If h is the height of the aerodrome above mean sea level and h' the height in the standard atmosphere which corresponds with the pressure (QFE) at aerodrome level, then the altimeter (when set to 1013·2 millibars) has an error of $(h'—h)$ (see Fig. 5). Since the height scale is linear, the value of QNH is then obtained as the pressure in the standard atmosphere corresponding with the height

$(h'—h)$. A table giving details of the international standard atmosphere may be used for this purpose; alternatively a table may be prepared for any aerodrome giving the value of QNH directly from the reading of the barometer.



(a) $h' > h$ (b) $h' < h$

(c) h' negative

FIG. 5. *Determination of QNH*

h' is the height of QFE in the standard atmosphere and QNH is the pressure corresponding to the height $h'—h$.

Example (*a*) – MSL pressure is less than 1013·2 millibars and $h'—h$ is positive.

Example (*b*) – MSL pressure is greater than 1013·2 millibars and $h'—h$ is negative.

Example (*c*) – Both QFE and QNH are greater than 1013·2 millibars, and h' the 'height' of QFE in the standard atmosphere is negative and $h'—h$ is also negative.

If with the altimeter set to 1013·2 millibars, the pilot requests QNE, the reply states what height the altimeter will read on landing at the aerodrome concerned. This is simply the height in the standard atmosphere corresponding with the pressure QFE. The use of QNE was introduced as a substitute for QFE at high-level aerodromes to cover occasions when the subscale did not have sufficient range to set QFE. With the introduction of an extended subscale and the use of alternative settings, QNE is now obsolescent.

Choice of altimeter setting

While on occasions the pilot of an aircraft may be free to exercise his personal preference in choosing an altimeter setting, at other times he may be required to adopt a setting prescribed by Air Traffic Control. In either case it is desirable that the pilot should be aware of the advantages and disadvantages of whatever setting is used. The main considerations influencing the choice of setting are concerned with landing and take-off, with cruising flight and the type of terrain, and with the flight of aircraft while under Air Traffic Control.

An altimeter set to QFE reads zero (strictly, the height of the altimeter above ground) when the aircraft is on the ground; this is sometimes referred to as the 'zero' setting and for obvious reasons has been widely used. While in the air, the QFE setting gives approximately the height of the aircraft above the aerodrome, but this information becomes of less use, or even misleading, when away from the aerodrome. Even when carrying out circuits and landings it may lead to difficulty, but still more so on a cross-country flight.

There is much to be said in favour of a setting which gives approximately the height above some recognized level, for which mean sea level is the obvious choice. Such a setting would be provided by QFF, but as this is the pressure reduced to sea level in accordance with isothermal conditions it leads to inaccuracies when used with the pressure altimeter. The error in the indicated height is small for heights near mean sea level, but the discrepancy increases in proportion to height and to the difference between the standard and actual conditions.

These difficulties are minimized by the use of QNH, since with this setting the reduction is made according to the standard atmosphere and the instrument therefore reads correctly on landing. It is the setting naturally suited to the construction of the pressure altimeter and gives the most satisfactory relation between indicated and true height under all circumstances.

Whatever altimeter setting is used, indicated altitudes in flight are subject to the effect of variations in temperature and pressure with time and place. Such variations are overcome in the United Kingdom by the regulation that aircraft cruising between heights within a flight information region (FIR) are required to use the forecast QNH value applying to the particular altimeter setting region (ASR) and are to make the appropriate change of setting on crossing the boundary between one ASR and another. When beginning the approach to land, a change may be made to some other setting. Because forecast QNH (the lowest QNH expected) values are issued hourly for each ASR, these procedures adequately cover the changes of sea-level pressure with time and place throughout the ASR concerned.

CHAPTER 3

TEMPERATURE

10. INTRODUCTION

Temperature is commonly measured according to one of two scales which are arbitrarily fixed with reference to the melting of ice and the boiling of pure water at normal pressure. On the Celsius or Centigrade scale these are taken respectively as 0°C and 100°C; this scale has been adopted internationally and is in general use for scientific purposes and in aviation. The Fahrenheit scale is defined with the melting point at 32°F and boiling point at 212°F; use of this scale for meteorological and other purposes is confined to some English-speaking countries. A third scale has a zero based on the following argument.

Heat is a form of energy. As heat is extracted from a substance, the internal energy must be reduced; this means that the random motions and vibrations of the molecules are reduced with the result that the molecules become arranged in a somewhat more orderly pattern than before. As more heat is extracted the cooling and rearrangement proceed further until eventually a state is reached at which the molecules have attained their maximum orderliness. In this state it is impossible to extract any more heat and the temperature has therefore reached its lowest possible value. Thermodynamic considerations show that this minimum temperature is the same for all substances, and it is accordingly referred to as the 'absolute zero'; its value is —273·16°C but for meteorological purposes it is often taken as —273°C.

The idea of the absolute zero is important for theoretical work and it forms the basis of the absolute or Kelvin scale on which the melting-point is approximately 273°K and boiling point 373°K. On the Fahrenheit scale the absolute zero is at —459°F, but the Fahrenheit absolute scale is rarely used. It is seen that an interval of 100 Celsius degrees corresponds with one of 180 Fahrenheit degrees so that if each of the letters K, C and F denotes one degree on the respective scale, the conversions from one scale to another may be written as follows:

$$F=\frac{9C}{5}+32, \qquad C=\frac{5}{9}(F-32), \qquad K=C+273.$$

With regard to differences of temperature, it may be noted that 9 Fahrenheit degrees correspond with 5 Celsius or absolute degrees.

This chapter not only includes a general description of the distribution of temperature and its variation with height, but attempts an explanation of how this distribution comes about. This requires some account to be given of the methods of heat transference in the atmosphere, especially by radiation and by adiabatic processes, while consideration of the latter makes this a convenient place to introduce the tephigram and to derive the basic criteria for vertical stability. The final section explains the direct effects of temperature on the operations of aircraft.

11. RADIATION

The molten interior of the earth is enclosed in a solid crust through which heat penetrates only slowly by conduction except here and there in volcanic eruptions and hot springs. The temperature increases downwards from the surface at an average rate of 1 degC per 100 feet, but the flow of heat upwards to the ground is insignificant compared with that received from the sun by day or lost by radiation at night. The temperature of the earth's surface and atmosphere depends in the last resort entirely upon heat received from the sun, only a negligible amount coming from the stars. Since the available evidence indicates that the mean temperature of the earth's surface has remained substantially unchanged for a long period, it follows that all the heat received from the sun must be returned to space by means of terrestrial radiation, and consequently that the average temperature of the surface and atmosphere becomes automatically adjusted to bring about the necessary balance between incoming and outgoing radiation.

In order to understand the processes of heating and cooling of the earth and atmosphere it is necessary to know something of the properties of radiation in general. Every body, whatever its state, emits radiation in the form of electromagnetic waves which travel through space at a uniform speed of 186 000 miles per second, identical with the speed of light. By analogy with waves on the surface of water, the radiant waves are characterized both by their wavelength or frequency, and by their energy or amplitude of vibration. On account of the uniform velocity of propagation, the wavelength is inversely proportional to the frequency or number of waves passing any point per unit time. Most bodies emit radiation over a wide range of wavelengths simultaneously, the relation between energy and wavelength depending both on the temperature and on the nature of the source. The distribution of energy in the spectrum of a theoretically perfect radiator, the so-called 'black body', is illustrated in Fig. 6 for temperatures roughly appropriate to the solar and terrestrial surfaces, both of which radiate practically as black bodies. It may be noted also that the total radiation emitted by a black body is proportional to the fourth power of the absolute temperature (Stefan's law).

The electromagnetic waves used in transmission by radio and radar are essentially of the same character as those concerned in solar and terrestrial radiation but of very much greater wavelength. The ranges of wavelength in the solar and terrestrial categories are shown in Fig. 6; for radar transmissions the range is from about 1 centimetre to 10 metres and for radio from about 5 to 30 000 metres. All these varieties of electromagnetic radiation are affected to some extent and in varying degree by the atmospheric conditions. The shortest waves, as we have seen, are of fundamental importance for heat transference; the radio waves have no significant effect on the atmosphere, but may be absorbed or refracted in the upper air (Section 3), while the behaviour of radar waves near the surface has given rise to the subject of 'radio meteorology' of which a brief account is given in Section 77.

The dependence of the character of radiation on temperature may be illustrated by the behaviour of a piece of metal when heated. It first emits radiation which is invisible but may be felt as heat; as the temperature rises the metal begins to glow with a dull red colour, and with further heating the colour brightens until ultimately the metal becomes white hot. These changes in colour indicate a progressive reduction of wavelength; in general, the higher the temperature the shorter is the effective wavelength and the greater the intensity of the emitted radiation. Now the temperature of the surface of the sun being some 6000°C, the solar radiation is mainly in the form of short waves visible to the human eye, that is ordinary light. Seen as a whole the sun's light appears white, but when the various wavelengths are separated, as by refraction through a prism, the white light is seen to be composite, the longest waves appearing red, the shortest violet, with intermediate colours

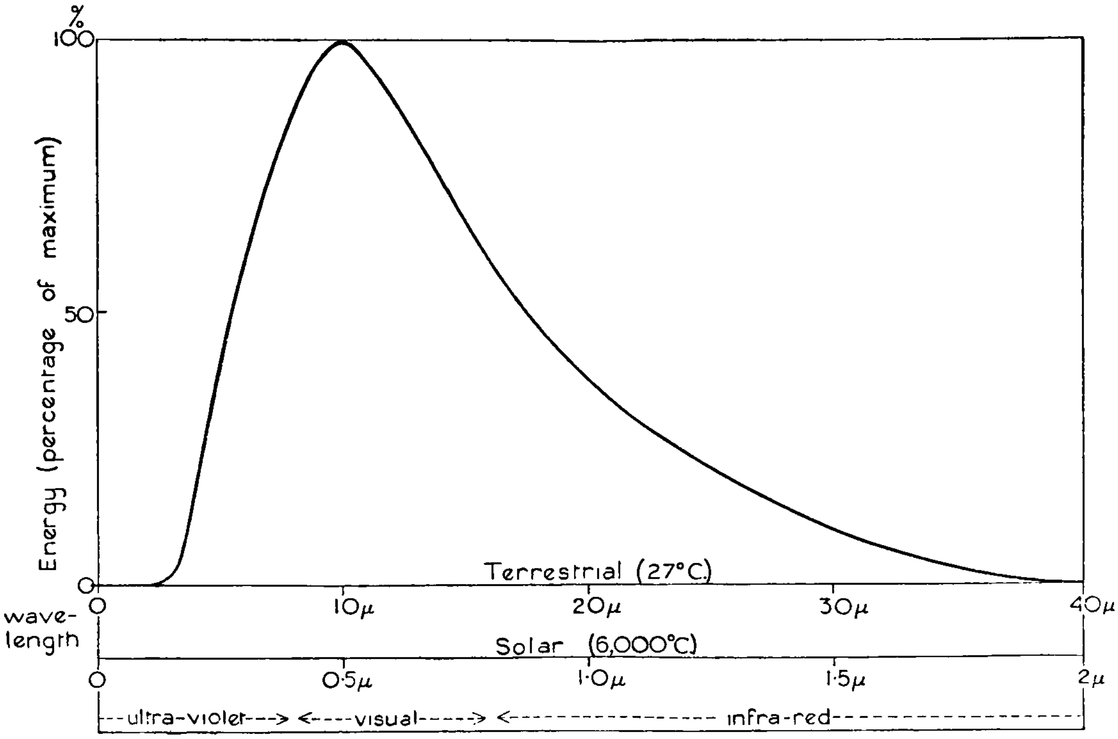


FIG. 6. *Energy and wavelength in the black-body spectrum at terrestrial and solar temperatures*

($\mu = 10^{-4}$ cm).

completing the familiar spectrum. Although the greater part of the solar radiation is comprised within this visible spectrum there is a part with wavelength too short to be observed by the eye; this is beyond the violet end of the spectrum and is known as ultra-violet light. At the other extreme there is a fringe of waves too long to be visible and therefore known as the infra-red. The complete solar radiation, including visible light, the ultra-violet and the infra-red, is responsible for all the heat which the earth receives from outside, and may be broadly described as short-wave radiation. This heat is lost again by radiation from the earth but in the form of relatively long waves, for terrestrial temperatures are much too low to give visible radiation. The earth therefore receives heat by short-wave radiation from the sun but loses it again by means of long waves.

Effect of solar radiation on the earth and atmosphere

Of the total radiation reaching the outer atmosphere from the sun, about one-third is reflected back to space by the earth or by the upper surface of clouds and plays no further part in terrestrial problems. This reflecting power is referred to as the 'albedo' of the earth. Some of the radiation is absorbed in the upper atmosphere, particularly by ozone in the ultra-violet range of wavelengths, and a small proportion is absorbed by clouds, but in clear weather about $\frac{5}{6}$ of the solar energy passes through the atmosphere to the earth's surface. Of this energy about 10 per cent is usually reflected, but over a snow surface reflection may amount to 80 per cent, and it is also large over the sea for small angles of incidence; a further portion is taken up in the evaporation of water or in the melting of snow and ice. The remainder of the incoming beam is absorbed, and tends to raise the temperature of the surface. Because of the almost complete transparency of the air to solar radiation, absorption takes place mainly at the earth's surface with the result, of great importance for meteorology, that the atmosphere is heated not directly from the sun but indirectly from the heated earth; it is heated from below and not from above.

12. TEMPERATURE AT AND NEAR THE EARTH'S SURFACE

In meteorology, the air temperature at the earth's surface is defined as the shade temperature measured in a louvered screen at a height of about 4 feet (1·25 metres) above ground. While the variation of air temperature on the whole follows closely that of the surface itself, there are on occasions large differences between the two. The primary influences controlling the variations of ground and air temperature are the incoming solar and outgoing terrestrial radiation, the nature of the surface, and the horizontal transference of heat by wind.

Insolation

Solar radiation has its greatest heating effect when the sun is highest in the sky; first, because it then passes through the atmosphere by the shortest path and suffers least loss by absorption and diffuse reflection; secondly, because the area of ground exposed to a beam of radiation increases with the obliquity of the path (it varies with the cosecant of the angle of elevation of the sun above the horizon, assuming the ground to be horizontal). There is a third effect, particularly over the sea; a much greater fraction of the radiation is reflected back to space when the sun is low in the sky than when it is high. The broad features of the distribution of average temperature over the earth's surface may be traced directly to the varying elevation of the sun, greatest in equatorial regions and decreasing towards the poles. The seasonal variations, warm in summer and cold in winter, may similarly be explained.

Nature of the surface

Of the solar radiation reaching the surface of the earth, that portion which undergoes reflection plays no part in raising the surface temperature, only the absorbed portion being concerned in this. Apart from the amount of heat lost or gained, the temperature attained by the ground depends upon the character of the surface. When heat is absorbed by a substance the increase of temperature is dependent on the thermal capacity or the amount of heat required to raise the temperature of the substance by 1 degC. By taking the thermal capacity of water as unity we may express that of any other substance as the ratio of its thermal capacity to that of water; this ratio is known as the specific heat. From this it follows that, other things being equal, the rise in temperature is inversely proportional to the specific heat. Water having a specific heat unity experiences relatively small temperature changes, but the solid materials of the earth's surface have a smaller specific heat and temperature changes are therefore greater.

A further factor is the amount of material which shares the loss or gain of heat. Over the solid earth it is the immediate surface which absorbs and radiates heat, and the change of temperature is spread downwards by the slow process of conduction. These effects will differ with various types of ground, but generally speaking changes of temperature during 24 hours affect only the first few inches, while even the seasonal changes affect only a few feet. Conditions over the sea are however very different, for some of the incoming radiation penetrates to a depth of several yards before being entirely absorbed so that the whole of this layer takes part in the temperature changes. Mixing of the surface waters of the sea also tends to spread any temperature changes through a considerable depth. The amount of heat available for raising the sea temperature is further considerably reduced by the transformation of some of the energy of radiation into latent heat of evaporation; this effect is also present over damp ground and vegetation, but it is almost negligible over deserts.

Temperature of surfaces exposed to direct sunshine

The solid ground or other objects exposed to strong sunshine may attain remarkably high temperatures during the middle part of the day if conditions are such that the absorbed heat is conducted but slowly downwards. For example, the surface of a runway or steel rail under the influence of sunshine becomes much hotter than the air, whose temperature is measured in the conventional manner. As a rough rule, the maximum temperature of the exposed surface may be some 20 degC above the maximum shade temperature of the air at 4 feet above ground, but wide variations from this figure may be expected according to the nature and thermal insulation of the exposed surface. Similarly the surface of an aircraft standing on a runway may reach equally high temperatures. However, the take-off performance of an aircraft is related to the air temperature at the level of the engines, and this temperature is given with reasonable accuracy by that measured in the thermometer screen since only an extremely shallow layer of air is affected by the abnormally high temperature of the ground.

Nocturnal radiation and water vapour

The loss of heat by long-wave radiation emitted from the earth's surface depends almost entirely on the temperature of the surface, but this loss is in part offset by a downward stream of long-wave radiation from the atmosphere. The difference between these two streams is the net outgoing long-wave radiation; this can be readily measured at night when the short-wave radiation is absent, and so has come to be known as the 'nocturnal' radiation although it actually takes place continuously and is usually greater by day when the surface temperature is higher. Dry air is practically transparent to the long waves but they are partially absorbed by water vapour, and this in turn emits long-wave radiation in all directions so that some radiation is returned to the ground in this way. Water vapour in the atmosphere therefore obstructs the free passage of radiation from ground to space and so hinders the nocturnal cooling of the surface. However, the vapour is transparent to certain wavelengths of the earth's radiation and the observed nocturnal cooling results mainly from this part of the radiation which escapes through the atmosphere, whatever the humidity, provided that the sky is clear. The lowest temperature on a radiation night is accordingly found at the ground itself while the screen temperature is commonly higher by amounts up to about 5 degC.

Diurnal variation of surface temperature

The differences already mentioned in the physical conditions of the surface have the effect of making diurnal temperature changes much smaller over the sea than over land. In fact sea surface temperature shows a variation from day to night of less than 1 degC, and the air temperature near the surface is equally steady in quiet conditions. On land, however, the diurnal variation of air temperature may average as much as 20 degC in the interior of continents where the moderating effect of the sea does not penetrate. Near the coast the diurnal variation depends on the direction of the wind; with a wind off the land the temperature changes will be almost as large near the coast as inland, but with a wind off the sea of any strength they become small. Even when there is no general wind off the sea the local sea-breezes, which develop regularly on warm sunny days, have a pronounced effect in tempering the heat of the day. In the absence of disturbing influences, surface temperatures are largely controlled by the difference between incoming and outgoing radiation. At night, cooling by long-wave radiation continues unchecked until the arrival of the refracted rays from the sun at dawn, when temperature is at a minimum; thereafter the incident radiation continues in excess of the outgoing until two or three hours after midday, when balance is again reached and temperature is at its maximum. Cooling of the ground and air near the surface is then resumed and proceeds throughout the rest of the day and the following night.

Other things being equal, the diurnal variation is greatest when the wind is calm. With more wind, the surface air becomes mixed with the air above so that the gain of heat by day and the loss by night become spread through a layer of air which may be some 2000 feet thick. In consequence the range of surface temperature is reduced and may become quite small when the wind is strong.

Effect of clouds on surface temperature

When a cloud obscures the sun the decrease in the amount of heat received can be sensed immediately; a small proportion of the radiation is absorbed by the cloud and some diffuse radiation transmitted, but the major part is merely reflected back to space and is lost to the earth. The intensity of the reflection is apparent from the the brightness of a cloud surface exposed to direct sunshine. A thin sheet of cirro-stratus cloud cuts off a fair proportion of the sun's radiation, and a thick layer of lower cloud may be almost opaque. The heating of the earth by day is therefore considerably decreased by a layer of cloud or by what is essentially the same thing, a deep fog. The cooling of the earth by long-wave radiation is even more considerably decreased, sometimes to zero, for a cloud layer readily absorbs all the earth's radiation while radiation is emitted by the lower surface of the cloud towards the earth.

The invisible water vapour, we have seen, absorbs and reradiates some of the long waves but is transparent to others, so the cooling of the earth by radiation is not thereby entirely prevented; on the other hand a layer of cloud, by absorbing and reradiating all the long waves, insulates the ground almost perfectly from above. A cloud radiates like a black body, and if the cloud layer is as warm as the earth it will send as much heat downwards as the earth radiates upwards with, consequently, no net loss from the ground. The diurnal variation of temperature at the ground becomes small when skies are obscured with low cloud; the lower the cloud the more nearly its temperature approaches that of the ground and the more effective it is in reducing nocturnal cooling.

If, as frequently happens, a clear sky at night clouds over after radiation has caused the air and soil at the surface to become cooler than the soil below, then a flux of heat upwards from the ground to the air produces a rise in temperature at the grass and screen levels.

Local differences in surface temperature

As a result of the varying character of the earth's surface and the topographical features, there are local differences in temperature which are of interest in aviation on account of the upward convection currents set up over relatively warm ground during the day and the possibility of mist or fog developing as the temperature falls at night. Under the influence of the sun's radiation, bare rock, dry soil, metalled roadways and runways take up a relatively high temperature compared with grass-covered land and wooded areas. The cause is to be found mainly in the loss of heat by evaporation of moisture from the vegetation. Water surfaces are particularly slow to heat for reasons previously explained.

The differences which occur at night-time when the ground cools by radiation are largely a matter of exposure. In a sheltered position, particularly in a valley, the air becomes much colder than that over freely exposed ground mainly because the air is left undisturbed and is allowed to cool continuously. Being then relatively heavy it remains as a pool of cold air in the valley, unaffected by light winds above. Even slight depressions in the ground may allow cold air to collect in this way. The temperature differences are sometimes astounding, as is well known to the fruit grower concerned with the occurrence of night frosts.

The general effect of height should be allowed for in hilly country, the air above being on the whole colder by about 1 degC for each 500 feet, although there are marked variations and the low land is, as we have seen, often the colder on a clear night.

Effect of the source of the air

The temperature of the air is not only dependent on local conditions but varies widely according to the source of the air supply; if the wind blows from a warm or a cold region it retains its relatively high or low temperature for some considerable time. Generally speaking air moving from high latitudes is felt as cold, from low latitudes as warm, although there are marked variations due to the distribution of land and sea and to the seasonal variation of temperature. Thus, over the British Isles a south-easterly wind in summer tends to be hot but in winter it may be very cold, coming from the interior of the cold continent; similarly westerly winds are usually mild in winter but comparatively cool in summer. These broad effects are just as important as those produced by radiation but their fuller discussion must be left to later chapters.

13. TRANSFER OF HEAT BY LONG-WAVE RADIATION

The vertical distribution of temperature in the atmosphere is the result of several processes, one of which is the transfer of heat by radiation (see Fig. 7). In discussing the cooling of the earth it was noted that perfectly dry air is almost transparent to the long waves but that water vapour readily absorbs radiation within certain wavelengths. The spectrum of terrestrial radiation may roughly be divided into two regions:

(i) That in which water vapour is transparent to radiation; this part of the radiation is transmitted directly through a cloudless atmosphere without in any way affecting the air temperature.

(ii) That in which radiation is absorbed by water vapour.

The radiant heat so absorbed tends to raise the temperature of the vapour and of the air; at the same time the vapour itself emits radiation in all directions, some being directed downwards towards the earth and some being passed upwards. Apart from lateral radiation (which tends to balance out) the emission from water vapour resolves into two main streams, one returning radiation back towards the earth, the other passing radiation upwards; the upward stream is augmented by the outgoing terrestrial radiation in the transparent bands of wavelength, and together

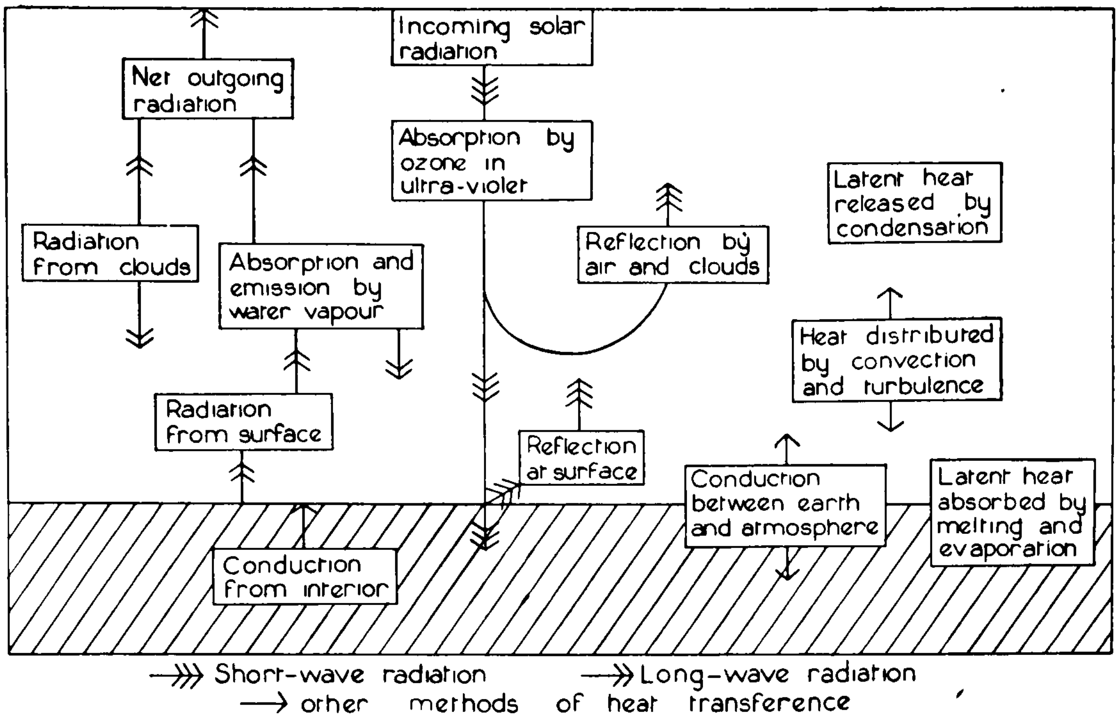


FIG. 7. *Heat processes in the atmosphere*

they provide for the radiative transmission of heat upwards through successive layers of the atmosphere and ultimately out to space.

At any height therefore the air tends to gain or lose heat in proportion to the net radiation absorbed or emitted by the water vapour. The consequential change in temperature from this cause is however rather small, being estimated at about 1 degC in 24 hours. The direction of the net flux of heat is necessarily from high to low temperature, i.e. upwards in the normal state of the atmosphere. Further, the greater the concentration of vapour the less easily is the heat transferred. If radiation were the sole agent concerned in the transfer of heat, the vertical distribution of temperature would adjust itself to the distribution of vapour in such a way that the total outward flux of radiation would on the whole just balance the heat which the earth and atmosphere receive from the sun. The rate of fall of temperature with height in the troposphere that would result from a radiative balance is however much greater than that which actually exists, and consequently the temperature in this region is not controlled by radiation alone.

Other methods of heat transference

Besides radiation, there are several other processes of heat transference in the atmosphere (see Fig. 7 and also Section 28). Since these are of comparable importance with radiation, it is desirable that they should be mentioned here although their fuller discussion cannot be given until later. Thus vertical currents of thermal origin transfer heat upwards or downwards in the process known as convection, while the transfer of heat horizontally by wind is referred to as advection. The irregular eddying motion of the atmosphere, or turbulence, may effect a redistribution of heat in any direction but is of most importance in the vertical direction. When two bodies of different temperature are in contact, heat passes from the hotter to the cooler body by conduction; this process is of little consequence in meteorology except for the flow of heat within the earth's surface and for the exhange of heat between the surface and air actually in contact with it. Lastly, the latent heat which is absorbed by the melting of ice or evaporation of water, mostly at the earth's surface, may subsequently be released elsewhere in the atmosphere by the processes of condensation or freezing. In the troposphere, all the above processes of heat transference are effective together with long-wave radiation. In the stratosphere neither convection nor latent heat have any influence but short-wave radiation becomes important on account of ozone (see Section 3).

14. ADIABATIC PROCESSES

Adiabatic processes in unsaturated air

It is a property of all gases that when they are compressed their temperature rises and when they are allowed to expand their temperature falls. Compression of the gas requires expenditure of energy, but in accordance with the principle of conservation the energy is not lost but is transformed into molecular energy of motion within the gas and so raises its temperature. This property is applied in the diesel engine, the temperature of air in the cylinders being raised by compression to a value above the firing point of the injected fuel oil; another popular illustration is provided from experience with a bicycle pump. Cooling by expansion is perhaps less familiar, but is the basis of a method of obtaining the extremely low temperatures required to liquefy certain gases and it is also applied in some household refrigerators.

When other effects such as conduction and mixing are eliminated so that the air is thermally insulated from its surroundings, then any change of temperature is said to be an adiabatic change and its origin is said to be dynamical as opposed to thermal. In that case the change of temperature (T_0 to T in degrees Kelvin) can be calculated for a given change of pressure (p_0 to p) by the following formula (Appendix I):

$$\log T - \log T_0 = 0{\cdot}288\,(\log p - \log p_0). \qquad (4)$$

Large changes in temperature can be produced in this way; for example, if the initial temperature is 20°C and pressure is reduced from 1000 millibars to 900 millibars, the final temperature is lower by 9 degC. The temperature so attained by air when it is compressed (or expanded) adiabatically to a pressure of 1000 millibars is known as the potential temperature (see Appendix I). By virtue of the definition, the potential temperature remains constant during any adiabatic change.

Now if a small mass of air is made to undergo a change of pressure by being moved fairly rapidly upwards or downwards, any transfer of heat by non-adiabatic processes (conduction, turbulent mixing, radiation) will be too slow to be effective except near the boundary of the displaced mass of air; the interior therefore changes temperature adiabatically. By expressing the right-hand side of equation (4) in terms of a difference in height it may be shown that the temperature of the displaced air changes at the rate of 3 degC (5·4 degF) per 1000 feet, which is known as the dry adiabatic lapse rate (see Appendix I). So long as the sample of air remains unsaturated, it is therefore cooled by ascent and warmed by descent at this rate.

The above rule requires modification if the air becomes saturated (i.e. cloudy) during ascent, but understanding of this requires some explanation regarding the change of state between water and vapour.

Evaporation, condensation and latent heat

When water changes from the liquid to the vapour state, a certain quantity of heat must be supplied; to change boiling water into vapour, more than five times as much heat is required as is needed to bring the same amount of ice-cold water to the boil. Once boiling has begun, the temperature remains constant and the heat supplied in this stage is said to become latent; it is stored up in the vapour and released only when the vapour recondenses. Boiling does not of course take place in meteorological processes but water may pass directly to the vapour state by evaporation, and the amount of latent heat to be provided is only to a small extent dependent on the temperature at which the change takes place. Cooling by evaporation is too familiar to call for illustration; it is made use of in the measurement of humidity by the dry-bulb and wet-bulb thermometers (Section 82). The importance of latent heat in the thermal processes of the atmosphere is also brought out by noting that roughly one-third of the radiant heat absorbed at the surface of sea and land is taken up in the evaporation of water and that most of this heat is subsequently released at other levels in the atmosphere. Given a quantity of dry air at a certain temperature there is a definite limit to the amount of water vapour which it can be made to hold. When the air contains just this amount it is said to be saturated, and the amount of vapour present is the saturation water-vapour content. The saturation content increases with the temperature, in other words saturated warm air holds more vapour than does saturated cold air. If therefore air containing moisture, but not saturated, is cooled it will ultimately reach a temperature at which it becomes saturated. Any cooling beyond this point – known as the dew-point – causes condensation of the surplus moisture which may then be deposited as dew or hoar-frost if the air is at the surface, or held in suspension as the water droplets or ice particles of cloud or fog.

Adiabatic processes in saturated air

When saturated air is made to rise, the adiabatic contribution to cooling continues to operate as with dry air, but the cooling now causes condensation and is in part offset by the latent heat which is liberated. The rate of cooling of ascending saturated air, termed the saturated (or wet) adiabatic lapse rate, is therefore less than the dry adiabatic lapse rate. Unlike the latter, its value is not constant but depends on the amount of vapour condensed. If the temperature is low, saturated air can hold but little moisture and the latent heat released is small; with warm saturated air, much more heat is set free in this way. As a rough value the saturated adiabatic lapse rate at low levels in temperate latitudes may be taken as half the dry adiabatic, or 1·5 degC (2·7 degF) per 1000 feet, but the figure may fall to about 1 degC per 1000 feet in the warm saturated air of tropical regions and may exceed 2 degC per 1000 feet when the temperature falls much below freezing-point, ultimately approaching the dry adiabatic value.

The result is very different when saturated air is made to descend provided little or no condensed water is held in suspension, for adiabatic warming then causes the air to become unsaturated almost immediately and thereafter it warms at the dry adiabatic lapse rate. On the other hand, if all the water condensed out on ascent were retained, the effect of descent would be gradually to evaporate the droplets so that the vapour would remain saturated and the temperature would increase at the saturated adiabatic lapse rate. In practice the larger droplets, if any are formed, fall out as precipitation, while the remaining cloud particles are rapidly evaporated by warming, so that descending cloudy air is generally taken to clear rapidly and to warm at the dry adiabatic lapse rate.

The further discussion of upper air temperature and lapse rates is greatly facilitated by the use of special diagrams, the principles of which will now be introduced.

15. THE TEPHIGRAM

When observations of temperature and humidity are obtained at a series of heights above a point on the earth, it is convenient to plot the data on a diagram so that the vertical distribution can be appreciated at a glance, while certain other results may quickly be obtained. A Cartesian diagram of temperature against height would go some way to meet this purpose. Such a diagram (Fig. 8) may be regarded as a

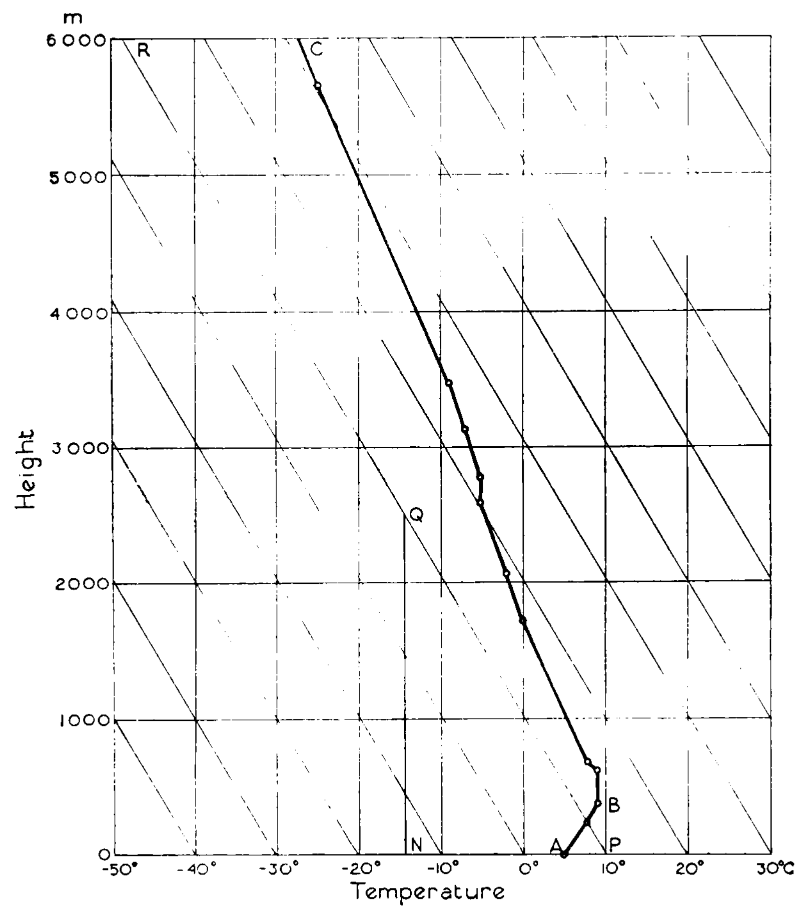


FIG. 8. *Temperature-height diagram showing dry adiabatics and a specimen ascent curve*

rectangular grid consisting of two sets of lines: horizontal lines on each of which the height is constant, and vertical lines on each of which the temperature is constant; the latter may be called isotherms. With a framework of this kind, the variations of lapse rate of a temperature sounding, such as that illustrated, become clear. However, such matters are facilitated by the incorporation of further sets of lines into the diagram. One such set is provided by the 'dry adiabatics'. We have seen how the dry adiabatic lapse rate has the constant value 3 degC (5·4 degF) per 1000 feet or about 1 degC (actually 0·98) per 100 metres. Let us now consider a sample of dry air represented by the point P (10°C at 0 metres). If the sample ascends adiabatically, the point which represents its temperature and height moves along the straight line PR which slopes upwards at about 1 degC for each 100 metres of height. For example, if the air is lifted to 2500 metres its temperature at that height is given by the ordinate QN through the point Q where PR intersects the 2500-metre height line, i.e. —15°C. Descent of air is treated in a similar manner. Thus the result of any adiabatic displacement of dry air may be obtained with the aid of the dry adiabatic line through the point representing the initial temperature and height of the air. A series of such lines, the dry adiabatics, are shown at convenient intervals in Fig. 8. The environment curve ABC shows an increase of temperature, i.e. an inversion, from the surface at A to 400 metres at B and a general fall of temperature with height up to C, apart from two shallow isothermal layers around 500 and 2700 metres.

The temperature-height diagram is not well suited to meteorological observations since temperatures are usually recorded not at fixed heights but at fixed pressures. Consequently it is more convenient to take the pressure as one of the co-ordinates, or better the logarithm of pressure since this is closely related to the height (see Section 7, equation (2)). There are in fact many types of diagram which can be used for the analysis of upper air soundings but the only one that need be described here in detail is that known as the 'tephigram'. This is in general use throughout the British meteorological service; the 1963 edition is reproduced as Appendix II, while part of it is shown in a simplified form in Fig. 9. The diagram has as rectangular axes the temperature and a function known as entropy (see Appendix I) but for

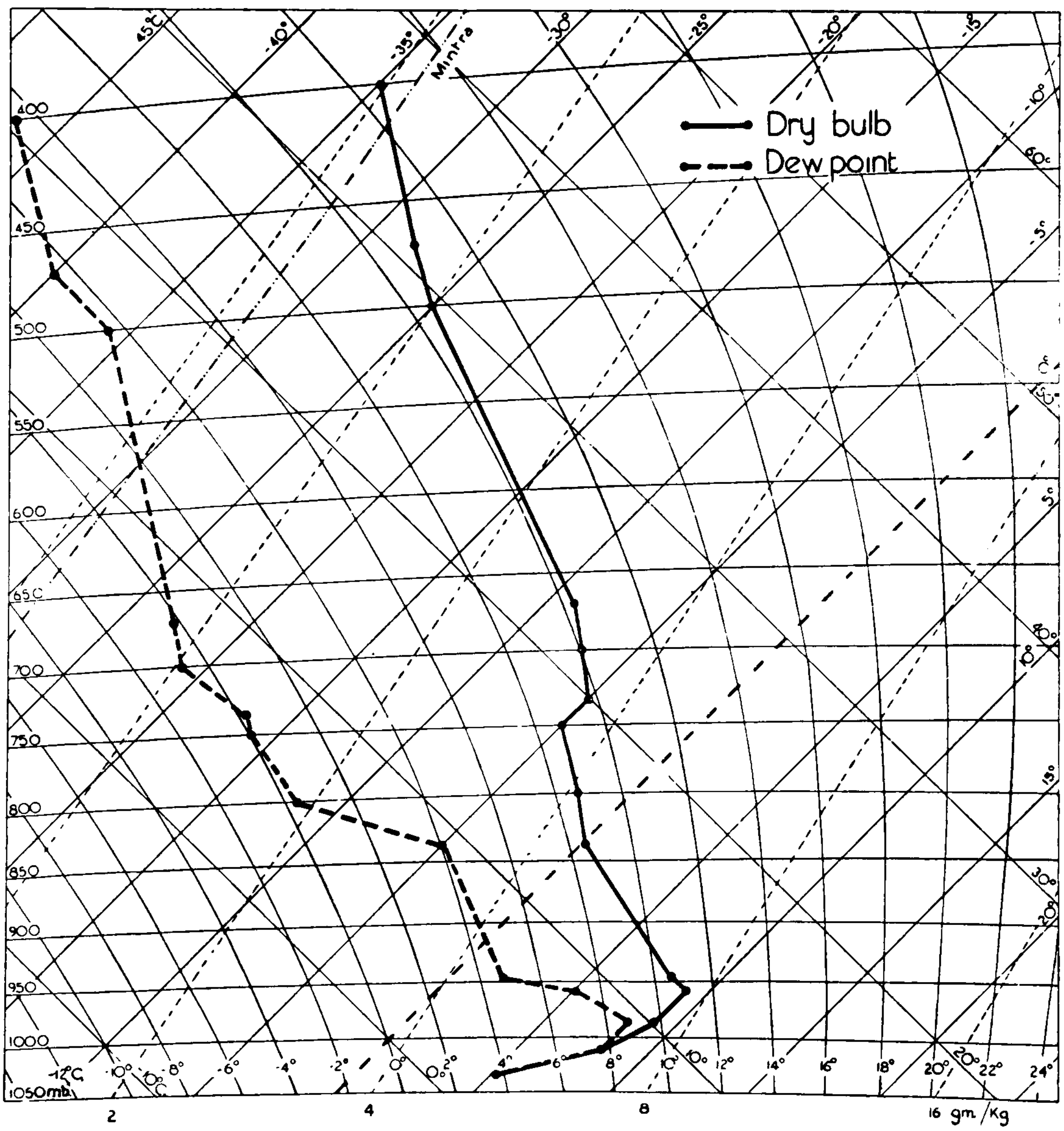


FIG. 9. *Section of tephigram, simplified and reduced, and showing examples of environmental curves*

convenience in plotting the results of upper air soundings the axes are rotated anticlockwise through 45 degrees so that the isotherms slope upwards from left to right and the lines of equal entropy downwards from left to right. Now the entropy of dry air may be expressed in terms of potential temperature and since the latter remains constant during an adiabatic change it follows that the lines of equal entropy are the dry adiabatics; they are labelled with values of the potential temperature at intervals of ten degrees Celsius. Thus the framework of the diagram is the rectangular lattice formed by the isotherms and the dry adiabatics. The lines running almost horizontally across the diagram are the isobars. The faint, pecked lines sloping from left to right are lines of equal humidity mixing ratio (see Section 31). The remaining set of full lines are the saturated adiabatics, curved lines which for the most part slope upwards from right to left; these indicate the relationship between the pressure and temperature of a saturated mass of air as it ascends under conditions which may be described as pseudo-adiabatic since latent heat of condensation is released during the ascent; moreover since the condensed water is presumed to fall out as it forms, then on descent the air would immediately become unsaturated and its representative point would move along a dry adiabatic; for this reason the saturated adiabatics are said to be irreversible.

The tephigram derives its name from the use of temperature (T) and entropy (ϕ, or 'phi') as basic co-ordinates. In ordinary use it may be regarded simply as a distorted pressure-temperature diagram in which the axes are inclined at about 45 degrees. The complete diagram should be carefully studied. Although at first sight it may seem complicated, it is easy to use once familiarity with it has been acquired. It will be found to have several important applications which will be described in due course, including that to the evaluation of height which now follows.

Height and the tephigram

An equation in Section 7 (equation (2)) relates a difference of height to the corresponding difference in pressure when the intervening temperature distribution is isothermal. Results computed from this equation are given on the tephigram in metres for pressure differences of 100 millibars starting at 1000 millibars and entered along the intervening 50-millibar isobar; for example the layer from 800–700 millibars with an isothermal temperature —11·7°C (261·5°K) has a thickness, by interpolation, of 1023 metres. Whatever the actual distribution of temperature, the temperature of an isothermal layer of identical thickness may be obtained by application of the equal-area rule, which is illustrated in Fig. 10. Here if ABC represents the observed distribution of temperature between two pressure levels DA, CE, the equivalent isothermal temperature is given by that isotherm DBE for which the

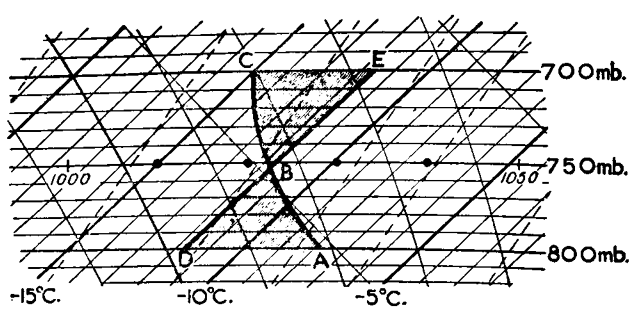


FIG. 10. *Height and the tephigram*

The isothermal DBE is drawn so that the areas ABD and CBE are equal, whence the thickness of the layer AC is, by interpolation, 1023 metres.

areas ABD and CBE are equal; for a layer extending between consecutive 100-millibar isobars, the thickness is then interpolated for this temperature from the height figures on the diagram. This method of determining height from observations of temperature and pressure is theoretically exact for dry air; if the example in Section 7 is reworked in this way, it will be seen that the two results agree to within a few feet.

When the limiting pressure levels are not multiples of 100 millibars the height interval may usually be determined with sufficient accuracy by interpolation or extrapolation of the nearest 100-millibar layer, or by use of equation (1), Section 7, with appropriate values for the temperature and pressure. Heights in feet and metres according to the international standard atmosphere are entered on the left of the diagram in Appendix II for pressures in steps of 50 millibars, zero height corresponding to 1013·2 millibars.

16. STABILITY AND INSTABILITY

The equilibrium of a fluid at rest is determined in theory by giving an element of the fluid a small displacement and noting whether the forces then acting on the element tend to return it to its original position or to increase its displacement. In the former case the fluid is said to be in a state of stable equilibrium and in the latter case in an unstable state – any small disturbance then continues to grow and the original state breaks down, perhaps violently. Similar remarks apply if the fluid is in motion, the supposed displacement being then relative to the original motion. Since vertical displacements in the atmosphere usually take place adiabatically, the vertical stability can be readily considered with the aid of a diagram such as the tephigram. Consider then a sample of air represented by the point P on a tephigram (Fig. 11) and suppose it is lifted adiabatically until its pressure falls to that of the isobar AB. If the air is unsaturated, the point representing the sample as

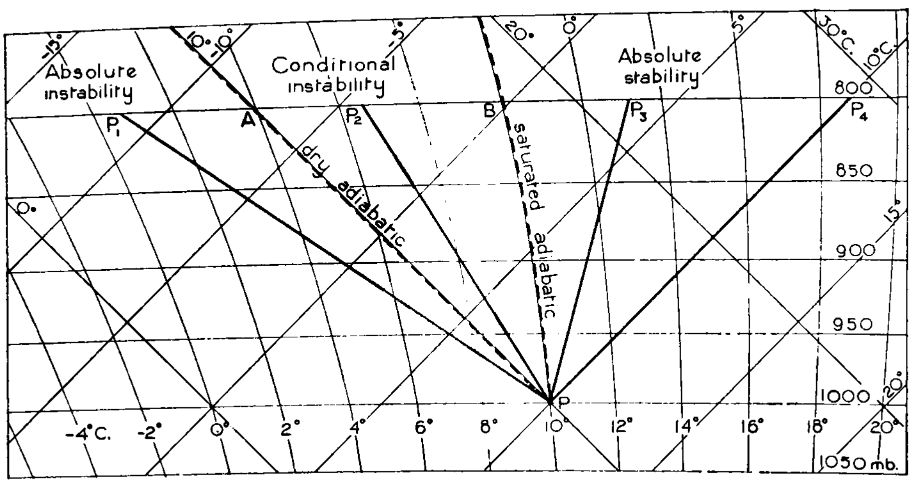


FIG. 11. *Stability and instability in relation to lapse rate of temperature*

it ascends will start to move along the dry adiabatic PA; if the air at P is saturated, its path will lie along the saturated adiabatic PB. We need to consider the supposed adiabatic displacement of the sample in relation to the observed or environmental lapse rate. Three cases can be distinguished.

Absolute instability—observed lapse rate exceeding the dry adiabatic, as PP_1. If the air is unsaturated the temperature of the displaced element at A is higher than that of the environment at P_1 at the same pressure level. The displaced element is therefore lighter, i.e. of lower density than its environment, and its buoyancy tends to increase the displacement still further. Similarly, if the air at P is saturated, the displaced sample at B is again warmer than the environment at P_1 and therefore buoyant. Hence if the lapse rate exceeds the dry adiabatic, any small upward displacement will lead to further ascent because of the buoyancy acquired by the displaced air. This case is referred to as absolute instability.

Conditional instability—observed lapse rate between the dry and saturated adiabatics, as PP_2. If the air at P is unsaturated, then when moved adiabatically to A it becomes colder and therefore denser than its environment P_2; it consequently tends to sink back to P and the lapse rate is stable. On the other hand, saturated air at P would rise to the point B where it would be warmer than the environment P_2 and therefore unstable. The lapse rate PP_2 is therefore stable if the air is unsaturated, but unstable if it is saturated. This case is referred to as conditional instability.

Absolute stability—observed lapse rate less than the saturated adiabatic, as PP_3. The displaced air at either A or B is then cooler than the environment at P_3 and tends to fall back to its original level. This is referred to as absolute stability. The degree of stability increases as the point P_3 lies further to the right from B, so that with an isothermal lapse rate PP_4, or still more with an inversion, the stability is very marked and any vertical displacements are short lived. Any lapse rate less steep than the saturated adiabatic may be described as a stable lapse rate.

Neutral equilibrium

If the observed lapse rate coincides with the dry adiabatic when the air is unsaturated, or with the saturated adiabatic when the air is saturated, then an element displaced upwards remains in its new position and the equilibrium is said to be indifferent or neutral.

Superadiabatic lapse rate

From the preceding considerations it follows that in the free air the lapse rate cannot much exceed the dry adiabatic for then the atmosphere would be unstable and vertical currents would redistribute the heat until a dry adiabatic lapse rate was established. Similarly in saturated or cloudy air, the lapse rate cannot for long exceed the saturated adiabatic. A different situation occurs at the surface when the ground is exposed to strong sunshine (Section 12, page 23). The rate at which heat is then absorbed by the air at the surface may be greater than the rate at which the heat could be conveyed upwards if the lapse rate were dry adiabatic; in consequence a superadiabatic lapse rate, many times the value of the dry adiabatic, is set up but at most this extends only a few feet above the ground.

Some further aspects of instability will be considered later in relation to the convective development of cloud and precipitation (Section 35).

17. VERTICAL DISTRIBUTION OF TEMPERATURE

Convection

With the observed temperature curve AB shown in Fig. 12, consider what happens when the surface air is heated. As the temperature is increased from A to, say, X the air becomes buoyant and begins to rise adiabatically. When it reaches the level where the dry adiabatic through X meets the environment curve at X′, its temperature is reduced to that of the environment and so it rises no further but mixes with its surroundings. If the surface temperature is further increased to Y, then the air would rise along the dry adiabatic through Y to the level of Y′. With continued heating of the surface, as at the ground on a sunny morning, successive small masses of air rise upwards in this way, their place at the surface being taken by air which has descended and is in turn heated by contact with the ground. The rising masses of air convey heat upwards from the surface, an example of the process known as convection. It is clear that the height, or rather the pressure level, to which the heated parcels of air can rise depends both on the rise of temperature at the surface and on the environmental lapse rate. When the lapse rate is stable, the surface air cannot rise far before being cooled to the temperature of the environment; with a steeper lapse rate or with intensified surface heating, the ascent may proceed far enough to produce saturation and cloud formation; these cases will be discussed in Chapter 6. It should be emphasized also that convection of unsaturated air tends to produce a dry adiabatic lapse rate, so far as the air remains unsaturated; thus in Fig. 12, by the time the surface temperature has increased to Y the original stable lapse rate or inversion AY′ has been changed into the dry adiabatic YY′, the part of the curve above Y′ remaining so far unaffected.

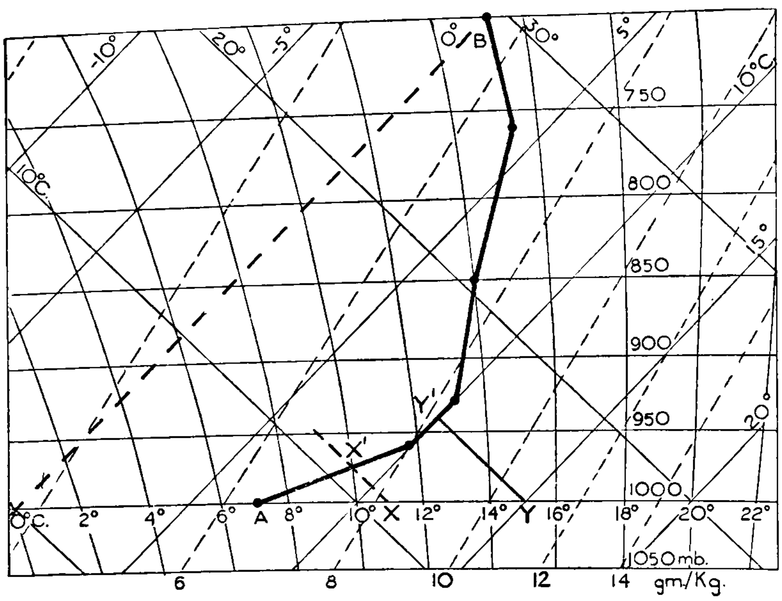


FIG. 12. *Surface heating and convection*

Diurnal variation of lapse rate in the lowest layers

The effect of sunshine in producing convection and a steep lapse rate near the ground has just been noted. On hot afternoons the lapse rate may become dry adiabatic up to a height of a few thousand feet, or up to the base of cloud, where it is replaced by the saturated adiabatic. After the surface temperature has passed its maximum the effect of radiative cooling of the ground begins to spread upwards by conduction and vertical mixing, but as the greatest fall of temperature takes place at the ground itself the lapse rate becomes progressively more stable until vertical currents are finally damped out. This stabilizing process is however liable to be interfered with by wind since this produces vertical mixing (turbulence) near the surface and so tends to keep the lapse rate steep (Section 25). In quiet conditions surface cooling is confined to the lowest levels and it commonly proceeds until a marked inversion has formed; on a clear night over land the depth of the surface inversion may exceed 500 feet. Sunshine usually destroys the inversion rapidly during the following morning as described above, although if fog or a cloud layer hinders solar heating the change will be delayed.

It will be clear that the regular change of temperature from day to night which is characteristic of conditions near the surface is to some extent propagated upwards by conduction and mixing. Observations however show that the diurnal variation of temperature decreases rapidly as height increases and that it ceases to be of practical significance above about 3000 feet.

Temperature and lapse rates in the upper air

The average distribution of temperature in the vertical was described in Section 3 and Fig. 1. This is now amplified by Fig. 13 which shows the variation with latitude in the troposphere and stratosphere. The representation is considerably smoothed as the mean temperature is far from uniform along the parallels of latitude. From the surface upwards the temperature decreases at a rate of about 2 degC per 1000 feet, largely independent of height or latitude; this is known as the average or

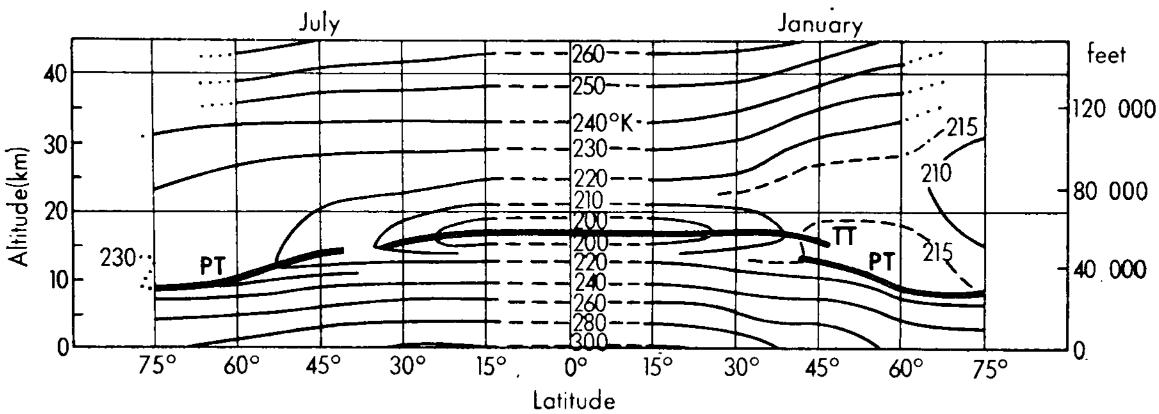


FIG. 13. *Temperature–altitude cross-section for January and July*

PT Polar tropopause TT Tropical tropopause

normal lapse rate. The decrease continues regularly throughout the troposphere but is succeeded in the stratosphere by a layer with practically uniform temperature except in tropical latitudes where the lapse rate quickly reverses sign. The height of the tropopause shows a marked variation with latitude and the part over the tropics is not always continuous with the parts over middle latitudes; sometimes, especially in winter, both may be recognizable at different heights in subtropical latitudes. In such cases of overlapping tropopauses it is safest to assume that the characteristic properties of tropospheric air are to be found only below the lower tropopause, and those of stratospheric air only above the upper tropopause; the intervening air should be regarded as having a transitional character. Because the tropopause is at a greater height over the equator the depth of troposphere in which the temperature decreases with height is greater and thus the temperature at the equatorial tropopause is lower than that at the polar tropopause; and, in fact, the temperatures in the lowest stratosphere above the equator are about as low as those at the same height towards the winter pole. From Fig. 13 it is seen that within the troposphere the temperature increases from the polar regions to the equator. In the stratosphere the temperature increases from the equator towards the pole in summer, but in winter in the lower stratosphere the highest temperature occurs in middle latitudes and from there decreases towards the equator and towards the pole. In the upper stratosphere there is a continuous increase in temperature from the winter polar regions through the equatorial region to the summer polar regions.

The remarkable difference between the average lapse rate in the troposphere and that in the stratosphere arises because the troposphere is warmed mainly by the vertical transfer of heat upwards from the ground by convection, but in the stratosphere radiation to and from its absorbent constituents, such as ozone, provides the most important control in determining the very stable lapse rate (see Section 3) and convection currents are almost absent. The average lapse rate throughout the troposphere is not far removed from the saturated adiabatic (slightly greater at low levels and less at high levels), and as remarked earlier (Section 13) it is very different from that which would result from radiation alone. It is concluded therefore that the average lapse rate is determined mainly by the effects of convection together with the release of latent heat. Convection currents and the associated clouds can rarely penetrate the tropopause where their ascent, if they reach so far, is terminated by the great stability of the stratosphere.

18. TEMPERATURE AND AVIATION

The indirect importance of temperature in aviation will be abundantly clear from the preceding sections since it is in many ways directly related to the development of weather. Apart from the forecasting aspect, the aircraft constructor must allow for all possible variations in temperature up to the ceiling height of the aircraft in connection with aircraft performance and cooling systems. The air temperature again is intimately related to the risk of ice accretion during flight, a problem which will be fully discussed in a later chapter.

Temperature and aircraft performance

The performance of an aircraft depends on several factors among which temperature is important. At a given pressure, high temperature implies low density and so has an adverse effect on both piston-engined and jet-aircraft; this effect is usually greatest when taking off but it should also be considered at other stages of flight, especially for jet aircraft. The efficiency of a jet engine depends in part on the difference between the outside air temperature and the maximum temperature attainable in the combustion chamber, which for safe running must not exceed a certain limit. When the air temperature increases above a certain value, depending on the altitude, both efficiency and true airspeed fall off (other things being equal) and the aircraft's operating height is reduced at the cost of an increase in fuel consumption per air mile.

Temperature accountability

This term refers to the incorporation of the effect of temperature in the performance data of an aircraft. Details of performance in terms of temperature, pressure, altitude, all-up weight, etc. are given in the appropriate aeroplane flight manual; thus, for civil flying, approved operational requirements regulate the maximum weight at which an aircraft shall operate at take-off, *en route* and on landing, for the altitudes and temperature concerned. In this scheme the temperature is to some extent at the choice of the operator; it may be either the air (or 'ambient') temperature at the time (actual or forecast) or a pre-arranged or 'declared' temperature. The declared temperature is intended to give on average a calculated incident risk equivalent to that arising with the use of ambient temperatures. It will vary according to circumstances; for example, for an aeroplane operating to a requirement which allows for failure of one engine at take-off, the declared temperature would be the mean monthly temperature plus a certain factor of the standard deviation* of hourly temperatures from this mean. If the factor were one-half and the mean temperature of a certain month at a given airfield were 19°C and the standard deviation 4 degC, then the declared temperature would be 21°C. A particular advantage of using a declared temperature computed on a monthly basis is that it greatly simplifies the long-term planning of operations; its use enables pay loads to be assessed for an indefinite period ahead, although it may well be the practice of operators to use the ambient temperatures when actually engaged on operations. A declared temperature may be based on the mean monthly temperature and corresponding standard deviation either for the whole 24 hours or for any part of the day; consequently if the mean and standard deviation are given for different hours of the day, the timing of departures and arrivals may be planned to minimize the adverse effects of high temperatures and to take the fullest advantage of low temperatures.

*The standard deviation of a set of observations is the square root of the average of the squares of the individual deviations from the mean.

CHAPTER 4

DENSITY

19. INTRODUCTION

The density of the atmosphere, defined as the mass of air contained in unit volume, is a factor which enters into many of the theoretical problems of meteorology, instances of which will be found from time to time throughout the book. In several ways it is of direct practical concern for the performance of aircraft, because the level of maximum efficiency, the lift, ceiling height and airspeed all depend on it to a marked extent. If density is low at the surface both landing and take-off speeds are greater and the thrust of the engines is reduced; allowance has to be made for these effects in planning the length of runways and in determining the take-off performance and maximum all-up weight of an aircraft. The effects of density on performance in high-altitude flight are discussed in Section 66.

Units

Air density in meteorology is usually stated in grammes (kilogrammes in SI units) per cubic metre. It is however often sufficient to specify it as a percentage of the standard surface density. Alternatively it may be expressed as the density altitude, defined as that altitude in the standard atmosphere to which the actual density corresponds. The engineering unit of density is the slug per cubic foot, where one slug is 32·2 pounds, so that one slug per cubic foot is equal to 0·516 grammes per cubic centimetre.

20. DENSITY OF DRY AND OF MOIST AIR

Density of dry air

The value of the air density – denoted by ρ – is most easily obtained by substituting observed values of the pressure and temperature in the fundamental gas equation in the form

$$\rho = \frac{p}{RT} \text{ (see Appendix I, page 382).} \qquad \ldots . \quad (5)$$

If the density is to be in grammes per cubic metre, then the value of the gas constant R is $2{\cdot}87 \times 10^{-3}$ when pressure is in millibars and temperature in degrees Kelvin. For standard surface conditions, namely 1013·25 millibars and 15°C (288°K), this equation gives the corresponding density as 1225 grammes per cubic metre. The density for any other pressure and temperature is accordingly given by

$$\rho = \frac{1225 \times 288{\cdot}16}{1013{\cdot}25} \times \frac{p}{T},$$

whence

$$\rho = 348{\cdot}4 \frac{p}{T} \text{ g/m}^3. \qquad \ldots . \quad (6)$$

Density of moist air

Equation (6) applies to perfectly dry air. In the atmosphere some water vapour is invariably present; being a gas, water vapour also obeys the fundamental gas equation but the value of the gas constant for water vapour is 8/5 times that for dry air. Now the total pressure p of moist air may be regarded as the sum of the partial pressures which would be exerted by the dry air and by the water vapour if each acted independently of the other. If the partial pressure of the vapour in millibars is denoted by e, then the density (or concentration) of the vapour is given from the gas equation by the expression $5e/8RT$. Further, the partial pressure of the dry air is $(p-e)$ so that the density of the dry air must now be written $(p-e)/RT$. The density of the moist air is the sum of the densities of the dry air and vapour and may therefore be written

$$\text{density of moist air} = \frac{348 \cdot 4}{T}\left(p - \frac{3}{8}e\right). \quad \ldots \quad (7)$$

Thus moist air has a lower density than dry air in similar conditions of pressure and temperature.

If the density of moist air is to differ from the density of dry air by less than one per cent, $3e/8$ must be less than about 10 millibars at surface levels, that is e must be less than 27 millibars, which is the vapour pressure in saturated air at 22°C. Such high humidities occur commonly in moist tropical climates, and in such cases should be taken into account in determining density; in temperate and high latitudes, in dry climates or at high-level aerodromes even in the tropics, the effect of humidity on density is small and can usually be ignored for the requirements of aviation. It may be noted that a decrease of density of about 1 per cent would also be produced by a fall of pressure of 10 millibars, by an increase of temperature of 3 degC, or by an increase in height of 300 feet.

21. VARIATIONS IN SURFACE DENSITY

It is a simple matter therefore to compute the density of dry or of moist air for any given pressure and temperature. At a given pressure, the density is inversely proportional to the absolute temperature so that warm air is comparatively light and cold air heavy. The relative values for dry air at different levels in the international standard atmosphere are given in Table 2, Section 4. In regard to conditions at an aerodrome, the average and minimum air densities occurring in any month may be provided for the use of operators since they indicate the average and the most adverse effects on take-off and landing speeds and on loads. Such information is commonly required in terms of density altitude which can be obtained from a detailed table of the standard atmosphere; there is then available a direct comparison between conditions at the surface of the aerodrome and at the equivalent height in the international standard atmosphere. The average value of the density for a given period may be obtained approximately by inserting average values of pressure and temperature (and, if necessary, vapour pressure) at aerodrome level in equation (6) or (7). The minimum density may occur with exceptionally low pressure, exceptionally high temperature, or with a combination of both. In tropical countries the effect of temperature is usually the more important and it is sufficient to combine maximum temperature with average pressure in the equation. In regions liable to disturbance by deep depressions, low barometric pressure may be the chief cause of low density. Since very low pressure rarely occurs at the same time as high temperature a reasonable estimate is obtained by combining the lowest pressure with the average temperature for the time of year.

For comparative purposes, the variation of density over the globe at sea level may be derived from charts of mean pressure and temperature. The results indicate a mean density in the neighbourhood of 1200 grammes per cubic metre at sea level in equatorial regions, increasing polewards to as much as 1550 grammes per cubic metre in Siberia in winter, where low temperature is combined with high pressure. In the northern summer, high temperature and relatively low pressure over south-western Asia and northern Africa produce a fall to below 1150 grammes per cubic metre. Variations during the course of a day arise mainly from the diurnal variation of temperature, the lowest densities occurring in early afternoon and the highest at night. Seasonal changes in density result from variations in both temperature and pressure. As already noted, the density decreases by about 1 per cent for every 3 degC rise of temperature or 10 millibars fall of pressure.

22. VARIATION OF DENSITY WITH HEIGHT

Reference to the standard atmosphere, Table 2, shows that throughout the troposphere pressure falls off upward much more rapidly than does the temperature, indicating a decrease of density with height at all levels. The decrease at lower levels is given very closely by subtracting 3 per cent of the value for any given level to obtain the value 1000 feet higher; successive applications of this rule give good approximations up to about 20 000 feet. It is also useful to remember that in the standard atmosphere the density has roughly half its surface value at 20 000 feet, a quarter at 40 000 feet, and one-tenth at about 60 000 feet.

In the upper atmosphere the density at any point fluctuates through a proportionate range similar in magnitude to that at the surface. A variation as much as 10 per cent from the average value, a very unlikely extreme, would be equivalent to a change in height of about 3000 feet and apart from a possible temperature effect would alter the ceiling height of an aircraft by a comparable amount. At sea level we have seen that the mean density is lowest near the equator and highest in high latitudes; a similar distribution is at first maintained aloft but as the reduction of pressure with height is more rapid in cold areas than in warm, the latitudinal variation diminishes until by about 26 000 feet the density is almost uniform over the globe in all seasons. At this level, density does not depart by more than 2 per cent from the standard atmosphere regardless of season or location. A second level of minimum departure from standard occurs near 80 000 feet but the amount of departure is more variable than at 26 000 feet. Between these two minima there is a level of maximum departure from standard at about 50 000 feet where density becomes greater over the tropics than over the poles. This reversal implies that while certain aircraft have a comparatively low ceiling in the tropics because the air density in the lower atmosphere is less there than in higher latitudes, the performance relative to that in higher latitudes nevertheless improves with height; at about 26 000 feet there is little difference over the world, while at higher levels and particularly in the lower stratosphere the air has more lift in low latitudes. For record high-altitude flying, conditions should therefore be more suitable in tropical countries than in other latitudes, and, in any other locality, more favourable during the summer season than at other times of the year.

CHAPTER 5

MOTION OF THE ATMOSPHERE

23. INTRODUCTION

The motion of the atmosphere embraces the whole range of phenomena from the major currents of the atmosphere down to the random molecular motions. The various types of motion, vertical as well as horizontal, are all intimately connected with weather in one way or another and most of them are of direct importance to the flight of aircraft. The term 'wind' is of restricted significance, referring to sustained horizontal movement. On the large scale, the wind is closely related to the horizontal variation of pressure and so to the ever changing pattern of cyclonic and anticyclonic pressure systems over the earth. Superimposed on the broad flow of air there may be smaller secondary disturbances which sometimes are too local or too transient to show up on the isobaric chart, while the wind itself consists of a succession of gusts and lulls, the effects of which may be seen on the detailed record of a suitable instrument. Finally, the motion may be broken down to that of the molecules of the various gases of which air is composed. Although this haphazard molecular activity does not require further discussion in this chapter, it nevertheless plays an essential part in many processes. It is by such activity that the pressure of the atmosphere is exerted; it is connected with the property of internal friction or viscosity which causes the energy of the winds to be continually dissipated and transformed into heat energy; it accounts for the 'skin-friction' to which an aerofoil is subject as it moves through the air; it is the means by which heat is exchanged between the earth's surface and the air in contact with it; and it is in part responsible for frictional turbulence. The main concern of this chapter is first with the relation between wind and pressure both at the surface and at higher levels, secondly with the modifications exerted by topography, thirdly with the various types of vertical motion, fourthly with the role of wind in heat transference, and finally with some applications to aviation. The last includes the effect of vertical currents in producing bumpiness, the use of wind and pressure data in navigation and in flight planning, and the effect of wind changes on sextant observations.

24. PRESSURE AND WIND

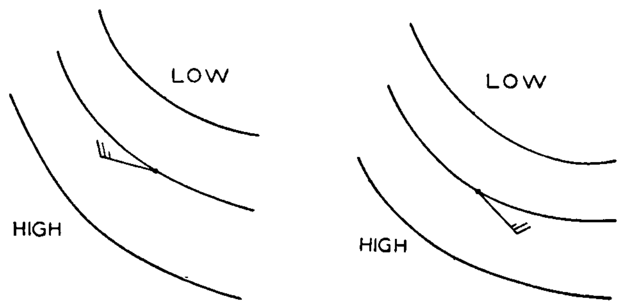
Relationship between isobars and wind, Buys Ballot's law

Perhaps the most striking feature of any synoptic chart is the obvious relationship between the wind direction and the isobars, a relationship which is known as Buys Ballot's law:

> If an observer stands with his back to the wind the lower pressure is on his left in the northern hemisphere, on his right in the southern hemisphere.

The general validity of this law may be confirmed on any of the weather maps given in this handbook although not without reservations. A few cases do not apparently fit the rule at all well, while, as illustrated in Fig. 14, the surface wind in general blows obliquely across the isobars from high pressure to low and in consequence the direction of the lowest pressure is not precisely at right angles to the direction of the surface wind. It is also evident from examination of the charts that strong winds occur when the isobars are near together and light winds when they are far apart.

These relationships do not concern surface winds alone. When upper air observations are studied it becomes apparent that the wind at some 2000 feet above the sea or low ground is directed almost exactly along the surface isobars with the lower pressure to the left (in the northern hemisphere) in accordance with Buys Ballot's law, while its speed is greater than that of the surface wind and inversely proportional to the distance between the isobars.



Northern hemisphere Southern hemisphere

FIG. 14. *Typical surface wind arrows*

This is also true of the wind at any height in the free atmosphere, in relation to the distribution of pressure at that height.

The importance of these rules cannot be exaggerated for they enable the forecaster to give an estimate of wind from isobars alone, and if he can anticipate the changes in the field of pressure he can anticipate also the changes in the wind. The reasons for this close connection between wind and pressure must now be considered.

Primary cause of wind

This may be traced to variations in density due to temperature differences resulting from the effects of solar and terrestrial radiation. The sole external force operating on the air is gravity, but this acts in the vertical direction and can produce horizontal motion only indirectly through the agency of pressure differences. Thus if air has come to rest locally, it cannot begin to move again until, on account of density variations, a pressure difference is set up; the air is then acted on by a force perpendicular to the isobars towards the low-pressure side. The magnitude of this force is given by the pressure gradient G, which may be taken as the difference of pressure between consecutive isobars divided by the distance between them. However, observations show that in the free air a steady wind is directed not across but along the isobars. This characteristic is explained in the succeeding paragraphs.

Effect of the earth's rotation

Once a body has been set in motion it will continue to move indefinitely at constant speed in the same straight line in space so long as there is no resultant force acting upon it. If a body is set in motion over the earth's surface, the action of gravity tends to prevent its leaving the surface; for example, an object set moving horizontally over the earth at not too great a speed might be expected (in the absence of friction) to continue indefinitely along a great circle and eventually to pass through the starting place again. In reality, rotation of the earth causes any great circle (other than the equator) to be continuously changing its orientation in space, and since the moving object tends to maintain a fixed direction in space, it must depart from a great-circle track. It may be shown that any body moving relative to the rotating earth is subject to the so-called 'geostrophic' force; this is also known after its discoverer as the Coriolis force. This force acts at right angles to the relative velocity and, in horizontal motion, is directed towards the right in the northern hemisphere and towards the left in the southern hemisphere; for a unit volume of air, the magnitude of the force is $2\Omega\rho V \sin\phi$ where Ω is the angular velocity of the earth (one revolution a day or $2\pi/24$ radians an hour), ρ the density, V the wind speed and ϕ the latitude. The formula shows that the Coriolis force is greatest at the poles and falls to zero at the equator (see Appendix I).

Geostrophic wind

It has been explained that air moving over the earth's surface is subject to forces due to the pressure gradient and to the earth's rotation. In steady motion along a great circle, these two forces are the only ones acting in a horizontal plane and they must therefore be equal and opposite. Writing the wind speed in this case as V_g and equating the expressions for the two forces

$$2\Omega\rho V_g \sin\phi = G$$

or

$$V_g = \frac{G}{2\Omega\rho \sin\phi} \quad \ldots\ldots (8)$$

The wind V_g defined in this way is termed the geostrophic wind. Since the Coriolis force acts at right angles to the wind direction, the pressure gradient must also be at right angles to the wind, so that the geostrophic wind therefore blows along the isobars with the lower pressure on its left in agreement with Buys Ballot's law. In the southern hemisphere the Coriolis force is reversed and the wind blows with the low pressure to the right.

It follows from the formula that the speed of the geostrophic wind is proportional to the pressure gradient or inversely proportional to the distance between the isobars – the closer the isobars, the stronger the wind. The speed is also inversely proportional to the sine of the latitude. As the equator is approached, $\sin\phi$ becomes small and the velocity would appear to increase indefinitely. This is of course impossible and means merely that the formula breaks down near the equator. In practice its use is limited mainly to extratropical latitudes where it gives in general a useful estimate of wind speed in the free atmosphere. Table 4 shows the speeds in various latitudes which correspond with that pressure gradient which gives a speed of 30 knots in latitude 45°, the density being assumed constant.

TABLE 4. *Geostrophic wind speeds at various latitudes for the same pressure gradient*

Latitude	90°	75°	60°	45°	30°	15°
Speed (kt)	21	22	25	30	42	82

For other pressure gradients the wind speeds are in the same ratio.

From this table it is seen that a pressure gradient which corresponds with only moderate winds in middle latitudes gives speeds of double or more in low latitudes, but appreciably lower speeds near the pole. Allowance should therefore be made for

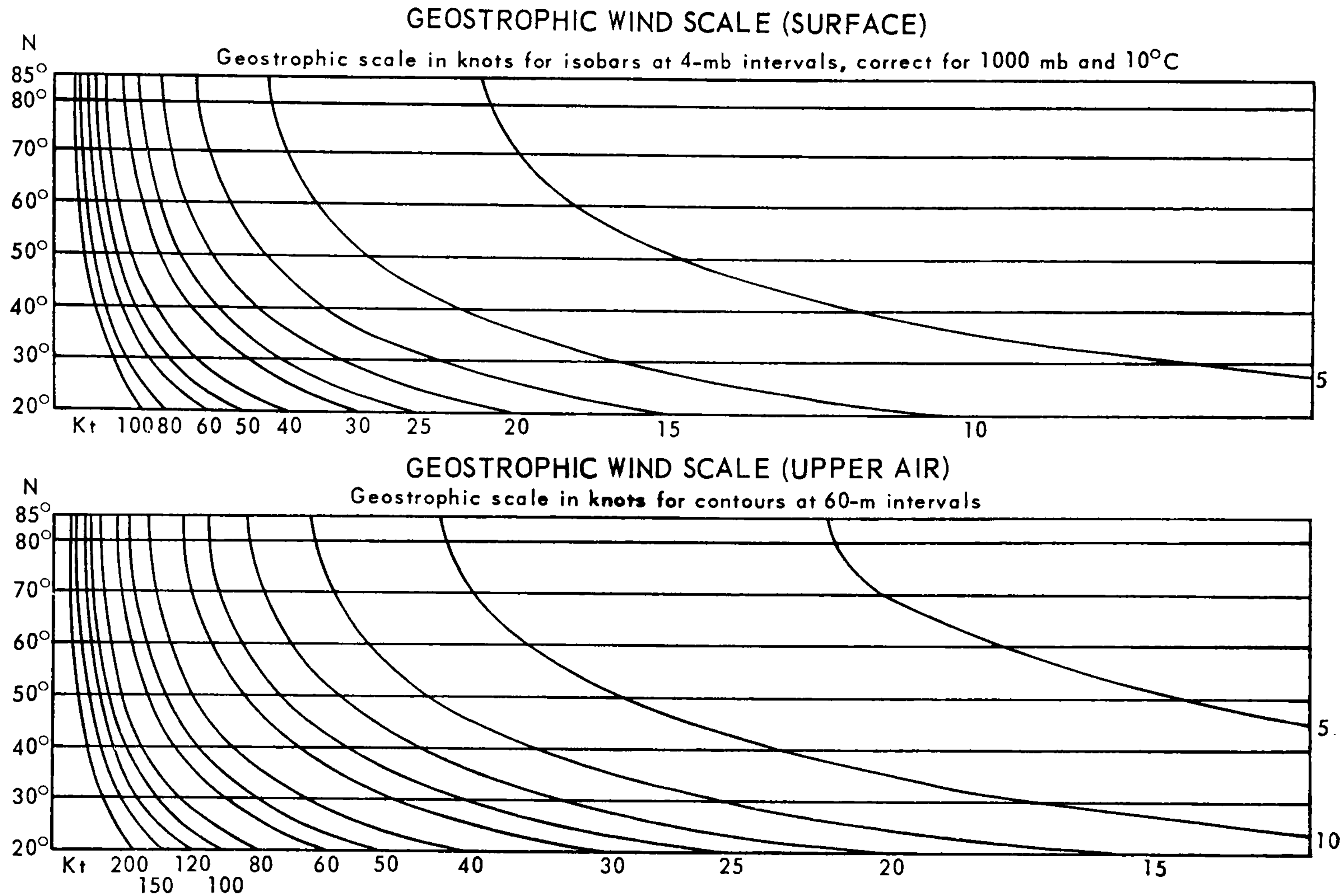


FIG. 15. *Examples of geostrophic wind scales*

These are for use with charts constructed on the polar stereogaphic projection with one standard parallel at 60°N, where the natural scale is 1:15 000 000.

this effect when interpreting weather maps. Pressure gradients generally are much more open in low than in high latitudes, and within the tropics steep gradients and strong winds occur as a rule only in tropical revolving storms.

Limitations of the geostrophic rule

The geostrophic wind is a precise measure of the true wind only when a balance is struck between the pressure gradient and the Coriolis force. This is an exacting condition, and is strictly fulfilled only with straight, parallel isobars. If the isobars are curved or if the pressure distribution is changing with time then additional forces are involved which may make the geostrophic formula inapplicable. Apart from the effect of curvature of the isobars, appreciable departures from the geostrophic rule arise with unsteadiness of the motion or with the introduction of a vertical component. Thus with transitory disturbances such as gusts and squalls the forces operating are unbalanced, so that these irregularities have little relation to the geostrophic wind. Similarly, with local winds such as land- and sea-breezes there is seldom sufficient time for a balance to be reached, so that as a rule they differ in both direction and speed from the wind which would normally be associated with the pressure gradient.

For these reasons there are many local or even general variations from Buys Ballot's law and the geostrophic rule. Nevertheless the geostrophic wind is fundamental in weather forecasting since often enough it gives a satisfactory approximation to the actual wind, while cases of serious departure from it are easily recognized in advance.

Geostrophic scale

The geostrophic wind is measured with the aid of a suitable scale engraved on transparent material and placed across the isobars. Apart from variations in density which can usually be ignored, the geostrophic speed in a given latitude depends only on the distance between the isobars, and the scale is merely a method of converting this distance into units of speed according to equation (8) on page 44. Types of geostrophic scale applicable over a range of latitudes are shown in Fig. 15. The use of the upper air scale is described on page 59. Fig. 16 illustrates the use of a single-latitude geostrophic scale which is placed across the isobars at right angles

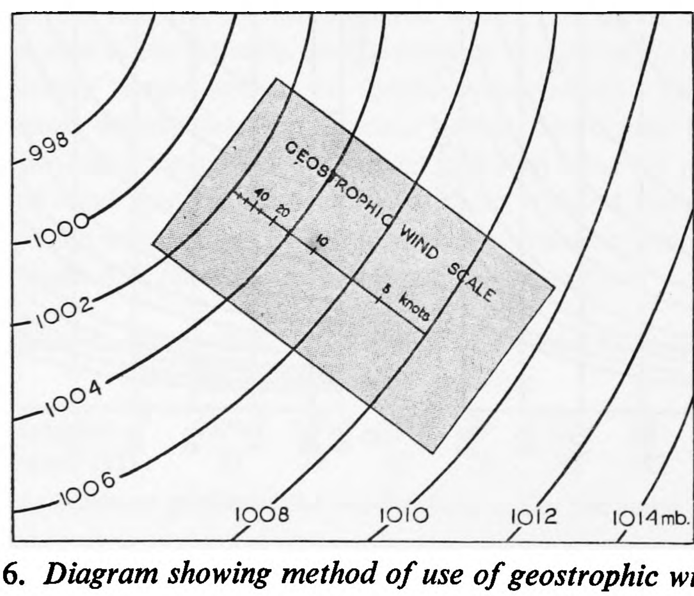


FIG. 16. *Diagram showing method of use of geostrophic wind scale*

In this case the reading is 14½ knots on a chart scale of 1 : 10 000 000 or 29 knots if the scale is 1 : 5 000 000.

with the left-hand extremity on one isobar and the speed is read off at the point where the next isobar crosses the scale. Clearly the spacing of the graduations depends on the scale of the chart in use as well as on the interval of pressure between consecutive isobars and both must be allowed for when using the scale; for example if the scale is constructed for use with 2-millibar intervals on a chart of $1:10^7$, the reading should be doubled when used with 2-millibar intervals on a chart $1 : 5 \times 10^6$.

Component of geostrophic wind in any direction

If, in Fig. 17, PQ is drawn normal to consecutive isobars PP′, QQ′ then the geostrophic wind blows at right angles to PQ with a speed inversely proportional to PQ, say

$$V = \frac{k}{\mathrm{PQ}}.$$

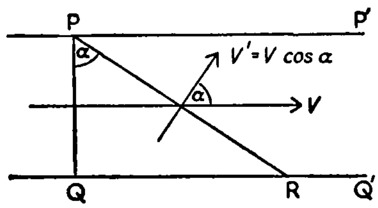


FIG. 17. *Component of geostrophic wind*

If now the geostrophic formula is applied to a line PR intercepted by the same isobars and inclined at an angle α to PQ, the speed obtained is

$$V' = \frac{k}{\mathrm{PR}} = V \times \frac{\mathrm{PQ}}{\mathrm{PR}},$$

or

$$V' = V \cos \alpha ,$$

which is the component of the geostrophic wind at right angles to PR. Hence the component of the geostrophic wind perpendicular to any line is obtained by laying the geostrophic scale along that line and reading off the speed in the usual way.

Cyclostrophic wind

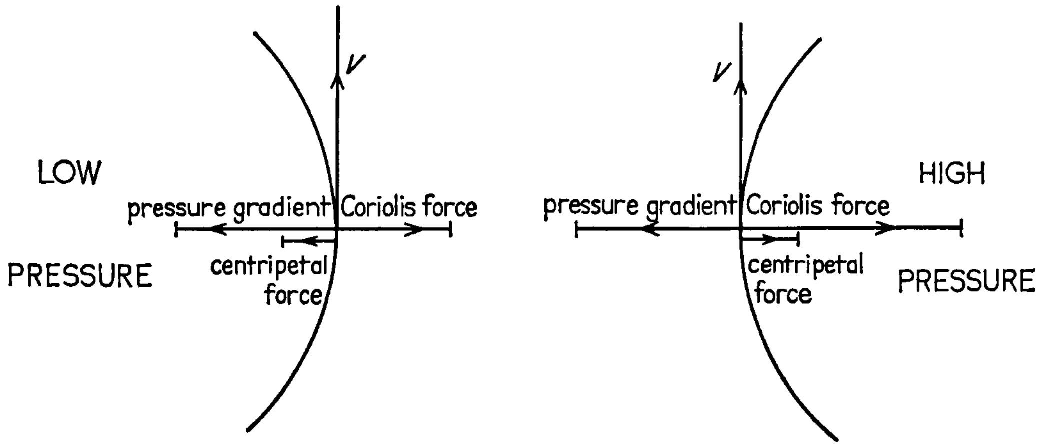
Since air is seldom constrained to follow a great circle, consideration must be given to the effect of the centripetal force which is necessary if air is to move on a path which is curved relative to the earth. If the air is moving steadily on a circular track of radius r and with a horizontal velocity V it has an acceleration to the centre of V^2/r. The centripetal force acting on unit volume of air is $\rho V^2/r$. If the Coriolis force is negligible this must be provided by the pressure gradient. Hence

$$\frac{\rho V^2}{r} = G \text{ and } V = \sqrt{\frac{Gr}{\rho}}. \qquad \ldots . \quad (9)$$

Motion under these circumstances is described as cyclostrophic and equation (9) gives the value of the cyclostrophic wind. In certain types of motion, as for example near the centre of a tropical revolving storm or in a circular tornado, the equation gives a close approximation to the actual wind.

Gradient wind

Generally, when isobars are curved both geostrophic and cyclostrophic effects should be considered. If air is moving steadily round a centre of low pressure on its left in the northern hemisphere (Fig. 18 (*a*)), the centripetal force acting inwards



(*a*) Cyclonic motion (*b*) Anticyclonic motion

FIG. 18. *Gradient wind and balance of forces in the northern hemisphere*

must be provided by the difference between the pressure gradient directed inwards towards low pressure and the Coriolis force directed outwards. Since the latter is at right angles to the wind it follows that all three forces must lie along the radius of curvature. Therefore

$$\frac{\rho V^2}{r} = G - 2\Omega\rho V \sin\phi , \qquad \ldots . (10)$$

and dividing by $2\Omega\rho \sin\phi$

$$V = \frac{G}{2\Omega\rho \sin\phi} - \frac{V^2}{2r\Omega \sin\phi},$$

$$\text{or } V = V_g - \frac{V^2}{2r\Omega \sin\phi} \qquad \ldots . (11)$$

where V_g is the geostrophic wind (equation (8)).

The pressure gradient, which with straight isobars would produce a wind velocity V_g, must in cyclonic motion provide the inward acting centripetal force. Thus in cyclonic motion the actual wind velocity will be less than the geostrophic wind velocity for the same pressure gradient.

In anticyclonic motion the centripetal force, acting towards high pressure this time, must be provided by the difference between the Coriolis force directed inwards towards high pressure and the pressure gradient directed outwards (Fig. 18 (*b*)). Thus in anticyclonic motion the actual wind velocity will be greater than the geostrophic for the same pressure gradient. The expression for the velocity now becomes

$$V = V_g + \frac{V^2}{2r\Omega \sin\phi}. \qquad \ldots . (12)$$

The value of V obtained from equations (11) and (12), which include both cyclostrophic and geostrophic terms, is known as the gradient wind. It depends not only on the pressure gradient but also on the curvature of the isobars and accurately represents the true wind only in steady and uniform circular motion.

The precise determination of the gradient wind is somewhat involved; the correction from geostrophic to gradient wind differs in magnitude as well as sign according as the curvature of the isobars is cyclonic or anticyclonic, while with moving systems it is the curvature of the trajectory of the air that is involved, not the curvature of the isobars. In practice, the geostrophic wind is determined in the usual way, and then, if required, reference to a table or diagram gives the correction to obtain the gradient wind. For example, suppose in a stationary system the geostrophic wind is 40 knots, radius of curvature of isobars 500 nautical miles, latitude 45°; then the gradient wind is 34 knots in cyclonic motion and 58 knots in anticyclonic motion. Again, if the radius of curvature is 1000 nautical miles, the values of the gradient wind are 36 and 46 knots respectively. The percentage correction to the geostrophic wind to obtain the gradient wind increases with the geostrophic wind speed, decreases with increasing radius of curvature, and decreases with increasing latitude. When the isobars are straight and parallel the gradient wind equals the geostrophic wind. When, as in low latitudes, the Coriolis term is negligible (equation (10) with small ϕ) the gradient wind is equal to the cyclostrophic wind.

The geostrophic formula being the easiest to apply is the most widely used and usually gives a useful approximation outside the tropics when the curvature is not too great. In other cases, for example near the centre of a pressure system where the radius of curvature of the isobars is small, it should be remembered that the geostrophic formula overestimates the speed in cyclonic motion and underestimates it in anticyclonic motion. Near the centre of a tropical revolving storm, the radius of curvature is small, as also is the Coriolis effect because of the low latitude; the cyclostrophic formula then gives a reasonable estimate of the wind. This is even more true of the circular tornado (Section 119) where the effects of small radius of curvature, steep pressure gradient and strong wind combine to render the Coriolis term relatively insignificant (equation (10)).

Effect of surface friction

The primary effect of friction with the earth's surface is to reduce the rate of flow in the lowest layers. The thickness of the layer affected – the friction layer – is variable; it depends primarily on wind speed, lapse rate of temperature and roughness of the surface. Since throughout this layer the wind speed increases from the surface upwards, the geostrophic wind as determined from the isobars at mean sea level is considered to apply to the unretarded air just above the friction layer, that is at a height of about 2000–3000 feet above the surface, provided the surface itself is not more than 500 feet or so above sea level. Within the friction layer the wind is slowed down and the Coriolis force, being reduced in proportion, is no longer sufficient to balance the pressure gradient. For this reason the wind near the surface blows somewhat across the isobars towards the side of low pressure. Both the reduction in speed of the surface wind and its inclination to the isobars vary considerably with circumstances; rough rules are that over the sea, where friction is small, the surface wind blows at about 15 degrees to the isobars while its speed is about two-thirds the geostrophic speed; over land, where friction is greater, the inclination to the isobars is about 25 degrees and speed about one-third to one-half the geostrophic value.

25. WIND NEAR THE EARTH'S SURFACE

In this section are considered the main features of the motion of the atmosphere within the layer affected by surface friction, the depth of which is usually between about 1500 and 3000 feet although at times less than 500 feet.

Turbulence and gustiness

The short-period and small-scale oscillations in wind which are classed under the general term turbulence are indicated very clearly on the records from an electrical anemograph (Fig. 19). The speed and direction pens both record a trace of

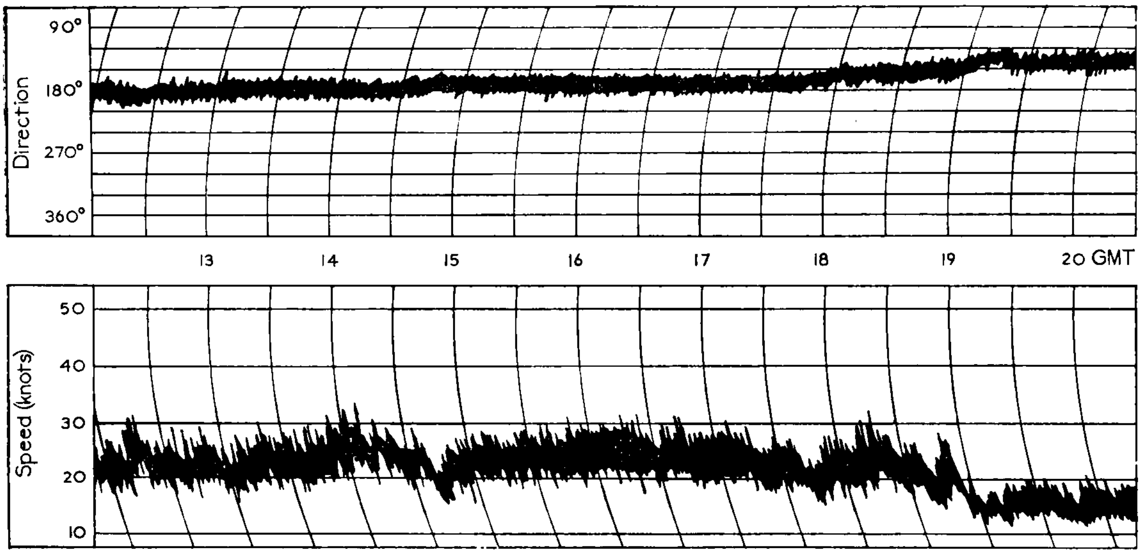


FIG. 19. *Diagrammatic representation of part of the record from the electrical anemograph at Valley on 18 January 1967 (see also Section 140)*

varying width caused by a series of oscillations in rapid succession which follow no obvious regularity but spread as a whole over a reasonably well-defined range. It is usually possible to imagine the position of a mean line through each trace and so to determine the mean wind at any time. The width of the trace is an indication of the degree of gustiness and a gustiness factor may be defined:

$$\text{gustiness factor} = \frac{\text{range of fluctuation in gusts and lulls}}{\text{mean wind}} .$$

The range is determined from the highest gusts and lowest lulls over a period such as an hour, ignoring any very exceptional variations. The factor is conveniently given as a percentage; for example, if the mean wind is 30 knots with gusts up to 45 knots and lulls down to 15 knots the range is 30 knots and the factor is 100 per cent.

Two types of turbulence

There are two distinct types of turbulence or gustiness, distinguished by the terms frictional and thermal. Either type comprises both vertical and horizontal fluctuations of wind, these being practically inseparable. The frictional type of turbulence is a characteristic property of fluid flow in the neighbourhood of a boundary when certain conditions are realized. Fluid actually in contact with a stationary boundary is itself at rest, while throughout the friction layer the mean speed increases with distance from the boundary until the unretarded free stream is reached. When for a given fluid and type of boundary the speed of the free stream is sufficiently low, the flow remains smooth or laminar, but when the free speed surpasses a certain limit the flow becomes unstable and breaks down into turbulent motion; eddies then form near the boundary, drift away into the stream, and cause the friction layer to become much deeper in turbulent than in laminar flow.

The thermal type of turbulence results from convection currents set up by surface heating (Section 17). The heating may result from insolation over the land or from the passage of a relatively cool mass of air over a warmer land or sea surface.

Some factors affecting turbulence

The flow of air over the earth's surface is usually although not invariably turbulent, and the degree of turbulence and the thickness of the friction layer vary from one situation to another. Frictional turbulence is widespread largely because the earth's surface is rough, dynamically speaking; moreover, gustiness is accentuated by flow over buildings, trees or rugged country. Further, since with either type of turbulence the eddies involve vertical as well as horizontal velocities, they develop more easily as the lapse rate becomes steeper. Factors unfavourable to the development of frictional turbulence are flow over open sea or relatively smooth ground, light wind or calm and a stable lapse rate; factors unfavourable to thermal turbulence are surface cooling and stable lapse rates. Over land there is in consequence a diurnal variation in turbulence which is most vigorous by day when the lapse rate is steep and least on a clear night with an inversion of temperature. The difference in degree of turbulence over land and over sea is revealed by the gustiness factor derived from anemometer records; thus, according to certain statistics, the gustiness factor for Kew Observatory (near London) is 100 per cent, while for Falmouth (Cornwall) with winds off the sea it is only 25 per cent. When the character of the surface in the neighbourhood of an aerodrome varies according to direction, as at a coastal site, then the gustiness varies with wind direction, often to a marked extent.

Surface turbulence and aircraft

To the occupants of an aircraft, turbulence is recognized as bumpiness and the difference between flying at low levels over land and over sea in this respect is well known. Strong winds are habitually turbulent but again the degree of turbulence increases with the roughness of the surface and is generally more marked over land than over sea. In turbulent conditions the landing and taking off of aircraft may be difficult, for sudden changes of wind speed or direction can cause loss of control when the aircraft is only just airborne.

Effect of turbulence on lapse rate of temperature

Turbulent mixing has the effect of steepening the lapse rate in the friction layer. If the lapse rate is stable to begin with, then air brought down to the surface from the upper part of the layer is warmed adiabatically to a temperature above that of the surface; similarly air carried upwards is cooled below the temperature of the upper layers. As mixing proceeds, the effect is to warm the lower layers and to cool the upper layers until the lapse rate becomes dry adiabatic, assuming the air remains unsaturated. Thus in Fig. 20, if the lapse rate is initially given by ABC . . ., the result of mixing would be, ideally, to replace the portion AB by the dry adiabatic A′B′ where BB′ corresponds to the top of the mixing layer. The position of A′B′ is such as to leave the mean temperature of the mixing layer unaltered. It is seen that a sharp inversion of temperature is formed at the top of the friction layer, although in practice it is likely to be rounded off as suggested by the broken curve B′C. Often too the wind is not strong enough to produce complete mixing, in which case the final temperature curve will be intermediate between AB and A′B′. The case when the rising air becomes saturated will be discussed under the heading of turbulence cloud, Section 33.

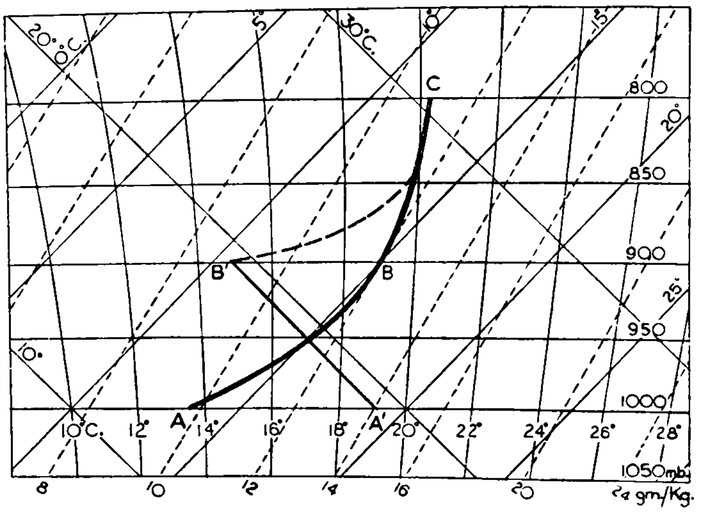


FIG. 20. *Effect of turbulence on lapse rate of dry air*
ABC is the initial lapse rate, A′B′C the lapse rate after mixing.

Thermal eddies

The eddies of thermal origin are often of larger dimensions and productive of stronger gusts than those produced frictionally. Therefore thermal currents are more noticeable as a rule to occupants of an aircraft than are the frictional eddies; they may also extend to considerable heights when the lapse rate is favourable, and they are effective in mixing the surface air with that of the upper layers.

Near the surface extreme gusts occur when some of the faster-moving air above the friction layer is brought down to the surface by the thermal eddies; in consequence the maximum gust speed in these circumstances may even exceed the speed of the geostrophic wind.

Squalls

In certain conditions thermal currents develop into large systems extending a mile or more horizontally and several thousand feet vertically; in these circumstances they may take many minutes to pass over a station and are no longer to be classed as turbulence or gustiness but are rather to be regarded as distinct secondary disturbances; the variations in wind take the form of squalls rather than gusts although rapid variations are superimposed by smaller eddies carried along in the system. The strongest gusts recorded are usually associated with such squalls. In the British Isles gusts of over 70 knots occur occasionally at almost any well-exposed station and recorded speeds have exceeded 90 knots from time to time on the western coasts. The strongest gusts are in many respects more important than the mean wind for they may be the cause of severe structural damage. A parked aircraft may easily ride out a gale defined as a mean wind of force 8 or over (34 knots and upwards), but the hazard is greatly increased if a single gust rises to say 60 knots even though the mean wind may not reach gale force.

The essential difference between squalls and gusts lies in the time factor. A gust is a transient increase in speed lasting perhaps for a few seconds, the squall is an increase in mean wind lasting usually for some minutes and then dying away again. A squall may be accompanied by a marked drop in temperature, a characteristic squall cloud and precipitation. Both speed and direction at the time of a squall may differ widely from the prevailing geostrophic values.

Diurnal variation of wind

When the field of pressure is steady, there is a noticeable difference in the surface wind as between day and night. It has been seen how the surface wind is reduced in speed compared with the free wind at a height of about 2000 feet and backed or veered according to the hemisphere. For a given pressure gradient, the precise value of the surface wind depends on the degree of turbulence, since this controls the mixing process in the friction layer. If this is vigorous the friction layer is deep, if weak it is confined to a shallow layer. The diurnal variation in turbulence due to thermal eddies over land entails therefore a diurnal variation in surface wind, resulting in lighter winds and greater deviation from the isobars at night, stronger with less deviation by day. In the upper levels of the friction layer the variation is in the opposite direction. During the night when turbulence is weak the undisturbed wind is reached at a lower height, perhaps at 500 or 1000 feet, whereas in the day-time these layers are well within the friction layer and are subject to reduction in speed and deviation in direction. Turbulent mixing has, in other words, the effect of smoothing out the differences in wind between the surface and the free air; in the day-time the difference is small, in the night-time large.

Diurnal variations of wind, depending on variations in convection, are to be expected only when thermal eddies develop by day and die away at night because of diurnal changes in surface temperature and lapse rate. There is therefore no observable diurnal variation over the open sea, while over the land it fails to appear when skies are continuously overcast. It is most apparent in fine weather with clear nights and sunny days. In these conditions a light wind of say 10 knots by day may fall practically calm at night. The diurnal variation should be kept in mind whenever a precise estimate of the speed and direction of the surface wind is required, or when it is desired to estimate the upper wind conditions on an aerodrome. Thus, if the wind on the aerodrome on a clear night is only 5 knots, the speed at 2000 feet may be as much as 20 knots, but with the same surface wind on a sunny day the upper wind would be unlikely to exceed 10 knots. It is impossible to lay down precise rules, as variations depend on various meteorological factors and particularly on local conditions. A veer in direction of 20 degrees and an increase in speed of the surface wind of 50 per cent from night to day is not uncommon; with light winds the percentage variation may be very much greater.

Wind and topography

The direction of the wind at low levels tends to some extent to conform with the contours of the land. Thus a stream of air flowing towards a mountain range tends to be deflected parallel to it, and in general the wind tends to flow round an obstruction if possible rather than over it; however, if the lapse rate is comparatively steep, upward motion takes place more readily and flow over high ground is facilitated with possibly the development of orographic cloud and precipitation as consequences. Should the range be broken by a pass or valley the air will be forced through with enhanced velocity, and gale force may be reached locally. In narrow valleys the wind flows almost invariably from one direction or the other along the valley and gives no reliable indication of the general direction of the wind; a small change in the pressure distribution may cause the wind in the valley to swing to the opposite point of the compass. The constriction of wind through valleys is spoken of as funnel effect or canalization; it is responsible for the great speed attained by the mistral in the Rhone valley (Fig. 21). Some characteristic types of wind which owe their existence to topography will now be considered.

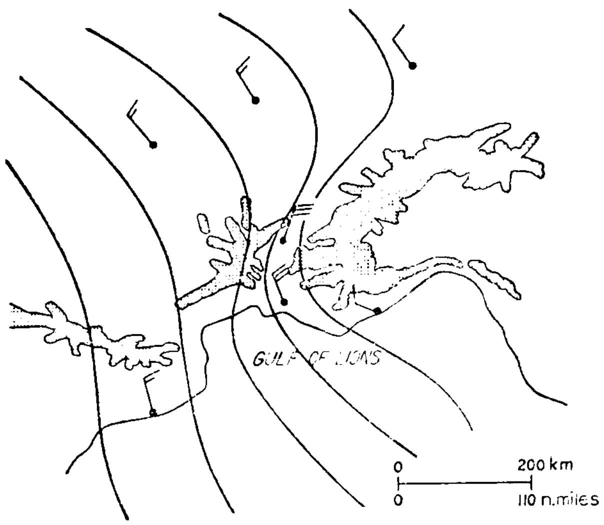


FIG. 21. *Funnelling down the Rhône Valley producing the mistral*

——— Isobars. Wind arrows in usual style.

Föhn wind

If on meeting a mountain barrier air is forced over the top, adiabatic cooling takes place which if it proceeds far enough leads to the formation of cloud and precipitation. In the cloudy stage, the rising air cools at the saturated adiabatic lapse rate; if some of the condensed water falls out as precipitation and if the air subsequently descends the lee slopes, then the latent heat which was liberated when the water vapour condensed to form clouds is only partly used to evaporate the cloud droplets in the descending air. After the cloud disperses the air is warmed at the dry adiabatic lapse rate. The larger the amount of water deposited on the mountain, the warmer the air after descent. The local name for these warm dry winds in the Alps is the föhn.

There may be a delay in the onset of the föhn wind in a mountain valley because of the reluctance of the cold, denser air to clear. The warm föhn wind may blow over a valley at considerable strength without affecting the lower altitudes at all. However, if the warm wind is strong a 'swilling out' may suddenly take place and the cold air is displaced very quickly. When this happens the valley wind can increase from zero to gale force within a few minutes, the time of onset being very difficult to predict. Taking the saturated lapse rate as half the dry lapse rate of 3 degC per 1000 feet, an ascent of saturated air and a subsequent equal descent at the dry adiabatic lapse rate gives an increase of temperature of 1·5 degC for each 1000 feet of displacement. High mountains may therefore produce warming of some 10 degC or more.

The term föhn wind has been adopted in meteorological terminology as a general name for the phenomenon wherever it exists. The chinook of the Rocky Mountains is another example, although it is not necessary to look so far afield for an illustration of the effect. It is noticeable on the lee side of high ground even in the British Isles, and accounts, for example, for the clear warm air sometimes found on the east coast of Scotland when moist south-westerly winds pass over the mountains.

Anabatic and katabatic winds

When the surface of the ground has an appreciable slope there is frequently a tendency for the wind to drift up or down the slope. If the slope is heated by the sun the air in contact with it becomes warmer than the free air at the same level; it is therefore lighter and tends to ascend. Such ascending winds are called anabatic winds. They are masked by irregular convection, and may not show clearly as a definite current of wind except perhaps where they are intensified by the funnel effect of a valley when the name 'valley wind' may be used. Except near a coastline where an up-slope wind is augmented by the sea-breeze, anabatic winds are seldom of much significance.

The reverse, or katabatic, effect is also found over sloping ground in favourable circumstances (Fig. 22). The necessary condition is that the air over the slope shall be colder than air at the same level in the free atmosphere, so that it will tend to sink; the condition is therefore present when the ground is relatively cold. As surface cooling takes place by radiation at night, katabatic winds are normally a nocturnal phenomenon. Even in gently sloping country with no great elevation a down drift of cold air occurs on a clear quiet night; the speed of the wind may be not more than a few knots, but it aids the formation of pools of cold air on low-lying ground and is one factor controlling the local incidence of frost, mist and fog.

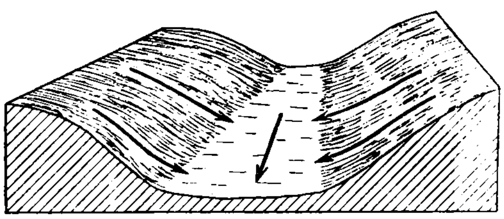


FIG. 22. *Katabatic winds*

Anabatic winds are in the opposite direction and are usually lighter.

On mountain slopes the effect is increased, and if the ground is snow covered it may occur during the day-time as well as at night. A well-known case of a vigorous wind of this type is the bora, an off-shore wind on the northern shores of the Adriatic. It normally sets in suddenly, and frequently reaches well over gale force with gusts of over 100 knots; it is extremely dangerous to shipping and to low-flying aircraft. In this and in several similar cases the wind is not purely katabatic, for it is influenced by the general pressure distribution and is locally intensified by the funnel effect down valleys extending towards the coast. Other examples may be found in different parts of the world; on the coasts of Greenland and the shores of the Black Sea they are well known.

The distinction between the föhn and the katabatic effects should not be overlooked. Both give downcurrents off high ground, but the former results from the general pressure distribution and is warm and dry, while the latter is a local development where surface cooling is more effective than adiabatic warming so that the wind is relatively cold even when it reaches sea level.

Ravine winds

These occur in and near ravines or narrow valleys which penetrate a mountain barrier. When there is a pressure difference, level for level, between the two sides of the barrier, air is impelled through the ravine by the pressure gradient. Such winds may be very strong not only in the ravine but also after leaving its mouth, where they flow out into open country. Examples are the ravine wind at Genoa due to the difference of pressure between the Po valley and the Gulf of Genoa, the kosava of the Danube south-east of Belgrade which sometimes exceeds 35 knots, and the vardar winds near Salonika.

Land-breezes and sea-breezes

In coastal districts the large diurnal changes of wind are a characteristic feature and are most marked in quiet conditions with sunny days and clear nights. The wind blows from the sea during the day-time, often setting in rather abruptly during the morning. If however there is a light off-shore wind to be overcome, its onset is delayed, perhaps even until the afternoon, but then takes place suddenly and may be accompanied by a minor squall with a sharp fall of temperature and an increase of humidity. The strength of a pure sea-breeze rarely exceeds 10 knots in the British Isles; above about 500 feet the speed rapidly decreases to become practically negligible in most cases by about 1000 feet. The sea-breeze sets in first near the coast, often being observed a mile or so out to sea, and the area affected gradually broadens both inland and seawards during the day. As a rule the breeze does not extend more than 10–15 miles either side of the coastline, although there are numerous cases of much greater extent, at least over the land, which are probably to be explained by the effects of topography or by inflow caused by the diurnal fall of pressure over the land. In temperate latitudes there is often sufficient pressure gradient to modify these local developments, and the diurnal variation in the vicinity of the coast then becomes complicated or even obliterated. It is necessary to make a special study of each section of coastline in order to forecast accurately the diurnal variation of wind.

Although the sea-breeze moves at first directly on shore, the Coriolis effect becomes more apparent as the air arrives after a longer track over the sea and the flow then tends to align itself with the coastline with the land on the left (on the right in the southern hemisphere). This process however takes time, and may not be reached until evening when the diurnal decline of the sea-breeze has already begun. The sea-breeze normally falls light soon after sunset, and after some hours may be replaced by an opposite drift from the land – the land-breeze. In the British Isles the land-breeze is rarely more than light, a few knots at most, and does not develop with any marked regularity; since it occurs only in stable conditions, its direction over the land follows the ground contours closely; it is also shallower than the sea-breeze, probably not extending above a few hundred feet.

On account of the greater intensity of insolation, the sea-breeze in lower latitudes may reach 15–20 knots; throughout most of the tropics and subtropics, where prevailing winds are normally light, it is of very regular occurrence.

The explanation of land-breezes and sea-breezes rests on the diurnal differences of temperature between the surfaces of land and sea (Section 12). The increase of temperature of the land by day causes the overlying air to warm and expand, so that above any upper level such as 500 feet the weight of air is increased. In this way the pressure aloft over the land becomes slightly greater than before, while over the sea it remains at first unchanged. Thus a drift of upper air commences from the land towards the sea. This in turn produces a slight increase of surface pressure over the sea and a reduction over the land; in consequence a flow of surface air sets in from sea to land and constitutes the sea-breeze. At night the temperature difference is reversed owing to radiative cooling over the land and a contrary flow is induced at the surface. In both cases the upper return flow of air is spread over so great a depth that it is no more than a gentle drift, with scarcely any observable wind; but if the sea-breeze has established itself against a light off-shore wind, then this may still be found above the sea-breeze at a height of perhaps 1000 feet or more.

On some occasions the effect of the sea-breeze extends to 3000 or 5000 feet. The warmer flow from the land lifts over the cooler air from the sea and the transition zone between the two types of air is often marked by a line of small cumuliform cloud. This transition zone may progress slowly inland as a sea-breeze front. It has small dimensions and although of interest to glider pilots, its effect would normally be unnoticed by powered aircraft.

A cross-section of a well-developed sea-breeze is shown in Fig. 23.

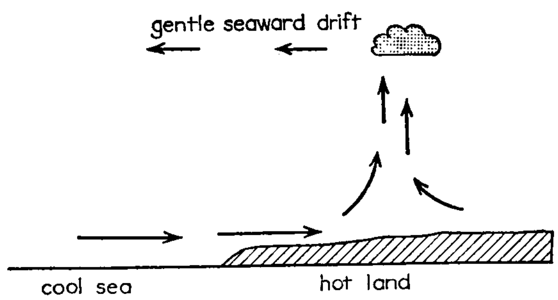


FIG. 23. *Cross-section through a well-developed sea-breeze*

A land-breeze is in the opposite direction and usually lighter.

26. WIND IN THE FREE ATMOSPHERE

Wind in relation to pressure contours

It has been explained in Section 24 how the wind just above the friction layer, that is at about 2000 feet, is related to the distribution of pressure at the surface. Both wind and pressure should strictly refer to the same level, and the relationship then holds good not only at 2000 feet but at any greater height. Thus, for example, at 20000 feet the formula (8) for the geostrophic wind remains true, but since the density at this height is reduced to about one-half its value at the surface, the geostrophic speed at 20000 feet is approximately double that at the surface for the same pressure gradient. Consequently allowance must always be made for the density if winds are derived from isobaric charts for the upper air, and the allowance will differ for each level.

There are considerable advantages to be gained from an alternative method of representing the distribution of pressure in the upper air. Consider a fixed pressure of, say, 500 millibars, and arrange that the heights at which this pressure occurs are plotted on a chart, using simultaneous observations from a network of upper air stations. The heights of course differ from place to place and are conveniently shown by drawing the lines of equal height, or contour lines, usually at intervals of 60 metres. The completed contour chart shows the variations in height of the chosen pressure surface, in exactly the same way as height of ground is indicated on an ordinary survey map (see Fig. 24).

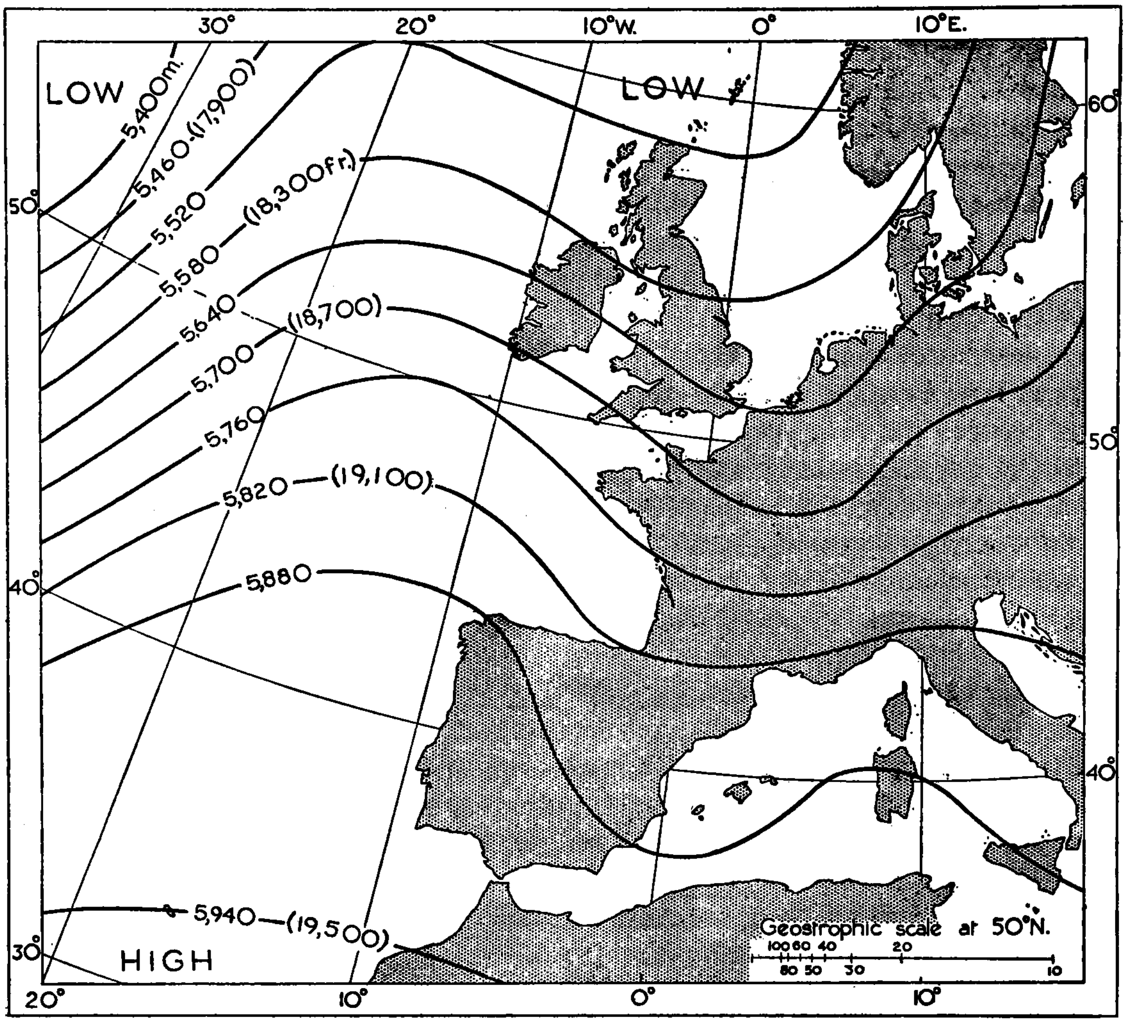


FIG. 24. *Contours of 500-millibar isobaric surface*

Any contour line is by definition horizontal and the pressure is the same at all points; it therefore satisfies the definition of an isobar on a horizontal surface. Thus in Fig. 24 the contour line for, say, 5520 metres (18 100 feet) is an isobar in the horizontal surface at that height. Consequently the direction of the geostrophic wind is along the contours, and it is easily seen that it blows with the lower values of the contours on the left in the northern hemisphere and on the right in the southern hemisphere.

The speed of the geostrophic wind is determined by a formula which is even simpler than that which applies to a chart of isobars. The upper part of Fig. 25 illustrates a portion of a contour chart for 500 millibars in which PBQ is drawn normal to the contours; the lower part shows a vertical section through PQ. By applying equation (8), the geostrophic wind at B is given as

$$V = \frac{1}{2\Omega\rho \sin\phi} \times \frac{p_C - p_A}{AC}.$$

Now the pressure at C exceeds 500 millibars by the weight of the column of air of unit cross-section represented by Q′C therefore

$$p_C - 500 = g\rho(h_Q - h_C).$$

Similarly the pressure at A is less than 500 millibars by the weight of the column P′A, and

$$500-p_A=g\rho(h_A-h_P).$$

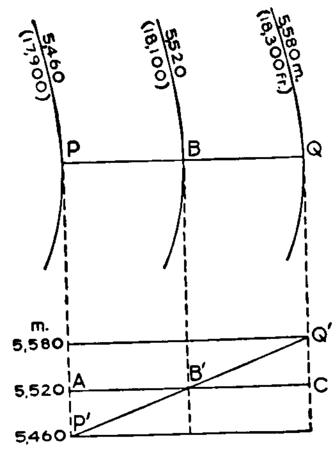


FIG. 25. *Vertical section through a pressure surface*

The upper part of the figure shows contours of the 500-millibar surface and the lower part shows the vertical section. The vertical scale is greatly exaggerated.

By addition, since $h_A=h_C$,

$$p_C-p_A=g\rho(h_Q-h_P).$$

If this is substituted in the above expression then, since AC = PQ, the equation for the geostrophic wind in terms of the contours becomes

$$V=\frac{g}{2\Omega\sin\phi}\times\frac{h_Q-h_P}{PQ}. \qquad \ldots . (13)$$

The density does not appear in this equation and the second factor is simply the gradient of the contours, that is, the difference in height between two contours divided by the horizontal distance between them, analogous to the pressure gradient on an isobaric chart. In a given latitude, the geostrophic speed varies only with the gradient of the contours. By the aid of this equation, a geostrophic scale for use with contour charts may be constructed similar to that in use with isobaric charts; since the equation does not contain the density, the same scale applies directly to the appropriate contour chart whatever the pressure level. The scale is illustrated at Fig. 15.

The relation of the geostrophic wind to the contours on a pressure surface may now be stated by the following rule:

> The geostrophic wind blows along the contours of a pressure surface with a speed which is proportional to the gradient of the contours and independent of the density; the direction is such that the lower contours are on the left in the northern hemisphere and on the right in the southern hemisphere.

As was explained in connection with isobars (Section 24), it follows that equation (13) gives the component of geostrophic wind at right angles to PQ, even when PQ is not normal to the contours.

The cyclostrophic and gradient winds may be determined from contour charts in the same way as from isobaric charts as already described, and there is no need to repeat the argument. In general, the limitations of the geostrophic wind derived from the surface chart, as discussed above, apply equally to the upper air charts except that any effects due to the proximity of an irregular land surface are absent except possibly in the vicinity of high ground. It should be remembered too that the geostrophic equation is in any case inapplicable near the equator.

Thermal wind

The general nature of the variation of wind with height may be envisaged by considering how the pressure field at the surface becomes modified at higher levels; this was discussed in Section 7 where the conclusion was reached that in areas where the temperature is high, the pressure in the upper air tends also to be high, and where the temperature is low the upper air pressure also tends to be low. In terms of contours of a given pressure surface, we should say that they are relatively low over cold places and high over warm places. In either case it has been shown that the wind at high levels tends to orientate itself with the lower temperature on the left (in the northern hemisphere).

This connection between wind and temperature becomes more precise when it is noted that the ratio of the pressures between two heights depends only on the mean temperature of the intervening column of air, in accordance with equation (2), Section 7. To take the simplest case, when there is no pressure gradient and therefore no wind at the surface the pressure distribution and hence the geostrophic wind at an upper level is determined solely from the horizontal distribution of mean temperature. This we now generalize into the statement that the vector difference of wind between two levels depends only on the horizontal distribution of mean temperature of the intervening layer of air. The thermal wind in a layer is defined as that wind which must be added vectorially to the geostrophic wind at the lower level in order to obtain the geostrophic wind at the upper level (Fig. 26).

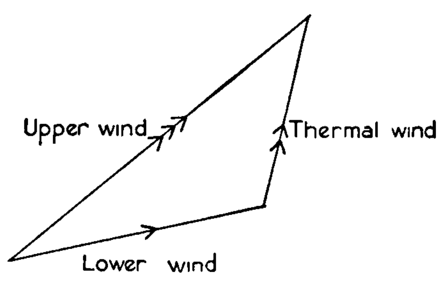


Fig. 26. *Vector diagram illustrating thermal wind*

Thus the thermal wind depends only on the horizontal distribution of mean temperature in the layer. The latter may be represented by drawing isotherms, and the relation of these to the thermal wind is given by the following rule:

> The direction of the thermal wind is along the isotherms of mean temperature with the lower temperature on the left in the northern hemisphere, on the right in the southern hemisphere; the speed of the thermal wind is proportional to the temperature gradient.

It is not necessary to reproduce the exact formula for the thermal wind, but in latitude 50° it leads to the following easily remembered rule:

If the gradient of mean temperature is 1 degC per 100 nautical miles, the speed of the thermal wind in knots is approximately equal to the thickness of the layer in thousands of feet.

If the latitude ϕ differs much from 50°, the speed so obtained should be multiplied by the factor sin 50°/sin ϕ.

As an example, suppose the geostrophic wind at 10 000 feet is 270° 40 kt while the gradient of mean temperature through the layer 10 000–20 000 feet is 3 degC per 100 nautical miles with the lower temperature due west; then the thermal wind at the top of the layer is 180° 30 kt which added vectorially to the wind at the bottom of the layer gives 233° 50 kt for the geostrophic wind at 20 000 feet.

Conversely, if the wind is known at two heights, the vector difference gives the thermal wind over the intervening layer and hence the magnitude and direction of the gradient of mean temperature in the layer.

Thickness charts

Isotherms of mean temperature, on which the thermal wind in a layer depends, are not normally drawn directly, but when contour charts are prepared it is usual to construct further charts showing the difference in height between pairs of standard pressure levels, e.g. 1000 and 500 millibars. As at any point the difference in height depends only on the mean temperature of the intervening column of air, isopleths of the difference coincide with the isotherms of mean temperature. Such charts are called 'thickness charts'; by representing how the thickness of the given layer varies from place to place, they show how its mean temperature varies in the horizontal, and from this, by use of the rules given above, the distribution of the thermal wind is made evident.

Variation of wind with height

The average variation of wind with height within the friction layer has already been described in Section 25; the present section describes the variation within the remainder of the troposphere and in the stratosphere. The notion of the thermal wind enables some immediate conclusions to be deduced as to the distribution of wind in the upper levels. It has been seen (Fig. 13) that temperatures in the troposphere for the most part decrease from the equator towards the poles. The average thermal wind throughout most of the troposphere accordingly blows from the west and the usual variation of wind with height is such that the westerly component increases throughout the troposphere. In the upper troposphere the winds are in consequence predominantly westerly. In terms of pressure this implies that at levels high enough for the surface distribution to be overcome there is a depression over each pole, while pressure is relatively high in the subtropics. Within the tropics such inferences cannot be made since there the formulae for the geostrophic, and consequently the thermal, winds break down. The actual distribution of wind in the tropics consists in the main of a belt of easterly winds extending up to at least 40 000 feet and covering about 10–20 degrees of latitude but varying in position according to the season.

Outside the tropics, as already implied, low-level westerly winds usually increase with height with little change of direction, while easterlies tend to decrease and eventually to give way to westerlies. On this account easterlies are often shallow, the reversal of direction perhaps taking place by 3000 feet although at other times they may persist to great heights; further, the reversal of direction may take place rapidly with a shallow calm layer separating the two currents, or by a gradual backing or veering spread over several thousands of feet. A westerly thermal wind also implies that northerly winds would be expected to back with height and southerly winds to veer.

In the low stratosphere in winter, temperature is lowest in polar regions and increases with increasing distance from the poles until a maximum is reached usually between 40 and 60° of latitude although the mean latitude of the maximum varies enormously from month to month and year to year. Thus in high latitudes in winter, westerly winds increase with height from the troposphere into the stratosphere. In lower latitudes, equatorward of the temperature maximum, westerlies decrease with height above the tropopause and eventually become easterlies which then go on to increase with height (see Section 72).

In the summer hemisphere on the other hand, temperatures in the stratosphere are highest in polar regions and lowest over the tropics resulting in an easterly thermal wind, and this distribution of temperature is relatively constant, in contrast to the variability in winter. In the same way as in low latitudes in winter, the speed of the westerlies above the tropopause is reduced and eventually there is a reversal to easterlies which then increase with height.

In winter, probably the strongest westerly winds of the northern hemisphere are high up in the stratosphere in high latitudes but the strongest westerlies of the troposphere occur in a belt whose axis lies between 25 and 40°N at about 40 000 feet. In summer the thermal gradients are weaker and the belt of maximum wind is displaced to between 40 and 45°N. The average wind speed* in these zones is about 70 knots in winter and 40 knots in summer. There is a somewhat similar distribution of wind in the southern hemisphere but with a smaller seasonal variation in speed. Superposed on these average conditions are variations with longitude depending on the local temperature gradients. On account of exceptional contrasts of temperature between continent and ocean in the vicinity of the south-east coasts of North America and Asia, the westerly winds in these areas average or even exceed 100 knots in winter at the heights stated (see also Chapter 22 and Figs. 155 and 156).

In the stratosphere the westerly component, which is usually strong at the tropopause, continues to show itself up to great heights but with decreasing speed, except in high latitudes of the northern hemisphere in winter where the westerlies continue to increase with height far into the stratosphere in accordance with the low temperatures over the polar regions at that time. Observations in middle latitudes of the northern hemisphere indicate a predominantly westerly direction in winter up to at least 100 000 feet but in summer easterly winds become established above about 60 000 feet and then steadily increase with height. The average speed of the westerlies exceeds 40 knots at 100 000 feet while individual speeds have been known to reach 130 knots; with the easterlies of summer the speeds range up to about 50 knots. The normal decrease of the westerly component with height is not however to be relied on invariably, as on particular occasions the distribution of both temperature and wind may be quite different from the average.

27. VERTICAL MOTION OF THE ATMOSPHERE

Vertical motion of the atmosphere ranges in type from irregular local gusts and lulls with a period of a few seconds, to widespread and sustained, although slow, movements lasting for days at a time; the vertical velocity, according to the type

*This is the speed of the mean vector wind. The average speed computed irrespective of direction is about 10 knots greater.

of motion, ranges from zero to many tens of feet per second. Such motions are both of immediate concern to the flying of aircraft and of fundamental importance to weather. The irregular upward and downward currents are responsible for bumpiness, while the more persistent local currents may cause aircraft to undergo marked vertical displacements or may adversely affect the rate of climb. They make difficulties for powered aircraft but for the glider pilot they are the means of sustained flight when used skilfully. The effects of all forms of vertical motion on weather are far reaching; in particular, upward motion is an essential prerequisite to the formation of clouds and precipitation, while downward motion is often associated with clear skies. These different types of vertical motion will now be described, but their consequences with regard to cloud and precipitation are discussed in the subsequent chapter.

Vertical currents produced by ground contours

In describing the flow of wind near the ground (Section 25), the turbulent character of the motion has been emphasized as a regular feature, even over a level surface; only on occasions of light wind and temperature inversion does the flow remain smooth. Individual eddies include upward and downward movements as well as horizontal variations in speed and direction; they may be of frictional or thermal origin, and when the size of the eddies is comparable with that of an aircraft, they are felt as short-period bumpiness. When there are obstacles in the path of the wind, such as trees and buildings or changes in the ground contours, then the character of the windflow undergoes a change. This takes place in one or more of the following ways: by a general deflection of the stream in order to get round or over the obstacle, by the development or intensification of turbulence, and by the formation of lee waves. The extent to which any of these changes takes place depends on several factors which include the wind speed and its variation with height, the lapse rate of temperature, the size and shape of the obstructions and their orientation relative to the wind.

Smooth and turbulent flow. The flow of air in the neighbourhood of an isolated hill, or an other object, is usually deflected, partly over the top and partly laterally (Fig. 27).

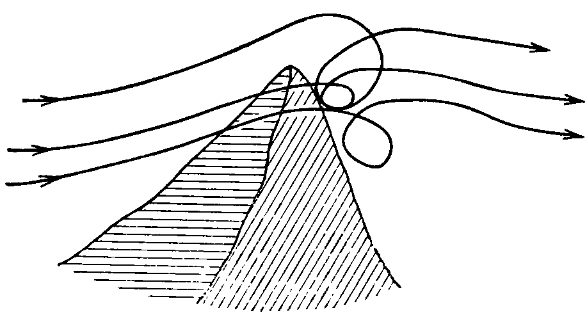


FIG. 27. *Flow near an isolated hill*

With a continuous line of hills stretching across the wind the possibilities of lateral displacement are limited and the height to which the airflow is influenced by the barrier is increased. It may reach to about four times the height of the ridge when the lapse rate is stable and no inversion is present, although much greater heights may be affected when lee waves develop.

When the wind is light, 15 knots or less, and the lapse rate is stable, the airflow over a range of hills is a smooth shallow wave with only feeble vertical currents and no downstream phenomena.

Over hills with steep slopes, or with slightly greater wind speeds, the flow may be modified by the formation of stationary eddies. The air flows smoothly above them through a shallow wave (Fig. 29.)

With strong winds, vertical currents may be quite extensive and turbulence may be greatly intensified. Frictional eddies form repeatedly in the vicinity of the ridge, break away, and drift downwind before they finally dissipate.

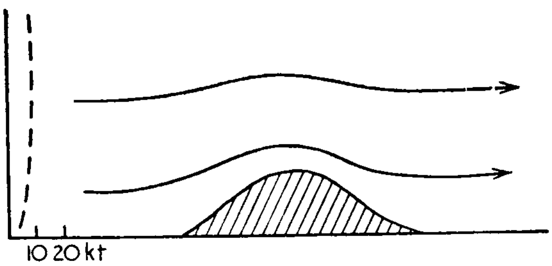


FIG. 28. *Steady flow over a hill, without eddies*
- - - - Vertical profile of wind

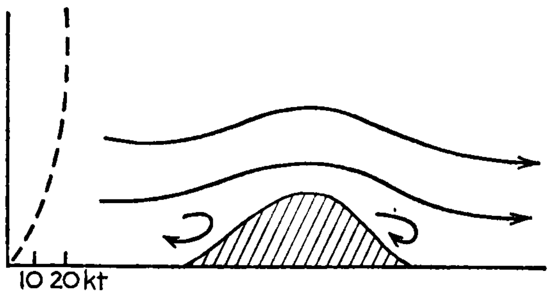


FIG. 29. *Steady flow over a hill, with stationary eddies*
- - - - Vertical profile of wind

Similar considerations apply to the flow of air over a cliff or escarpment or over obstructions such as aircraft hangars and other buildings. Fig. 30 illustrates the motion for light and for strong winds. Generally, the smoother or the more streamlined the obstruction, the less the disturbance it causes to the flow; the more bluff the object, the more intense is the resulting eddy motion.

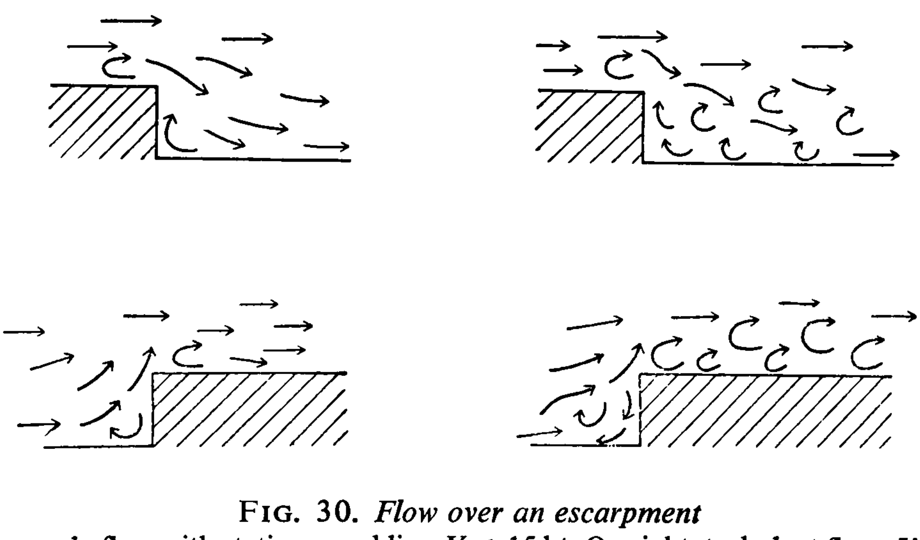


FIG. 30. *Flow over an escarpment*
On left, steady flow with stationary eddies, $V < 15$ kt. On right, turbulent flow, $V \geqslant 15$ kt.

The vertical currents so far described have a noticeable effect on aircraft encountering them. When, as is usually the case, the size of the eddies is comparable with the size of an aircraft, they produce short-period bumpiness and this may at times be severe, particularly to leeward of the obstruction when the wind is strong. Such turbulence increases the difficulty of landing on or taking off from an aerodrome situated downwind of a hill. Moreover in taking off over a hill and generally when flying near hills adequate clearance should be allowed to avoid the effect of possible downcurrents on the slopes. The stationary eddies are to be avoided not only because of the downcurrents but also because an aircraft encountering the reversal of direction might have its airspeed momentarily reduced below the stalling speed. It is clear that special attention should be given to the wind flow when flights are made near hills and obstructions, and aerodromes should be located well away from hills whenever possible.

Standing waves. In suitable conditions in addition to the disturbance over the ridge, waves frequently occur to the lee of the hills or mountains. The waves remain in a fixed position with the air moving through them. They can be regarded as oscillations about the stable state of the undisturbed airstream with the mountain providing the source of the disturbance and gravity providing the restoring force.

Over the mountain the flow is deflected by the high ground and in the lower layers tends to follow the profile of the ground. Beyond the mountain a lee-wave flow is established which has regular undulations which bear little relationship to the ground below (Fig. 31).

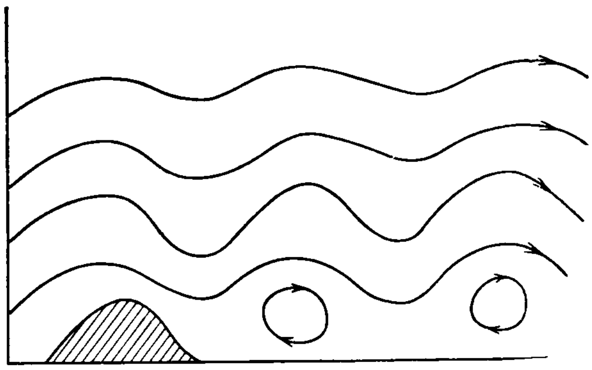


FIG. 31. *Flow over a ridge, with lee waves*

Vertical currents may be quite extensive and vigorous; speeds of 30 feet per second (10 metres per second) have been encountered over mountains in the British Isles while in powerful waves in the Sierra Nevada, California, speeds of 75 ft/s (25 m/s) or more have been observed. The amplitude of the mean wave increases from the ground upwards, achieving a maximum in the middle troposphere, and then dies away. The wave effect spreads upwards, sometimes as high as 80 000 feet; occasionally cloud at cirrus levels and noctilucent cloud show a wave-type formation.

The existence of waves requires certain conditions to be fulfilled regarding the distribution of wind and temperature with height. For strong wave development the airstream must consist of a deep current of air in which wind direction changes little with height and speeds increase upwards through the troposphere. The wind must blow within about 30° of the perpendicular to the ridge and the speed at the crest must be above 15 knots for small mountains and above 30 knots for large mountains. Powerful waves develop when, at levels where the airstream is disturbed by the mountain, there is a layer of marked stability, such as an inversion or an isothermal layer, bounded by less stable air above and below.

Fig. 32 shows computed streamlines for a hill about 2000 feet high, where a stable layer extends from 4000 feet to 9000 feet above the low ground and the wind speed increases with height throughout the layer in which the lapse rate is stable.

Wavelength and amplitude. When the wind and temperature profiles of an airstream would favour the formation of waves, their natural wave-length is commonly a few miles, although it may be anything up to 30 miles. Light winds or marked stability through a substantial depth make for short wavelength, whilst strong winds and slight stability are associated with a long wavelength.

The amplitude of the waves depends largely upon the size and shape of the high ground. In general, the higher the mountain the greater the amplitude of the lee waves, but in addition the width of the mountain is relevant. For a given airstream and height of hill, the lee waves of largest amplitude are generated by those hills which have a cross-section which coincides with the wavelength natural to the wind and temperature conditions of the airstream. A ridge whose wavelength matches the one natural to the airstream will give lee waves of larger amplitude than will another much larger hill.

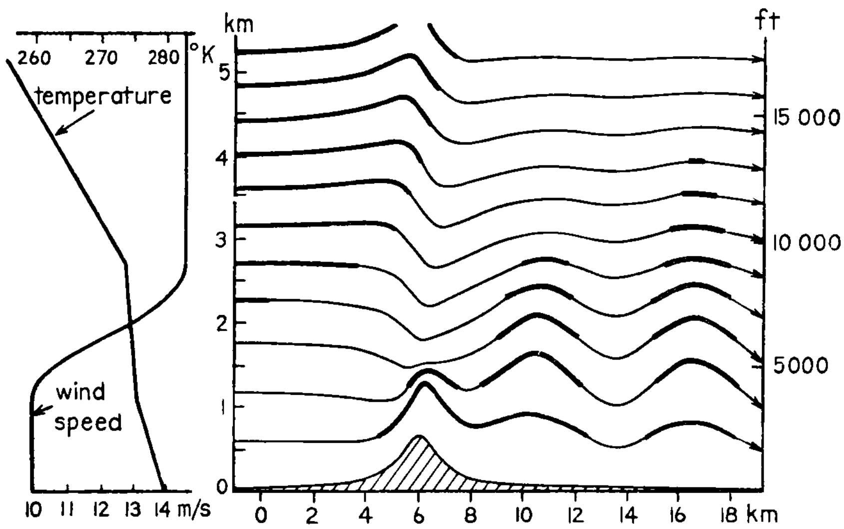


FIG. 32. *Computed streamlines over a ridge in conditions favourable for the development of standing waves* (after R. S. Scorer*)

Vertical profiles of wind and temperature are shown on the left.

Turbulence. The friction layer is more variable in depth over irregular mountainous country than over level ground; eddies form in the layer and turbulence is vigorous. At higher levels, flight through mountain waves is often very smooth, but turbulence can occur as a result of wave breaking and is likely to be severe in the rotor zones; these zones tend to form beneath the crests of the waves and sometimes form a roll cloud.

Turbulence at high levels in the vicinity of the tropopause and the jet stream appears to be more frequent and more intense over mountainous country than elsewhere.

Rotor streaming. This should not be confused with the rotor zones which form under the crests of standing waves.

If a very strong wind extends through a restricted depth comparable with the height of the mountain, i.e. the wind becomes much lighter at some level in the upper air, then standing waves are unlikely and rotor streaming may occur (Fig. 33). In these circumstances severe turbulence is encountered in layers level with the hill top and probably up to two or three times the height of the hills.

Clouds. Orographic clouds frequently form over mountains and where a lengthy lee slope occurs the turbulence may be rendered visible by the cloud fall sweeping down the lee slope. Clouds sometimes form in the rotor zone and lenticular clouds

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*London, Ministry of Aviation. Atmospheric turbulence and its relation to aircraft. Proc. Symp. RAE Farnborough, November 1961. London, HMSO, 1963.

frequently form in the crests of lee waves. These clouds are a visible indication that lee waves are operating, but such waves often occur without cloud formation.

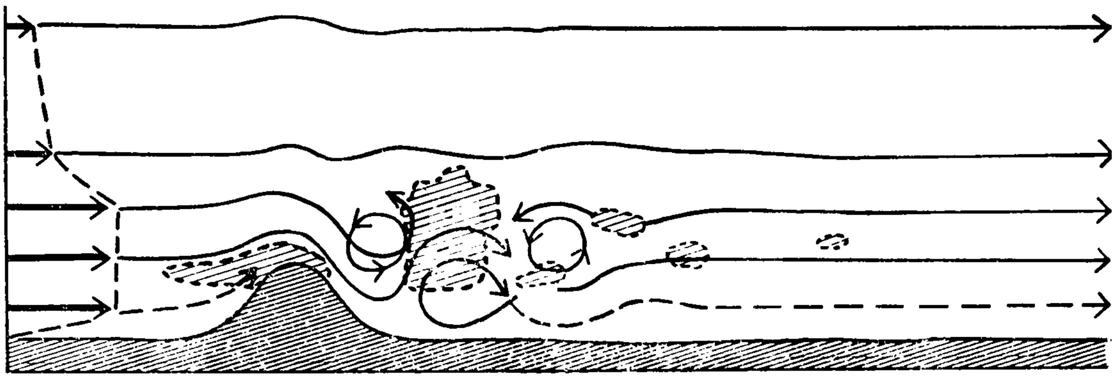

FIG. 33. *Rotor streaming* (after Förchtgott)

The vertical profile of wind is shown at the left with bold arrows.

Icing. The displacement of air in mountain waves allows little chance of entrainment so the ascending motion is very nearly adiabatic. In stable air the 0 °C isotherm is lowered by the passage of air over high ground. The supercooled water content may increase. For these reasons mountain waves are likely to increase the severity of ice accretion on aircraft in cloud.

Safety heights over high ground. When mountain waves are operating, the safety height for an aircraft overflying the high ground should be increased to allow for the downdraughts and to avoid the turbulent rotor zone. The additional height margin will vary widely in different circumstances and methods of calculating it can be found in the relevant civil information circulars (see Appendix III).

In a mountainous area such as the Alps where air flows over a succession of ridges, wave systems from one crest may interfere with those from the following crest. In certain cases the waves may be damped out but in others, where the waves are in phase, greater vertical currents may develop.

Convection currents

Thermal currents resulting from the heating of the atmosphere at the ground were noted in Section 17. They often appear in the form of small irregular eddies which, rising from the ground and being carried along by the wind, are practically indistinguishable from the eddies of frictional turbulence, or they may take the form of more pronounced upward and downward currents possibly arranged in a regular pattern. Differential heating of the ground has the effect of localizing the vertical currents; upward currents tend to originate over the warmer parts of the ground such as dry soil, rock or sand, while downward currents tend to be associated with relatively cooler patches such as grassland, wooded areas and especially water surfaces. The bumps so produced are often marked and their association with variations in terrain is familiar to pilots. Convection currents and bumpiness also occur at other levels in the troposphere, usually in association with cloud formation, whenever the lapse rate becomes unstable.

The height to which convection currents extend is controlled largely by the lapse rate of temperature. A marked inversion with base at a height of a few thousand feet limits the upward penetration of convection currents originating at the surface. In such cases considerable, even intense, bumpiness is experienced from ground level up to the base of the inversion, while above the air is smooth. On the other hand, when the lapse rate in the free atmosphere is less stable, or if the surface heating is sufficient to overcome the stable tendency the vertical currents may become very strong and perhaps extend through the entire troposphere. Wherever upward currents exist there must be compensating downcurrents in the vicinity, but these are usually spread over a wider area and so are less intense; moreover the downcurrents are often found in the clear air between the clouds. The bumpiness accompanying extreme currents is severe; even the downcurrents may cause an aircraft to lose height rapidly. In these conditions ample clearance must be allowed above ground, particularly when flying over hilly country, while conditions for landing and taking off become difficult, especially for light aircraft.

Pronounced ascending currents, as well as many of less intensity, are responsible for the development of convection clouds and will be further considered in the next chapter.

Widespread vertical motion

So far we have considered the vertical motion associated with disturbances of relatively small horizontal dimensions caused by turbulence, ground contour or local convection currents and covering areas of at most a few miles in extent. Now even the largest types of wind systems, the large anticyclones and depressions, involve vertical as well as horizontal motion. Being distributed over extensive areas, the upward or downward velocity is very small, only a few feet per minute, and its direct effect on aircraft is quite insignificant. The importance of this type of motion lies in its effects on the weather; for the motion may continue for several days and, though slow, may cause large masses of air to rise or fall through many thousands of feet. The problem is bound up with the general theory of depressions and anticyclones and is discussed in this connection in later chapters, but it is appropriate to consider some of the more general characteristics at this stage.

Divergence and convergence

Fig. 34 illustrates the types of motion known respectively as divergence and convergence. Air moving towards the points A and C from opposite sides constitutes an example of horizontal convergence, while in the vicinity of B there is horizontal divergence. There is in such cases a tendency for air to accumulate at A or C which must be exactly counterbalanced by vertical motion at these points; if the horizontal flow is taking place in the lower levels, the vertical motion

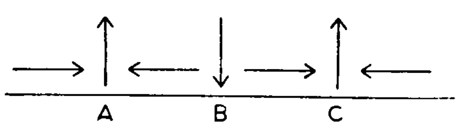

FIG. 34. *Horizontal divergence and convergence near the surface and the related vertical motion*

must necessarily be upwards. Similarly at B the net effect of horizontal motion is to remove air, its place being taken by air descending from above. Opposing motion is not essential for the production of convergence or divergence, as may be seen from the conditions of Fig. 34 by superposing a horizontal wind strong enough to overcome the flow in one direction. An example is provided by wind blowing from sea to land, even where there is little difference in height of the surface. The wind speed as the air passes over the land is reduced by the greater friction with the result that air arrives at the coastline faster than it is removed inland; the excess is forced upwards and sometimes results in a line of cloud parallel to the shore.

The association of convergence with ascending motion and of divergence with descent of air holds not only near the surface but also at higher levels. In fact dynamical considerations indicate, and it is certainly found in practice, that the large-scale movements of convergence, ascending motion and cyclonic circulation are usually linked together in one system and divergence, descent and anticyclonic motion in another.

Vertical motion on a large scale may have marked effects on the lapse rate of temperature. Thus descending air is subject to adiabatic heating; moreover the air must at some level spread out horizontally, or diverge. The lower layers of the subsided air accordingly become warm and dry compared with the air immediately below, where the latter has not taken part in the descent. In this way an inversion of temperature is formed, known as a 'subsidence inversion'; it is usually recognizable from the results of an upper air sounding by the presence of dry air above the inversion. On the other hand, general ascending motion often leads to a steepening of the lapse rate, but since in this case condensation and cloud formation are usually involved, its discussion will be deferred to a later chapter (Section 35).

Non-divergence of geostropic wind. Suppose PQ, P′Q′ (Fig. 35) are two neighbouring isobars, not necessarily straight and parallel, and that the flow of air

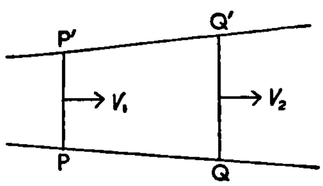

FIG. 35. *Non-divergence of geostrophic wind*

between them is geostrophic at all points. The rate at which air is transported across the section PP′, supposed of unit vertical thickness, is $V_1 \times \rho_1 \times \mathrm{PP}'$ where V_1 is the geostrophic wind speed at this point, while the transport across QQ′ is $V_2 \times \rho_2 \times \mathrm{QQ}'$. By the geostrophic equation (8), if the pressure is p on PQ and p' on P′Q′

$$V_1 = \frac{p-p'}{\mathrm{PP}' \times 2\Omega\rho_1 \sin\phi},$$

with a similar expression for V_2; therefore

$$V_1\rho_1 \times \mathrm{PP}' = \frac{p-p'}{2\Omega \sin\phi} = V_2\rho_2 \times \mathrm{QQ}'.$$

The mass of air carried across PP′ is thus identical with the flow across QQ′, and, since in geostrophic motion there is no flow across the isobars, there can be no tendency to accumulation of air within the area PQQ′P′. The motion therefore remains horizontal, and this is always so when the winds are geostrophic. This result is expressed by saying that geostrophic winds are non-divergent. Thus the spreading apart or closing together of isobars do not by themselves imply the existence of divergence or convergence. The terms 'diffluence' and 'confluence' respectively have been introduced to denote such changes in the pattern not only of the isobars but also of isopleths such as streamlines and thickness lines.

It follows from this result that when vertical motion is present the horizontal wind cannot be exactly geostrophic. The importance of vertical motion for cloud and precipitation has been emphasized and it now follows that such conditions are necessarily related to departures from geostrophic flow. Regions in which the wind is not geostrophic and in which vertical motion occurs can be recognized with the aid of synoptic charts as will be explained in Part III, but quantitatively the problem is one of great difficulty. This is partly because both the geostrophic departure – the 'ageostrophic' component (Greek prefix 'a' meaning 'not') – and the associated vertical component are too small to be readily measured, and partly because theoretically the problem has not yet been adequately solved.

Divergence and convergence due to surface friction. One effect of surface friction is to reduce the air flow near the surface to below the geostrophic value and to cause a flow of air across the isobars towards the side of lower pressure. When the total flow across a closed isobar is considered, it is seen as in Fig. 36 to lead to convergence towards the centre of a low-pressure system and divergence away from the centre of a high. This type of convergence is confined to the friction layer; calculation shows that the effect is not of much importance. While it may perhaps account for a certain amount of cloud in depressions, frictional convergence is entirely inadequate to explain any but the lightest rainfall, so that other causes of vertical motion must be sought.

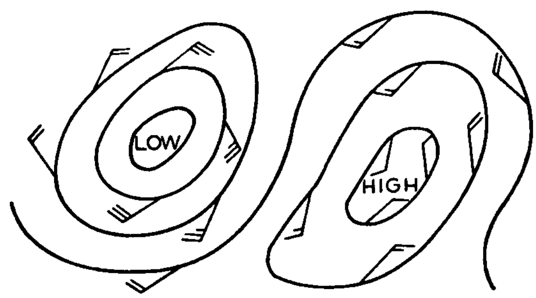

FIG. 36. *Convergence in a low and divergence from a high due to surface winds*

Divergence and convergence due to changing pressure. The geostrophic wind results from a state of balance between the pressure gradient and Coriolis force, but this balance is not achieved instantaneously. Consider air which is accelerated from rest by the pressure gradient resulting from a local fall of pressure; the Coriolis force, being proportional to the wind speed, is small at first and the deviating effect slight, so that in the early stages the wind is directed across the isobars towards the region of falling pressure. If the rate of fall of pressure is maintained, the lag in the adjustment of wind to pressure gradient continues and an ageostrophic component continues to flow towards the region of falling pressure. Similarly wind may be shown to have a component directed outwards from an area of rising pressure. The speed of these ageostrophic components not uncommonly amounts to about 10 knots; also, unlike divergence or convergence due to surface friction, they may affect a great depth of atmosphere. The upward motion associated with such convergence is sufficient to produce thick masses of cloud and continuous precipitation but even so the magnitude of the vertical currents, which are spread over a large area, is far too small to have any noticeable direct effect on the flight of an aircraft.

28. WIND AND HEAT TRANSFERENCE

The processes by which heat is first absorbed in the atmosphere, then redistributed horizontally and vertically and finally reradiated out to space are of considerable theoretical interest and are intimately related both to the day-to-day motion of the atmosphere and to the general circulation. The radiative balance of the atmosphere as a whole was discussed in Section 13; on average the atmosphere neither gains nor loses heat, and since the only means of heat exchange with outer space is by radiation, the absorption of short-wave radiation from the sun must be exactly balanced by the long-wave radiation escaping from the earth's surface and atmosphere. We have seen how the absorption of solar radiation takes place for the most part at the earth's surface; on account of the smaller declination of the sun in low latitudes, much more radiation is absorbed per unit area of the earth's surface in low latitudes than in high, when averaged over the year. Further investigation shows that the outgoing terrestrial radiation is only a little greater in low than in high latitudes. Consequently between the equator and about latitude 35 degrees more radiation is absorbed than emitted, while in higher latitudes there is on average a net loss of heat by radiation.

Thus there is a source of heat near the surface in low latitudes but the 'sink', or area from which most of the heat is lost, is found in the polar regions. In order that a steady distribution of temperature should be maintained, the excess heat must somehow be conveyed from the tropical source to the polar sink. This transference is brought about mainly by wind, that is to say by advection. Heat is transported polewards by streams of relatively warm air, but these displacements must be compensated by the movement equatorwards of air from high latitudes; hence in middle latitudes both warm and cold masses of air are brought into proximity and their interaction creates the variable cyclonic and anticyclonic circulations which are a characteristic feature of those regions. Another effect of wind is to generate ocean currents and these also make an appreciable contribution to the northward flux of heat.

Apart from the effects of radiation, exchanges of heat in the vertical direction are brought about directly or indirectly by means of vertical currents. In this respect the effect of molecular motions is inappreciable except for the transfer of heat between the earth's surface and air actually in contact with it, i.e. by conduction. A more effective agent for the vertical transfer of heat is turbulence and even more effective, when conditions are suitable, is convection. The turbulent transfer of heat, which also operates horizontally, is nevertheless a small-scale process seldom appreciably affecting layers of air more than a few thousand feet in thickness; on the other hand convection may affect the whole depth of the troposphere. Both horizontal and vertical currents also produce a redistribution of heat by the transport of water vapour; thus latent heat absorbed by evaporation at the surface in one place is usually released by condensation in the upper air at some other place. The whole process of the redistribution of heat in the atmosphere, including the exchanges between the atmosphere and the earth's surface, is extremely complicated but in many ways of fundamental importance for the understanding of weather (see Section 13 and Fig. 7).

29. EQUIVALENT HEADWINDS AND TAILWINDS

In order that the time of flight of an aircraft on a given route may be determined, the ground speed must be known at each point and this depends on the wind components both along and at right angles to the track. The amount by which the airspeed exceeds the ground speed at any point of a track is known as the equivalent headwind at that point. When this conception is applied to the route as a whole it leads to the following definition:

> The equivalent headwind over a route is defined as that constant wind which, blowing along the track at all points, results in the same average ground speed as that produced by the actual distribution of wind at the time of flight.

The wind information for the route when expressed in this way is thus reduced to a single figure, as far as the average speed and time of flight are affected. A negative value of the equivalent headwind is to be interpreted as an equivalent tailwind. It should be noted that the result depends on the airspeed and so to some extent on the aircraft concerned.

The above ideas lead to great simplification in preparing statistics for a given route when approximately the same track is to be flown on each occasion. From the frequency distribution of wind strength and direction at points of the route in the neighbourhood of the operating height, it is possible to deduce the frequency distribution of equivalent headwinds, whence a table or diagram may be prepared showing for example how often any particular value of an equivalent headwind or tailwind is likely to be exceeded in a given period. Such a diagram is illustrated in Fig. 37 for the route London to New York in the winter months. An operator who

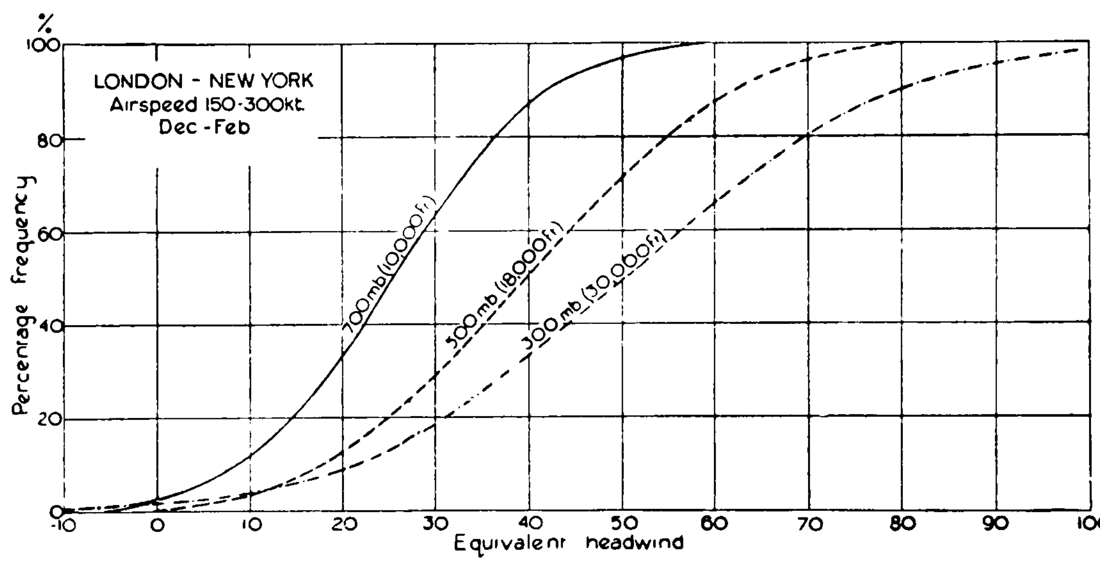

FIG. 37. *Curves of frequencies of equivalent headwinds less than any given value*

London to New York, December–February (airspeed: 150–300 knots).

requires to provide a service on this route with a regularity of, say, 85 per cent can then see at once from the diagram that he will have to reckon with effective headwind components up to 40 knots if flying at 700 millibars, 60 knots at 500 millibars and 75 knots at 300 millibars. A similar diagram prepared for the reverse direction of flight will differ in detail since the effect of a wind component at right angles to the track is always adverse. Thus on this route in the winter season the mean equivalent headwind for westbound flights at 18 000 feet is 40 knots, but the mean equivalent tailwind for eastbound flights is only 34 knots.

Frequencies of equivalent headwinds are prepared on the assumption of a particular value for the airspeed. This is conveniently taken as 200 knots for piston-engined aircraft but it is an advantage of the method that the results are not closely dependent on the assumed airspeed, so that values derived for 200 knots may be applied with little error to the range from about 150 to 300 knots. For jet aircraft an appropriate basic figure would be 400 knots but the results can be easily derived from figures calculated for 200 knots.

30. PRESSURE-PATTERN FLYING

There are several navigational techniques which make use of the distribution of pressure in order to determine the effect of wind on an aircraft. Pressure-pattern flying is the name given to these techniques; they may be grouped into three main categories, concerned respectively with:

(i) altimetric determination of drift,
(ii) single-heading flight, and
(iii) economy of time and fuel.

For the detailed application of these techniques, reference must be made to other publications; the following sections are concerned mainly with the meteorological aspect.

Determination of drift from altimetric observations

Drift is defined as the angle between the heading of an aircraft and its track. In Fig. 38, if the aircraft's heading is in the direction PN while the track made good is along PQ then there is a drift to port given by the angle NPQ or α. If NQ

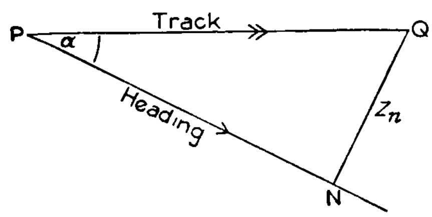

FIG. 38. *Drift* (α) *and displacement* (Z_n) *of an aircraft*

is drawn perpendicular to PN, the component of wind at right angles to the heading blows parallel to NQ; the distance NQ or Z_n is the total displacement of the aircraft perpendicular to its heading (assumed constant) during the passage from P to Q and is given by

$$Z_n = \frac{g}{2\Omega \sin \phi} \cdot \frac{H_Q - H_P}{A} \quad \text{(see Appendix I, p. 389)},$$

where H_Q and H_P denote the true height of the aircraft at Q and P respectively and A is the true airspeed of the aircraft. The altimeter correction D at any point (the subscale being set to 1013·2 mb) is the difference between the true height at that point and the pressure altitude h (Section 8); the change in the altimeter corrections between P and Q is therefore

$$D_Q - D_P = (H_Q - h) - (H_P - h) = H_Q - H_P.$$

If the altimeter corrections are in feet, A in knots and Z_n in nautical miles then

$$Z_n = \frac{21\cdot 47}{\sin\phi} \cdot \frac{D_Q - D_P}{A} \qquad \ldots . (14)$$

and

$$\sin\alpha = \frac{21\cdot 47}{\sin\phi} \cdot \frac{D_Q - D_P}{As} \qquad \ldots . (15)$$

where s is the ground displacement in nautical miles between P and Q.

The direction of the drift may be readily inferred from the geostrophic relationship and can be stated explicitly as follows:

If the value of D is increasing, the drift is to port; if D is decreasing the drift is to starboard. This rule is reversed in the southern hemisphere.

The altimetric drift obtained in this way is subject to some uncertainty on account of instrumental and observational errors and the assumption that the wind is exactly geostrophic. The last assumption is unlikely to lead to serious error over a long flight since the geostrophic departure would be expected to make several changes of sign *en route* and the total displacement on this account would be small. Provided certain further restrictions are introduced, namely that airspeed, latitude and heading may be regarded as constant throughout the flight, equations (14) and (15) may be shown to hold however long the flight path may be. These equations, in which ϕ should be regarded as the mean latitude, then give the total displacement and drift in terms of the altimeter corrections at the terminals.

Pressure-pattern position lines

In Fig. 39 the point A represents the position on a navigator's chart for which a reliable fix has been obtained together with the altimeter correction D_A. The line ABC represents the aircraft's heading, further observations of D being made at, say, half-hourly intervals as represented by the points B, C... From the observations D_A and D_B the displacement Z_n perpendicular to AB is calculated and the pressure-position line RR′ is drawn parallel to AB at distance Z_n to port or starboard as the

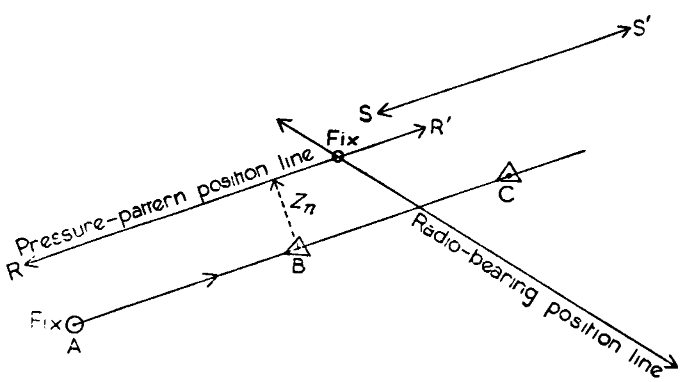

Fig. 39. *Pressure-pattern position line*

case may be. The position of the aircraft at the time of the observation D_B will be somewhere along RR′, depending on the wind component in this direction; a fix can be determined only by other navigational means, for example a simultaneous position line may be obtained from the bearing of a radio beacon. Subsequently, further pressure-position lines such as SS′ may be determined from the accumulated displacements between observations made at A and B, and at B and C, C and D, etc., provided the heading remains constant. A pressure-position line gives all the positional information which is obtainable from altimetric observations. The assumption of a geostrophic wind introduces a source of error into the result.

Single-heading flight

This refers to a flight made with a constant heading, the pressure altitude and airspeed also being approximately constant. The required constant heading is obtained by a direct application of the expression (14) which gives (under the conditions stated) the displacement Z_n in terms of the altimeter corrections at the two terminals. Suppose the terminals are represented on the navigator's chart by the points A, B, in Fig. 40. Join AB by a straight line, draw a circle centre B and radius

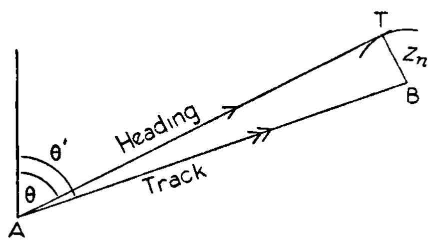

FIG. 40. *Single-heading flight as plotted on the navigator's chart*

Z_n (in terms of the scale of the chart), and draw AT tangential to this circle (to port or starboard according to the sign of Z_n); then AT gives the required constant course or heading with respect to a rectangular grid on the chart. If θ is the required bearing, θ' the bearing of the straight line AB, then

$$\sin(\theta' - \theta) = \frac{Z_n}{d} = \frac{g}{2\Omega \sin\phi} \cdot \frac{D_B - D_A}{Ad}, \qquad \text{. . . . (16)}$$

where d is the distance between A and B as measured along the straight line on the chart.

It must be emphasized that the heading which is maintained constant is measured from a system of parallel lines on the navigator's chart and is not necessarily measured from true (or magnetic) north. Only if the chart is on a Mercator projection will the course be in a constant geographical direction. Fig. 41 illustrates the importance of the reference grid for single-heading flight. The projection of the chart is such that the great circle from London to New York appears as a straight line (a); the rhumb line, that is the line with a constant geographical bearing is the curved line (c). Two tracks for a single-heading flight on a particular occasion have been calculated on the respective assumptions that the aircraft maintained (i) a constant course with respect to a rectangular grid superimposed on Fig. 41 and (ii) a constant geographical course, i.e. with respect to a rectangular grid on a Mercator chart. The two tracks, shown as (b) and (d), are seen to diverge widely from each other although they are somewhat similarly disposed to the great circle and rhumb line respectively.

An advantage of single-heading flight is that the course to be flown can be calculated from values of upper air pressure at the terminals alone, without knowledge of the winds that will be encountered *en route*. For this purpose, forecast values of

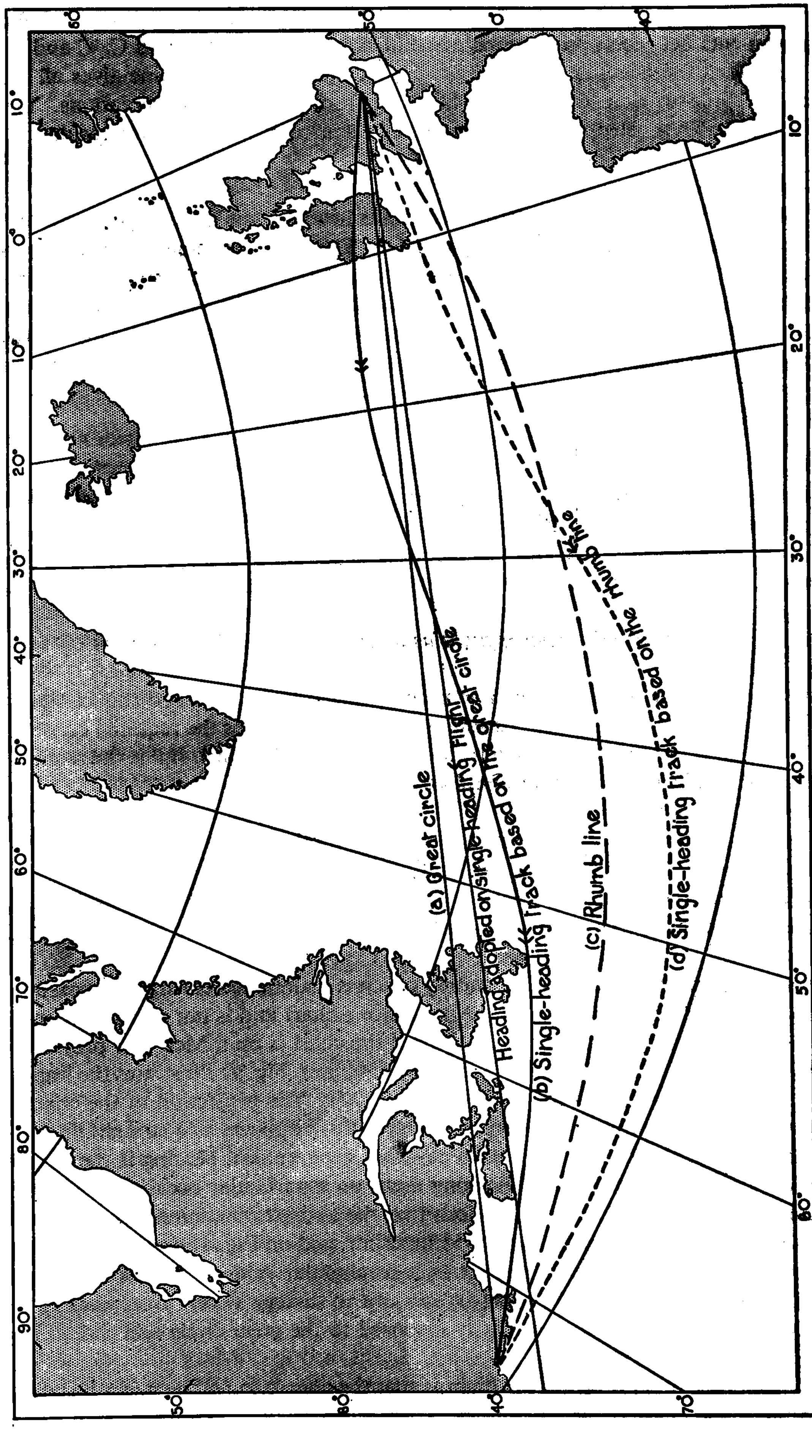

FIG. 41. *Relation of single-heading flight path to the 'map' straight line on which it is based*

D_A and D_B may be used; alternatively if the flight is made on one of the standard meteorological pressure surfaces, $D_B - D_A$ is replaced by $H_B - H_A$, the values of H_A and H_B being derived from a forecast contour chart. These procedures clearly do not require any special equipment to be carried on the aircraft.

In most cases, a single-heading flight gives a saving in time and fuel compared with the direct flight following the straight line on the map used for the computation. However, single-heading flight does not usually provide the maximum saving which can be effected by careful selection of route and it may take even longer than the great-circle route. Referring to Fig. 41, flight along the track (b) might be expected to be quicker than that along (a), and along (d) quicker than along the rhumb line (c), but there is no reason to expect flight along (d) to be quicker than along the great circle (a).

The actual track to be followed on a single-heading flight can be predetermined only when the winds are known along the route; when there are strong beam winds it may vary widely from the straight-line track on which it is based. For this reason a modified procedure may be preferred, for example the route may be divided into two or more legs and a separate single-heading course flown on each leg.

Minimum-time tracks

The time of flight of a given aircraft between two terminals varies to a considerable extent with the winds encountered. On long flights especially it becomes desirable to avoid strong headwinds; a route can usually be found which, although geographically longer than the direct route, can be flown in less time. While there are many factors other than wind which together determine the selection of a suitable track on any given occasion, it is usually practicable to effect an appreciable saving of time by careful choice of track. Once the winds over an extensive area covering the route have been forecast, the computation of the quickest track is a navigational matter and beyond the scope of this handbook. Computations of minimum-time tracks for the route London–New York show that they often vary widely from the great circle and change from day to day according to the winds. Such variations are very inconvenient operationally. An alternative procedure (which approximates to the desire for minimum time of flight) is to lay down several fixed tracks for a given route and to select the most appropriate one for any particular occasion, having regard not only to time of flight but also to other factors such as avoidance of bad weather. The most suitable track of the set will usually be apparent from inspection of the current and forecast weather charts; it will avoid adverse headwinds as far as possible and will take account of the variation of wind in the vertical as well as horizontally. In cases of doubt the times of flight of two or more routes can be computed. Once the most suitable of the standard tracks has been selected, the flight may if desired be flown with the single-heading technique applied to each leg.

It should be mentioned that an aircraft's performance and its most economical cruising speed both vary considerably with the altitude at which the flight is made. On this account the quickest route and operating height do not necessarily coincide with that method of flying the route which will effect the greatest economy in fuel consumption.

CHAPTER 6

FORMATION OF CLOUD AND PRECIPITATION

31. WATER IN THE ATMOSPHERE

The water vapour of the atmosphere must be transformed into water drops or ice crystals before cloud can appear, and the cloud particles must increase still further in size before they are able to fall out as precipitation. Before going on to describe the methods by which cloud and precipitation are formed from the vapour, it is necessary first to define more fully than has been done hitherto the terms used to specify the state of the atmosphere in respect of its water-vapour content, and to consider the changes of state between vapour, water and ice.

The dry air and water vapour composing the atmosphere may each be considered as separately satisfying the laws of gases. The total pressure is the sum of the partial pressures of these two constituents (Section 20). The partial pressure of the vapour, or more briefly the vapour pressure, is denoted by e and the total pressure as usual by p, so that the partial pressure of the dry air is given by $(p-e)$.

When dry air is confined in contact with a surface of water or ice, evaporation takes place and the vapour pressure increases. At any one temperature there is a limit to the amount of vapour that can be taken up in this way. When this limit is reached the air is said to be saturated and the vapour pressure has reached its maximum value, the saturation vapour pressure e_S for that particular temperature. If the vapour pressure is less than the saturation vapour pressure the air is unsaturated.

Apart from vapour pressure there are several methods in common use of specifying the state of the atmosphere in regard to its vapour content, and the more important of these are described below.

Humidity mixing ratio (r) (see Appendix I) is the mass of vapour contained in unit mass of dry air. It is usually expressed as the number of grammes of vapour per kilogramme of dry air (g/kg). Isopleths of the humidity mixing ratio for saturated air are shown as pecked lines on the tephigram (Appendix II).

Relative humidity is the percentage degree of saturation. It may be defined as 100 times the ratio of the actual vapour pressure to the saturation vapour pressure at the same temperature, or $100\ e/e_S$.

Vapour concentration or *absolute humidity* is the mass of vapour contained in unit volume. It is usually expressed as the number of grammes of vapour per cubic metre (g/m^3).

Dew-point is the temperature to which moist air must be cooled in order to just reach the condition of saturation with respect to a plane water surface. Further cooling results in condensation on solid surfaces. Even slight cooling beyond the dew-point will ensure condensation on dust particles in the air, forming fog or cloud. Fog may form even in unsaturated air, as explained later, and its presence should not be taken as indicative that the temperature is at or below the dew-point.

Frost-point is the temperature to which moist air must be cooled in order to just reach the condition of saturation with respect to a plane ice surface. Further cooling induces deposition of ice in the form of hoar frost on solid surfaces including other ice surfaces.

Wet-bulb temperature is measured with a wet-bulb thermometer (Section 82); in conjunction with the dry-bulb temperature it forms the standard method of measuring humidity at the earth's surface. As heat is absorbed in the process of evaporation, the wet-bulb becomes cooled below the air temperature unless the air is already saturated. The wet-bulb temperature is the lowest temperature to which air may be cooled by evaporation of water.

Water vapour can change directly into ice, a process known as sublimation, and conversely ice evaporates directly into vapour. Moreover, water droplets in suspension readily remain liquid or supercooled at temperatures below 0°C. Consequently at such temperatures it is necessary to consider the saturation vapour pressure not only with respect to ice but also with respect to supercooled water. It is found that the saturation vapour pressure over ice is slightly less than that over supercooled water at the same temperature (see Table 5); hence if unsaturated air is cooled progressively below 0°C, the frost-point is reached while the air is still unsaturated with respect to water.

TABLE 5. *Saturation vapour pressure* (*SVP*) *over water and ice*

Temperature (°C)	−60	−50	−40	−30	−20	−10	0	10	20	30	40
SVP over water (mb)	—	0·06	0·19	0·51	1·25	2·86	6·11	12·3	23·4	42·4	73·8
SVP over ice (mb)	0·01	0·04	0·13	0·38	1·03	2·60	6·11	—	—	—	—

From Table 5 it is seen that the saturation vapour pressure increases rapidly with temperature so that warm air, when saturated, holds more water vapour than cold air.

The amount of vapour present in a given mass of air can be varied only by means of evaporation or condensation and (at a given pressure) does not depend on the temperature unless this is reduced below the dew-point, when the excess vapour is condensed. In contrast, the relative humidity varies widely with temperature. A rise of temperature implies a rise in the saturation vapour pressure and consequently a fall in relative humidity, although this may be partly offset by evaporation from the surface or from vegetation; conversely a fall of temperature increases the relative humidity unless the air is already saturated. The diurnal variation of temperature over land is accordingly reflected in a diurnal variation of relative humidity, the lowest values occurring at the warmest time of the day and the highest at night when, if 100 per cent is approached, fog can readily form. Even in desert regions the relative humidity at the surface may reach 100 per cent at night due to intense radiative cooling, so giving heavy dew or even fog. The lowest values of relative humidity occur in air subjected to prolonged heating such as might result from insolation over dry land, from advection over a warm dry surface or from adiabatic subsidence; the highest values occur during prolonged evaporation from the surface — as when air is in contact with the sea or wet ground — or as a result of cooling by radiation, advection or adiabatic ascent. On account of the prevalence of convection, air in the upper troposphere is frequently near saturation. In the lower stratosphere extremely low relative humidities are the rule; in temperate latitudes the average is about 2 per cent and it rarely exceeds 10 per cent.

Condensation, sublimation and freezing

When the air temperature is reduced below the dew-point, water droplets condense on minute particles suspended in the air which act as condensation nuclei. Without such particles the vapour pressure would surpass the saturation value without condensation, that is the vapour would become supersaturated, but in practice suitable nuclei are invariably present in large numbers so that supersaturation with respect to water is not likely to be large. As the condensation nuclei are often hygroscopic, that is, having a special affinity for water, they may bring about condensation even before saturation is reached; this explains the occurrence of mist and fog with relative humidity below 100 per cent. The main origin of the nuclei is probably the combustion products of domestic, factory and other fires. Sea salt particles probably contribute about one-tenth of the nuclei involved in cloud droplet formation.

When the vapour is cooled below the frost-point, it may be that ice crystals are formed on sublimation nuclei. Laboratory evidence however indicates that condensation does not take place until after the vapour has become saturated with respect to water and the first product is then a supercooled droplet; subsequently the droplet may freeze spontaneously if it contains a solid nucleus which is active as a 'freezing nucleus', presumably because its shape is similar to that of an ice crystal. In any case, the rarity, if not the complete absence, of sublimation nuclei in natural conditions means that water vapour on cooling readily becomes supersaturated with respect to ice while still unsaturated with respect to water, a condition which is of common occurence. If on the other hand ice crystals are already present, having perhaps fallen from higher levels, and if the vapour pressure exceeds the saturation value with respect to ice, then direct sublimation takes place on to these ice crystals.

Supercooled drops are in an unstable state and usually start to freeze when brought into contact with ice particles or other objects; this aspect will be considered further in relation to the formation of ice on aircraft (Chapter 8). When freezing of supercooled droplets takes place spontaneously, the larger droplets tend to freeze more readily than the smaller ones. Generally, the freezing of droplets becomes increasingly probable as the temperature continues to fall. In cloud a few frozen particles may be present at a temperature of —10°C. At still lower temperatures the number of ice particles increases and at —32°C there is a further marked increase, but it is not until the temperature falls to about —40°C that freezing is likely to become general. Thus cloud at temperatures between 0°C and —10°C may be considered as consisting almost entirely of supercooled drops; between —10°C and —40°C there are both supercooled drops and ice particles, the latter being more numerous the lower the temperature; while cloud at temperatures below —40°C consists mainly or entirely of ice crystals, giving the cirrus types, although there are exceptional cases of water-drop clouds existing even below —40°C.

Clouds with strong upcurrents often contain a relatively high proportion of water drops which are carried up before they have an opportunity to freeze, but the ice crystals grow more rapidly than the water drops because adiabatic cooling of the ascending air maintains a state of saturation with respect to water and therefore supersaturation with respect to ice.

32. GENERAL CAUSES OF CLOUD AND PRECIPITATION

A cloud is formed by the condensation of water vapour into droplets, or occasionally into ice crystals. The immediate cause of condensation to water drops is the reduction of air temperature below the dew-point. In nature this may be effected in any of the following ways:

(i) loss of heat by conduction to a cold surface,

(ii) loss of heat by radiation from the air, or

(iii) adiabatic cooling due to ascent.

The first process may result only in a deposition of dew on the cold surface (or of hoar frost if the temperature is below freezing-point) but if the air near the surface is subject to slight turbulent mixing the cooling is spread upwards and temperature may fall below the dew-point through an appreciable depth. Condensation then takes place within the air itself and a cloud is formed resting on the surface. This type of cloud is known as mist or fog according to its opacity and will be dealt with in Chapter 9 together with other causes of reduced visibility.

Direct cooling by radiation from the air itself, or rather from the water vapour contained in the air, process (ii), may modify the formation of fog and cloud, but its precise importance is not easy to estimate. When a fog or cloud has formed, radiation from its upper surface may affect its further development but it is doubtful whether radiation from the air is ever the sole cause of condensation.

Apart, therefore, from fog due to surface cooling, the methods of cloud formation reduce to only one, adiabatic cooling by reduction of pressure, which is almost entirely associated with vertical motion. The different types of cloud are then to be accounted for by the various modes of upward motion which, in accordance with the discussion in Section 27, may be classified as follows:

(i) turbulence,

(ii) orographic ascent,

(iii) convection, and

(iv) slow widespread ascent.

Two or more of these effects may occur simultaneously, giving a wide variety of cloud types which do not fall definitely into any simple class.

Condensation level

When cloud is formed by adiabatic cooling, the process can be followed on the tephigram as illustrated in Fig. 42. Air with a temperature of 14°C is represented by the point T, while the dew-point temperature, 8°C is marked as D on the same isobar. The pecked line through D indicates the humidity mixing ratio of the air represented by T. If a sample of air from T ascends adiabatically, its temperature follows the dry adiabatic through T; meanwhile its vapour content – as given by the humidity mixing ratio – remains unchanged, so that the dew-point moves along the isopleth of humidity mixing ratio through D. The lines of constant humidity mixing ratio are for this reason known also as dew-point lines. Thus when the temperature of the ascending air is represented by T′, the corresponding dew-point is given by D′ on the same isobar. Where the dry adiabatic and dew-point lines intersect, the temperature and dew-point coincide, so that the air becomes saturated at the point C. This point determines the base of cloud formed in the ascending air and its height (or pressure) is termed the 'condensation level.' Beyond this point, any further ascent follows the saturated adiabatic CS.

If the wet-bulb temperature is marked as the point W, then the condensation level is also determined by the intersection of the saturated adiabatic through W with the dry adiabatic through T. This result follows from Normand's theorem which may be stated as follows:

On the tephigram, the dry adiabatic through the dry-bulb temperature, the saturated adiabatic through the wet-bulb temperature, and the dew-point line through the dew-point, all meet in a point.

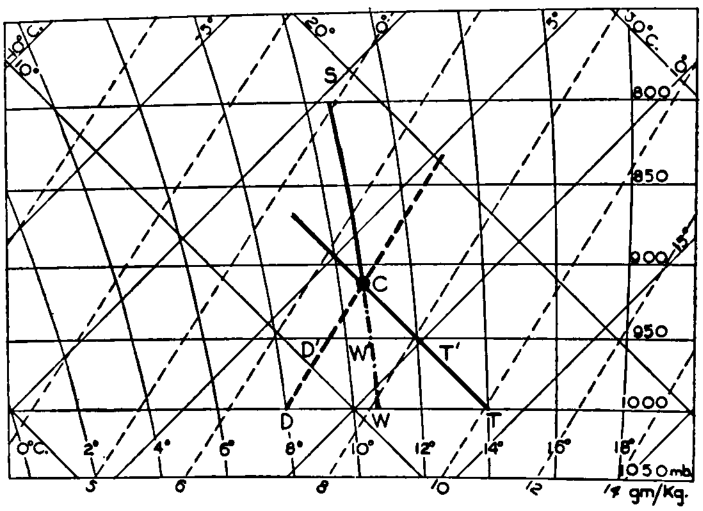

FIG. 42. *The condensation level of ascending air*

It follows from Normand's theorem that the ascent path which is followed above the condensation level by a sample of air of given temperature and humidity is determined solely by the wet-bulb temperature. Moreover, even in the unsaturated stage of the ascent, the wet-bulb temperature follows the same saturated adiabatic. The temperature at which the pressure on the saturated adiabatic becomes 1000 millibars is called the wet-bulb potential temperature by analogy with the potential temperature which was defined (Section 14) in relation to the dry adiabatics. It is seen that the wet-bulb potential temperature remains constant during any change which follows a saturated adiabatic. On the tephigram, the saturated adiabatics are labelled with values of wet-bulb potential temperature at 2 degC intervals.

Size and terminal velocity of drops

Cloud particles, as distinct from raindrops, have an average diameter of about 0·02 millimetres and their rate of fall in still air is about 0·012 metres per second. This rate of settlement is trivial and the particles practically float in the air or are carried upwards in the vertical currents which give rise to the cloud. Before precipitation can occur, some of the particles must increase in size until they become heavy enough to fall out of the cloud. A falling drop quickly attains a limiting velocity relative to the air, when the air resistance just balances its weight. This is known as the terminal velocity, some values of which are given in Table 6 for drops of various sizes.

The terminal velocity increases at first with the size of the drops but reaches a limit of about 9 metres per second for drops of about 5·5 millimetres diameter.

TABLE 6. *Terminal velocity of water drops in stagnant air at pressure 1013·2 mb and temperature 20°C*

Type of drop	Diameter of typical drop *mm*	Terminal velocity *m/s*
Cloud droplets	0·01	0·003
	0·02	0·012
	0·1	0·27
Drizzle drops	0·2	0·72
	0·4	1·62
	1·0	4·03
Raindrops	2·0	6·49
	4·0	8·83
	5·0	9·09
	5·8	9·17

The larger drops are deformed by the resistance of the air and if greater than 5·5 millimetres they become broken up into smaller drops. Hence there is an upper limit to both the size and the terminal velocity of raindrops.

There is no sharp demarcation between cloud particles and raindrops. Since vertical velocities due to convection not uncommonly exceed 9 metres per second, drops up to the largest size may be found in the body of a convection cloud. Generally the average size of drops in a cloud increases with the intensity of the upward motion.

Precipitation

The mechanism by which some of the particles in a cloud attain a size and terminal velocity great enough to enable them to fall out as rain or snow must now be considered. In many clouds no such process is effective, and since convection clouds often develop to great heights without giving precipitation, it is clear that continued cooling by adiabatic ascent, although essential, is not of itself sufficient to give rain in all cases. Drizzle drops have a diameter of about 0·2 millimetres; although they form readily from cloud particles, they can fall to the ground only when the upcurrents are very slight. The reason for the growth of cloud particles to the size of raindrops has presented much difficulty. Although the problem is still under investigation it is realized that adequate growth can be achieved only by collisions between cloud particles; moreover an essential factor for this purpose is the difference in the fall velocity of droplets of different sizes, so that the faster falling particle sweeps up the smaller particles in its path. Present theories suggest two main methods for the release of rain from a cloud.

Ice-particle theory of precipitation. The theory that release of precipitation requires the presence of ice crystals in the upper part of the cloud was advanced by Bergeron in 1935. From what has been said (page 80) about the freezing of droplets, it is seen that when the adiabatic reduction of temperature in the vertical currents is carried beyond 0°C the cloud particles become supercooled. If the cloud is of sufficient development, the region consisting wholly or mainly of supercooled drops extends over a height range of many thousands of feet above the level of the 0°C isotherm. It was seen that as lower temperatures are reached, an increased proportion of droplets is likely to become frozen, giving a mixture of supercooled drops and ice crystals. For reasons already explained, the ice crystals in this part of the cloud grow by sublimation and so acquire a fall velocity relative to the supercooled drops. They increase in size by collision with these droplets and subsequently by the overtaking of drops at temperatures above 0°C in the lower part of the cloud; eventually they fall out from the base of the cloud as snowflakes or raindrops according to the temperature.

If the ice-particle theory were true universally, it would imply that every raindrop originates as an ice crystal and that no precipitation (other than drizzle) can be released except from clouds extending above the level of the 0°C isotherm. This in turn implies a considerable vertical development on most occasions, especially in summer and in low latitudes; it explains the observation that precipitation in temperate latitudes in winter sometimes occurs from comparatively shallow clouds, whereas clouds of similar depth in summer would probably fail to give precipitation. There is little doubt that this theory correctly explains many cases of precipitation, but observations have nevertheless shown that precipitation (other than drizzle) can occur from clouds with temperatures entirely above 0°C.

Coalescence theory of precipitation. This theory aims primarily at explaining the occurrence of precipitation in clouds warmer than 0°C. As long as condensation continues in a cloud, there is a tendency for the droplets to become of nearly uniform size, but, as with the ice-particle theory, it is necessary that some larger drops should be present before rain can be released. While there is no precipitation, the concentration of liquid water in an upcurrent must increase with height and it is suggested that eventually some larger droplets are formed by the coalescence of neighbouring particles. With continued ascent, the smaller drops tend to evaporate on reaching the top of the cloud, where the vertical currents die out; meanwhile the larger drops sink back into freshly rising cloudy air where, by virtue of their relative fall velocity, they grow rapidly by collisions to raindrop size. These drops are then likely to be broken up by the air resistance and the process is repeated with the fragments so that a chain reaction is set up. Hence only a very small proportion of the original cloud particles needs to be of large size in order that precipitation should be released in this way, although it is clear that, for the chain reaction to be possible, the vertical currents must be strong enough to sustain large drops in the cloud.

Another theory is that a few unusually large droplets may form on especially favourable condensation nuclei such as are known to originate as droplets of sea spray and so would be present to a significant extent in air that had recently been over the sea. Once taken up into the cloud, these larger droplets would start a chain reaction as in the coalescence theory.

33. CLOUD FORMED BY TURBULENCE

Formation of layer cloud by turbulent mixing

The prevalence of frictional turbulence in the first few thousand feet above the earth's surface and the effect of turbulent mixing on the lapse rate were discussed in Section 25. It is now necessary to consider the distribution of humidity. The mixing process tends to even out the water vapour content and to produce a condition in which the mass of vapour in each unit mass of dry air, that is the humidity mixing ratio, becomes equalized throughout the friction layer. Now suppose the lapse rate from the surface upwards is represented on some occasion by the line ABC on the tephigram, Fig. 43, while the dew-point curve is shown by A′B′C′. The effect of vigorous mixing of this air between 1000 and 900 millibars will be to make the lapse rate change towards the dry adiabatic represented by the line JL placed so that the mean temperature of the layer is unaltered. Another effect will be the tendency to produce a constant humidity mixing ratio throughout the layer; thus the original dew-point curve A′B′ will tend towards J′L′ located so that the average mixing ratio for the layer is unaltered. If this line intersects the dry adiabatic at a point K within the mixing layer, then saturation occurs at this point and a cloud layer extends from this level to the top of the mixing layer, the lapse rate in the cloud approximating to the saturated adiabatic KM.

The lowest point K at which condensation occurs is known as the mixing condensation level. It is in general different from the 'condensation level' (as defined in Section 32) of the original undisturbed surface air, although after mixing has taken place, the two condensation levels become identical. Reference to Fig. 43 shows that the tendency towards an adiabatic lapse rate is accompanied by a rise of temperature in the lower part and a fall in the upper part of the mixing layer.

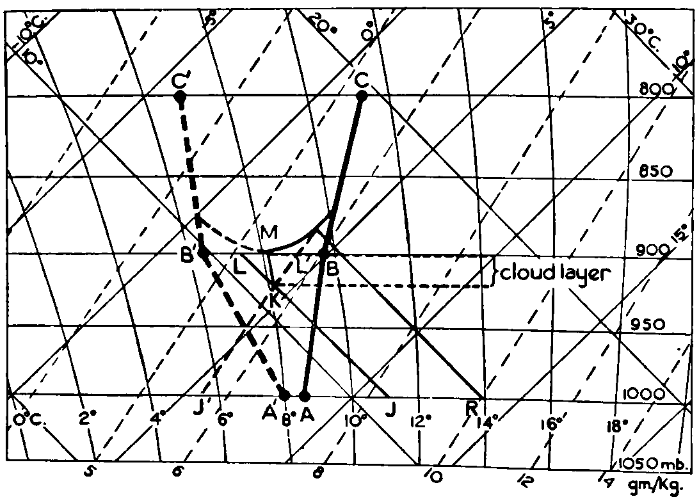

FIG. 43. *Effect of turbulence on lapse rate and humidity*

After mixing, the temperature curve ABC becomes JKMC, and the dew-point curve A′B′C′ becomes J′KMC′.

The top of this layer therefore coincides – at least in typical cases – with the base of an inversion which by its stability limits any further upward spread of turbulence. Consequently the top of the cloud layer extends only a little above the inversion base. Often both upper and lower surfaces of the cloud are undulatory while sometimes breaks are seen, owing to cloud being formed in the turbulent upcurrents and evaporated in the downcurrents. The cloud type is then stratocumulus, some photographs of which are reproduced at Plates IX and X (Chapter 12). At other times, usually when the wind is quite light, no structure is apparent in the cloud, which is then classified as stratus. Both types of cloud may also be formed by processes other than frictional turbulence, although vertical mixing is never entirely absent. Thus when air is subjected to gentle surface heating, thermal turbulence may be the sole factor, but more often thermal and frictional turbulence reinforce each other. With rather more active heating and an inversion aloft, stratocumulus may be formed by the spreading out of cumulus, a development which is considered in the next section.

The formation of turbulence cloud – stratocumulus and stratus – is further illustrated in Fig. 44. It is clear that certain conditions are required for its formation by turbulent mixing:

(i) Turbulence must be sufficiently active to steepen the lapse rate in the friction layer so that it tends towards the dry adiabatic.
(ii) Humidity must be great enough for the mixing condensation level to be reached within the friction layer.
(iii) The lapse rate of temperature above the friction layer must be stable.

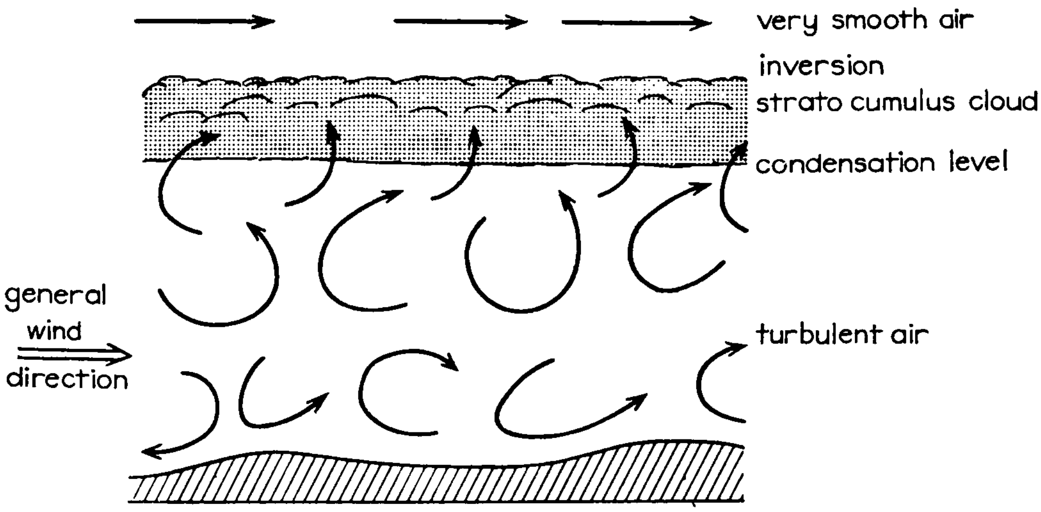

FIG. 44. *Formation of a layer of turbulence cloud*

If condition (i) is not fulfilled and if there is cooling by radiation or by advection over colder ground, then the lowest temperatures are likely to be found near the ground and fog rather than low cloud will result. If condition (ii) is not satisfied, then no cloud is formed. In regard to (iii), it has been said that mixing results in an inversion or stable layer immediately above the friction layer. On the other hand, a pre-existing stable lapse rate is favourable to the formation of layer cloud for reasons which will be given presently. In contrast, an unstable lapse rate aloft would permit convective developments which are usually incompatible with the formation of layer cloud.

General characteristics of cloud formed by turbulence

Forming within the surface turbulent layers, the cloud is low. Fog is really only the extreme limit of turbulence cloud, so that the cloud base may vary from only slightly above the surface to possibly 3000 or 4000 feet, beyond which height turbulence from the ground rarely penetrates. The vertical thickness may be as much as 3000 feet but is usually less; there is, however, no lower limit to the thickness and thin wispy patches of cloud may occur forming or dissipating locally according to variations in temperature, humidity and turbulence. The cloud type varies from stratus fractus to a continuous layer of stratus or stratocumulus. The base is generally at a fairly uniform height but often undulating; at the greater heights the cloud may assume a semblance of structure with denser and thinner patches occurring with some regularity; the lower bases are often associated with the lighter winds.

The presence of an inversion above the cloud has been attributed to an effect of mixing in the friction layer but there may be other processes at work tending to produce or to intensify an inversion. Thus a pre-existing subsidence inversion (Section 27) at a suitable height confines the upward diffusion of water vapour to within the friction layer, so raising the humidity there and favouring cloud formation. Cooling of the upper surface of the cloud at night by emission of long-wave radiation also helps to maintain not only the inversion above, but the steep lapse-rate within the cloud. Being cut off from the supply of surface moisture, the air above the cloud often has a much lower relative humidity than that below; this effect too is intensified if the air above is subsiding.

For adequate frictional turbulence, the surface wind speed must exceed about 10 knots, but once formed the cloud may persist with less wind. Bumpiness is generally noticeable below and within the cloud, but except in strong winds it is not more than moderate; above the cloud layer flying conditions are invariably smooth.

As the clouds are continually undergoing formation in ascending currents and dissolving in descent and have little vertical extent they rarely pass beyond the cloud stage to that of falling precipitation – they provide the phenomenon of dull overcast skies without rain. When the cloud is dense, drizzle or light rain or snow may fall, but if precipitation is considerable it may be assumed with confidence to be falling through the turbulence cloud from a higher level.

The effect of a cloud canopy in reducing loss or gain of heat by the surface has already been mentioned. With low thick clouds there may be little diurnal variation of temperature.

Some situations giving rise to turbulence cloud

One necessary condition for cloud to form is that the relative humidity of the surface layers shall be sufficiently high. This is favoured by prolonged surface cooling; turbulence cloud is therefore very common when air originating over sub-tropical oceans arrives in temperate latitudes. In the neighbourhood of the British Isles moist south-westerly winds frequently bring a continuous layer of very low cloud; over the western approaches this may occur at any season of the year, but in summer over the land it usually clears during the day-time by surface heating. If the wind is very light, fog is the more probable. In this type the origin of the air and the surface cooling are important factors.

Prolonged surface cooling is also characteristic of the air in anticyclonic regions over the continents in winter-time when, if turbulence is set up by an increase of wind or by surface heating as the air drifts over a warmer region, notably over the sea, a cloud layer is formed. The layer may be persistent and extensive, maintaining itself partly by the cooling of the upper surface of the cloud by radiation to the clear sky, while in industrial areas the accumulation of smoke particles in the cloud produces the well-known anticyclonic gloom of quiet weather in winter. Similar conditions may occur in other seasons if the air is sufficiently humid.

While certain occurrences of these clouds may be anticipated with confidence, there are numerous occasions when the formation demands a precise adjustment of humidity and turbulence and when a slight variation in one or the other may make all the difference between clear and overcast skies. A particularly difficult factor to allow for is the effect of the diurnal variation in turbulence, caused by the heating of the ground by day and cooling by night, combined with the associated variations in relative humidity. The variations in turbulence and in relative humidity are greatest in the evening and again in the early morning and it is at these times that turbulence clouds are most likely to form or to dissipate – but the nature of the change is not always the same. The heating after dawn tends to an increase in turbulence and a decrease in relative humidity; the former tends to the production of cloud, the latter to clearance, and which effect is the more important is a question of circumstances. Often following a clear night, a sudden formation of low cloud just after dawn appears to be attributable to an increase in turbulence; the cloud may clear later in the day, slowly or rapidly, as a result of further heating. If the cloud is dense the sun's heat may be unable to penetrate to the ground and the cloud may persist, a common occurrence in winter-time over the British Isles and the neighbouring continent. Even in summer many a potentially brilliant day is spoiled by an unexpectedly thick formation of low morning cloud which fails to clear.

On the other hand turbulence cloud may form at night after a clear day, as a result of cooling and an associated rise in relative humidity. This is particularly so with air coming from the sea; the relative humidity is then often high so that only slight cooling is required to produce cloud, while the increased turbulence over the land due to the greater roughness is another favourable factor. If dense, the cloud may persist through the night, but if thin and the wind is light, further surface cooling may stabilize the lapse rate so much that the formation either ceases or lowers to the surface as fog. Also turbulence cloud already existing over the sea may spread more or less rapidly inland some time in the evening, to be followed by dispersion the following morning. In all the circumstances the forecasting of the appearance and dispersal of the cloud becomes difficult and it may not be possible to go further than the issue of warnings of the possibility of sudden formations about dawn and towards sunset or later, on occasions when the humidity is high.

As a further type of turbulence cloud one may mention the very low ragged cloud – stratus fractus – which forms below a rain-bearing cloud such as nimbostratus, altostratus or cumulonimbus. The lower air is rendered almost saturated by falling rain, and turbulence easily causes low cloud formation within it.

In circumstances where the wind changes rapidly with height through a humid layer in the upper atmosphere, cloud sheets maintained by turbulence may be expected. Some cases of altocumulus or high stratocumulus may perhaps be explained in this way, although there is usually some other contributory factor giving rise to the humid layer.

Dissipation of turbulence cloud

Cloud formed by turbulence often dissipates over the land under the influence of insolation. If sufficient solar radiation penetrates to the ground and so raises the temperature of the air at the surface, then the condensation level rises so that the cloud thins out by a lifting of the base. For example, with conditions as in Fig. 43, the cloud would be expected to disperse completely if the surface temperature were to increase beyond the point R. Sometimes absorption of solar radiation at the upper surface may cause evaporation and consequent reduction in the height of the cloud top. Cloud which is present on a summer night often disperses entirely within a few hours of sunrise; at other times the cloud may persist all day, especially in winter, but even then a lifting of the cloud base during the day by perhaps 500 or even 1000 feet is usually noticeable. If the general conditions are unchanged, the cloud may be expected to re-form or to lower during the evening as already described.

34. FAIR-WEATHER CUMULUS

When the air is heated at the surface the thermal currents together with frictional turbulence mix the surface layers and the lapse rate tends towards the dry adiabatic; if the dew-point is reached in the cooler levels then cloud is formed, but the amount of cloud and the extent to which it develops upwards are matters of circumstances. If the lapse rate of the air through which the currents ascend is greater than the saturated adiabatic, the cloud is unstable and continues to develop upwards spontaneously until eventually halted in a stable layer; if the prevailing lapse rate is stable, development of the cloud is restricted from the outset. The processes governing the formation and development of convection cloud are advantageously followed with the aid of the tephigram.

The adiabatic ascent of air of given temperature and humidity was considered in Section 32; it is now necessary to consider such ascent in relation to the prevailing lapse rate. Suppose for example an early morning upper air ascent shows the temperatures and humidities indicated in Fig. 45. The lapse rate is stable, the air is unsaturated and no convection is present. During the day the surface temperature T increases while the dew-point D may be assumed to remain unchanged. When the surface temperature has increased to say T_1, a sample of air ascending adiabatically from the surface has its temperature reduced to that of the environment at the point P where the dry adiabatic through T_1 meets the observed temperature curve, and because the environmental lapse rate is stable the ascent comes to a stop at that level.

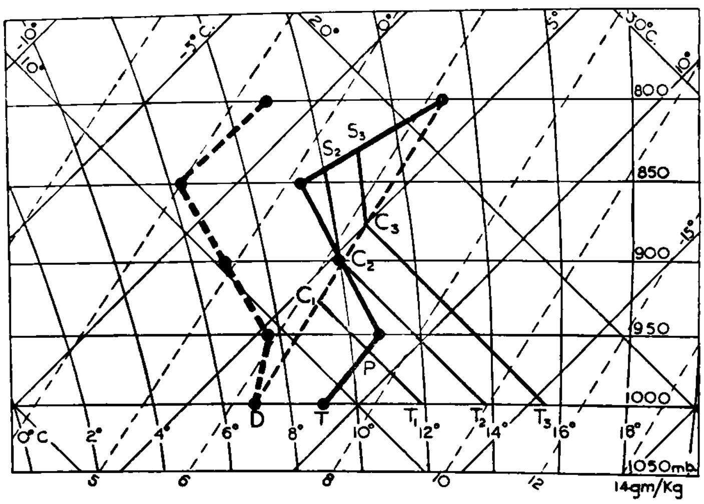

FIG. 45. *Formation of fair-weather cumulus*

As the condensation level C_1 (the intersection of the dry adiabatic through T_1 with the dew-point line through D) is not reached, no cloud is formed. Evidently the lowest surface temperature at which convection cloud can form is obtained by taking the lowest intersection C_2 of the dew-point line through D with the environment curve, and then drawing the dry adiabatic through C_2 to meet the surface pressure in T_2. Surface air heated to T_2 is then just able to rise to its condensation level. If the environmental lapse rate at this level is less than the saturated adiabatic, the ascent will be quickly arrested as it loses its momentum and any cloud formed will be shallow, not more than a few hundred feet thick. If, on the other hand, the lapse rate is steeper than the saturated adiabatic, as in the case illustrated, then further spontaneous ascent takes place along the saturated adiabatic through C_2 until the environment curve is again encountered at S_2. Thus the cloud base is formed at the level of the condensation point C_2 (900 millibars, approximately 2860 feet above the 1000-millibar level), and it extends upwards as far as S_2 (845 millibars, approximately 4500 feet), or possibly individual clouds will be carried up somewhat above this level by the momentum of the ascending currents. With a further rise of surface temperature from T_2 to T_3, both the base and top of the cloud are raised.

In discussing the formation of cloud by turbulence it was seen how mixing of the surface layers by friction tends to set up an adiabatic lapse rate and to spread the water vapour so that the upper part of the friction layer is cooled below the dew-point with the formation of cloud. If there is also thermal turbulence, the general character of the situation may be little altered except that any increase of surface temperature raises the condensation level and so increases the height of the cloud base. So long as the condensation level falls within the friction layer, a more or less continuous cloud layer is to be expected, but in other cases cloud can be formed only in currents penetrating above the friction layer, that is in the larger-scale thermal currents; these produce isolated clouds, giving the familiar broken sky with cumulus clouds drifting with the wind. The cloud bases are flat and clear-cut at the uniform condensation level, which is usually above 1500 feet, while the domed upper surfaces indicate the limit of penetration of the vertical currents. Sometimes when these currents are terminated by a marked inversion above the condensation level, conditions may resemble those for the formation of turbulence cloud; the cloud then spreads out beneath the inversion into a layer of stratocumulus, usually with light or clear patches in it.

Isolated convection clouds of limited vertical extent are referred to as 'fair-weather cumulus', the vertical development being inadequate for precipitation. The thermal currents are originated by surface heating due to insolation or advection, or a combination of the two. Air warmed by the ground beneath does not rise as a continuous stream but rather as a series of bubbles. Successive bubbles breaking away from the heat source at the ground combine with others to form larger bubbles. A series of bubbles starting in this way make up the convection currents (Fig. 46).

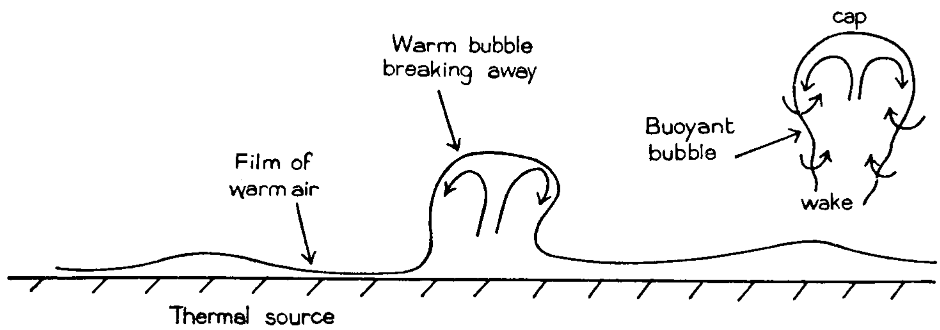

FIG. 46. *Development of thermal bubbles*

The upper and outer skin of the bubble mixes with the environment and part of this mixed air flows, relative to the bubble, down its sides. When condensation takes place the cloud droplets at the top and sides of the bubble mix with the drier environment and evaporate. The latent heat of evaporation is supplied by the air which therefore becomes cooler and denser and so sinks to form the outer edge to the wake of the bubble.

The convective bubble then consists of a rising dome of cloudy air and a wake, the outer part of which is sinking while the inner part is rising. The inner regions of the bubble continue to rise until the thermal is completely eroded or the buoyancy is counteracted by the increasing stability of the environment. Because of mixing with the comparatively cooler and drier environment, the rising bubble, or thermal, cools at a rate greater than the saturated adiabatic lapse rate and this leads to loss of buoyancy and to a brake on development.

As the cloud grows, a number of thermals develop in it and the vertical motion becomes more complex (Fig. 47). Cloudy thermals rising through, and entraining air from, the residue of previous thermals, cool at a rate greater than the saturated adiabatic but the loss of buoyancy is less than when the dry cool environmental air is entrained.

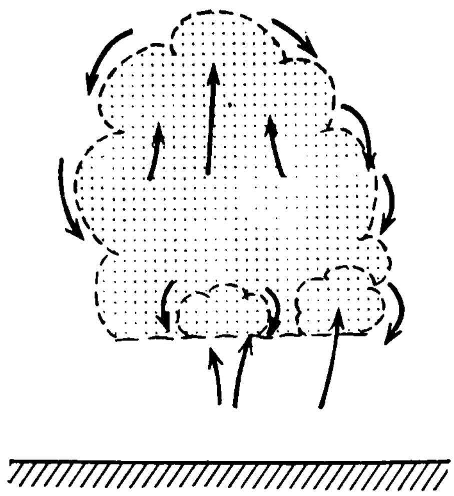

FIG. 47. *Cumulus cloud developing from thermals*

When formed over the land by diurnal heating alone, fair-weather cumulus is a phenomenon of the day-time only. It usually develops during the morning, reaches a maximum development in the afternoon and clears rapidly when the ground cools in the evening, but if the individual clouds have combined to form a more or less unbroken layer, the cloud may persist into the night especially if aided by frictional turbulence. When produced by advective heating, the cloud may also be present at night although the added effect of insolation enhances its development overland by day. Over the open sea only the advection type occurs, but because evaporation from and warming by the surface is continuous, the cloud often develops into stratocumulus when the lapse rate aloft is stable, or into intense convection cloud with showers when the lapse rate is deeply unstable.

In general, anticyclonic conditions are most favourable for fair-weather cumulus, particularly in summer. In winter anticyclones, thermal currents are less intense and are usually limited by a marked inversion; if cloud forms at all it is commonly a flat layer at a low height, more properly regarded as turbulence cloud than as convection cloud.

35. INSTABILITY CLOUDS AND SHOWERS

An instability cloud is a general term for a convective cloud with intense vertical development, and therefore for one liable to give precipitation in the form of showers. Small cumulus clouds of the type considered in the previous section are prevented from developing further by the stability of the lapse rate aloft, but if the lapse rate above the condensation level exceeds the saturated adiabatic through a deep layer, the cloud becomes correspondingly extensive; cooling less rapidly with height than its environment, it becomes continuously more buoyant and extends upwards with increasing speed until finally checked by stable conditions.

Entrainment of drier air round the edges of cumulus clouds restricts their development and as a result not all cloud tops reach the height anticipated from the tephigram. As the clouds increase in size the entrainment of environmental air has a proportionately decreasing effect and a stage is reached when the centre of the rising current cools more nearly at the saturated adiabatic lapse rate.

When instability is great enough some clouds develop sufficiently to produce precipitation. Once precipitation has started cloud turrets may be seen to grow rapidly and, in favourable circumstances, form towering masses of cumulonimbus which frequently reach the tropopause; occasionally severe storms extend some distance above the tropopause. In temperate latitudes convective cloud does not usually precipitate until the tops have penetrated some thousands of feet above the 0°C level. The base of the cloud may be at, say, 2000 feet and the vertical thickness may be as much as 30 000 feet or more. In rain the base may fall very low with ragged turbulence cloud almost to the surface.

Instability conditions when well developed present violent manifestations of energy. The speed of ascent often exceeds 9 metres per second (30 ft/s) and is sufficient to prevent even the largest raindrops from falling to the ground; in vigorous development of cumulonimbus cloud updraughts exceeding 50 metres per second are possible. The cloud, cooling with height approximately at the saturated adiabatic lapse rate, often extends well beyond the 0°C level and the topmost region may consist entirely of ice crystals. This part is therefore of cirrus type. Such extensive vertical development promotes the precipitation of rain (and often hail) according to the ice-particle theory, but in other cases – more frequent in low latitudes where the 0°C level is high – the clouds may precipitate, as we have seen, without reaching the ice stage.

On reaching a stable layer aloft, the cloud spreads out horizontally and forms the well-known anvil shape typical of thunder and shower clouds. In the process of spreading out, and in the subsequent decay of the whole cloud system when the energy has been expended, patches or layers of cloud may form at several levels and give a chaotic appearance to the sky.

The extreme development of instability cloud is the thunder cloud, which is described in the next chapter. The stages by which cumulus clouds develop into cumulonimbus are still being investigated, but some distinction needs to be made between cumulonimbus clouds which develop in a succession of massive towers giving isolated thunderstorms of short duration and well-organized cumulonimbus of severe thunderstorms which may be accompanied by large hail and possibly tornadoes and, moving across the country, persist for several hours. Tops of cumulonimbus in some of these severe storms have been known to exceed 60000 feet.

In the formation of instability cloud and showers it is necessary to recognize two more-or-less independent conditions, both of which are essential: first, the lapse rate must be greater than the saturated adiabatic through a deep layer; secondly, the air must become saturated, that is cloud-laden, for if the air is not saturated it does not become unstable until the much steeper lapse rate, the dry adiabatic, is reached. It has been explained (Section 16) that any lapse rate between the saturated and the dry adiabatics is conditionally unstable and that instability cannot be realized unless the air is saturated. Such a condition is of common occurrence, and the more important means by which the instability may become effective are described on pages 93 to 97.

Instability cloud and showers in low latitudes

Cumulonimbus often attains a greater height and intensity in the tropics than in temperate regions. The greater height is permitted by the high tropopause of low latitudes, while the greater intensity depends essentially on the higher temperature at the condensation level, the consequently higher water vapour content and the greater amount of energy released by the subsequent condensation. Developments up to 50 000 feet or more are not uncommon and tops have been known to exceed 60 000 feet. Often the cirrus plume takes the form of an extensive anvil, and this or other associated cirrus formations may persist in the upper troposphere and even up to the tropopause itself long after the rest of the cloud has disappeared.

In contrast to the greater height of the cumulonimbus tops, the height to which a convective cloud develops before giving precipitation is sometimes considerably less in the tropics than it is in temperate latitudes. Showers may be produced when the cloud tops reach less than 10 000 feet above the base and so are well beneath the 0°C level. Precipitation is then to be attributed to coalescence; this process is more likely in the tropics than elsewhere, since temperatures and dew-points are high and relatively large amounts of water are condensed in the rising air.

Instability due to surface heating

Given a sufficiently high humidity in the surface layers, heating as we have seen may result in fair-weather cumulus. If, however, a cloud turret penetrates into a region which is conditionally unstable it continues to rise and a larger convective system develops. This process may be followed on a tephigram in the manner similar to that explained for small cumulus cloud. The development is illustrated in Fig. 48 which shows the plot of a morning upper air sounding. It is seen that if the surface dew-point D remains unchanged, then no convection cloud can form until the temperature increases from T to at least T_1; the condensation level is reached at C (875 millibars) and further ascent follows the saturated adiabatic through C to S (580 millibars). In this example the cloud has its base at about 4000 feet and its top a little over 14 000 feet.

The simple theory just given takes no account of the possibility that momentum acquired by rising air may cause it to continue rising somewhat beyond the level given by S on Fig. 48. Neither is account taken of the compensating downcurrents which must be interspersed with the upcurrents, nor of the mixing which takes place between the convection currents and the environment. The presence of the downcurrents implies that the cloud forms in isolated masses, more or less extensive and numerous according to circumstances. Of these variations the most common is caused by mixing and entrainment which frequently limit the cloud tops to levels below those suggested by the tephigram construction in Fig. 48.

Instability is common in air which has undergone prolonged surface heating by travelling from a cold to a warmer part of the earth, the lapse rate tending to become steep throughout a great depth of air. If the humidity is moderately high, as when the air has been for a considerable time over the sea, the cumulus clouds are easily formed and develop readily to the shower stage. Surface heating may take place by insolation or by advection; both processes operate to give the frequent passing showers and bright intervals associated with north-westerly winds over the British Isles. Any precipitation leaves the air progressively drier downwind and showers develop less frequently over south-east England; north-westerly winds in summer often bring brilliant weather, but the lapse rate aloft remains steep and if cloud does form it is likely to develop to a great height, perhaps giving a thunderstorm. Over the land, instability showers caused by diurnal heating occur mostly in the day-time, being most frequent in the afternoon and clearing completely at night; when initiated by advective heating they may also occur at night.

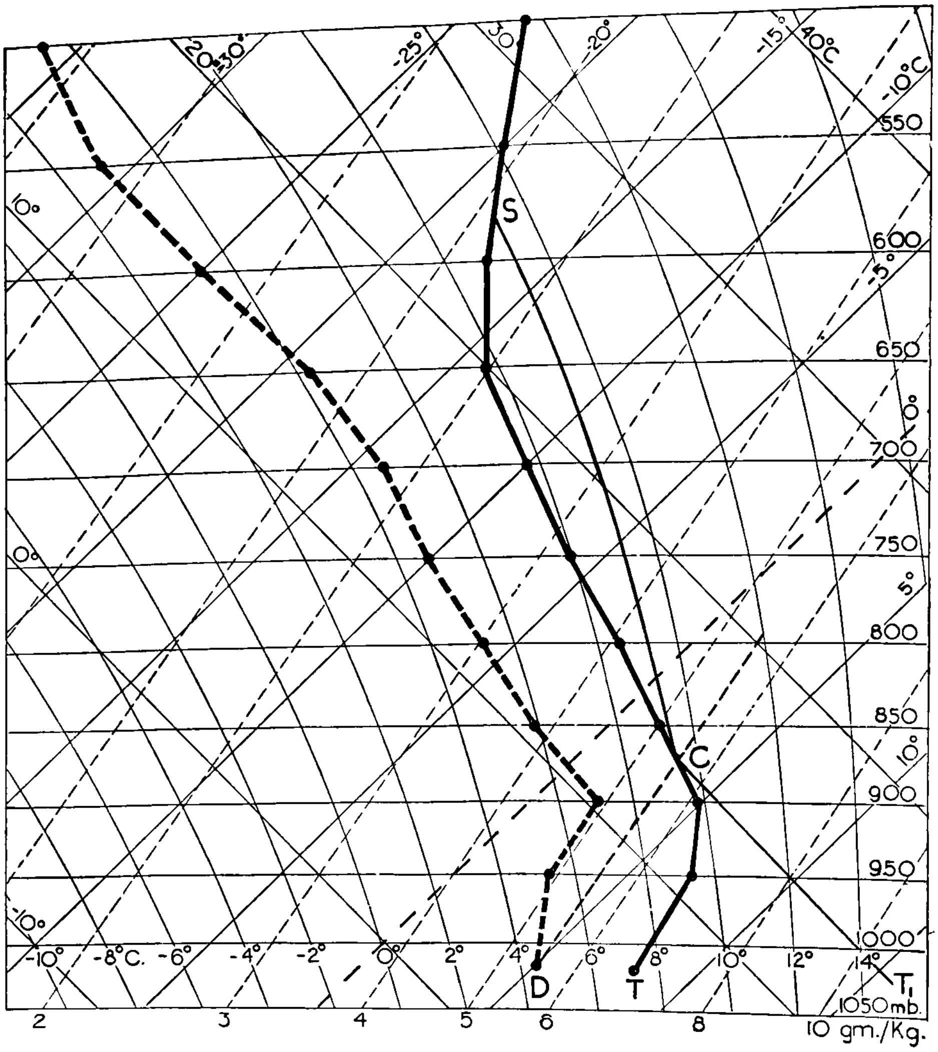

FIG. 48. *Development of instability cloud*

Latent instability

The lapse rate shown in Fig. 49 is seen to be conditionally unstable from 950 to 800 millibars but since the air is unsaturated at all heights, the situation is actually stable. If now a sample of air is by some means made to ascend from the surface, its temperature will follow first the dry adiabatic TC, and then the saturated adiabatic CQS, but not until the level of Q is passed does the rising air become warmer than its environment and so unstable. Any such lapse rate in which a parcel of air forced to ascend from any level eventually becomes unstable is said to possess 'latent' instability. Whether this property exists or not in any one case clearly depends both on the humidity of the lifted parcel and on the observed lapse rate, and it is seen that for instability to exist the saturated part of the ascent curve must at some stage lie to the right of the environment curve. From what has been said in Section 32, we derive the condition for latent instability: the saturated adiabatic through the wet-bulb temperature W must intersect the environment curve at some higher level.

It is a property of the tephigram that the area enclosed between the ascent curve and the observed temperature curve is proportional to the energy liberated or absorbed by an ascending parcel of air. In the case illustrated, energy must be supplied to move the parcel up through the stable region TQ, so that the area TCQT counts as negative. Above Q the air rises spontaneously to S and the energy

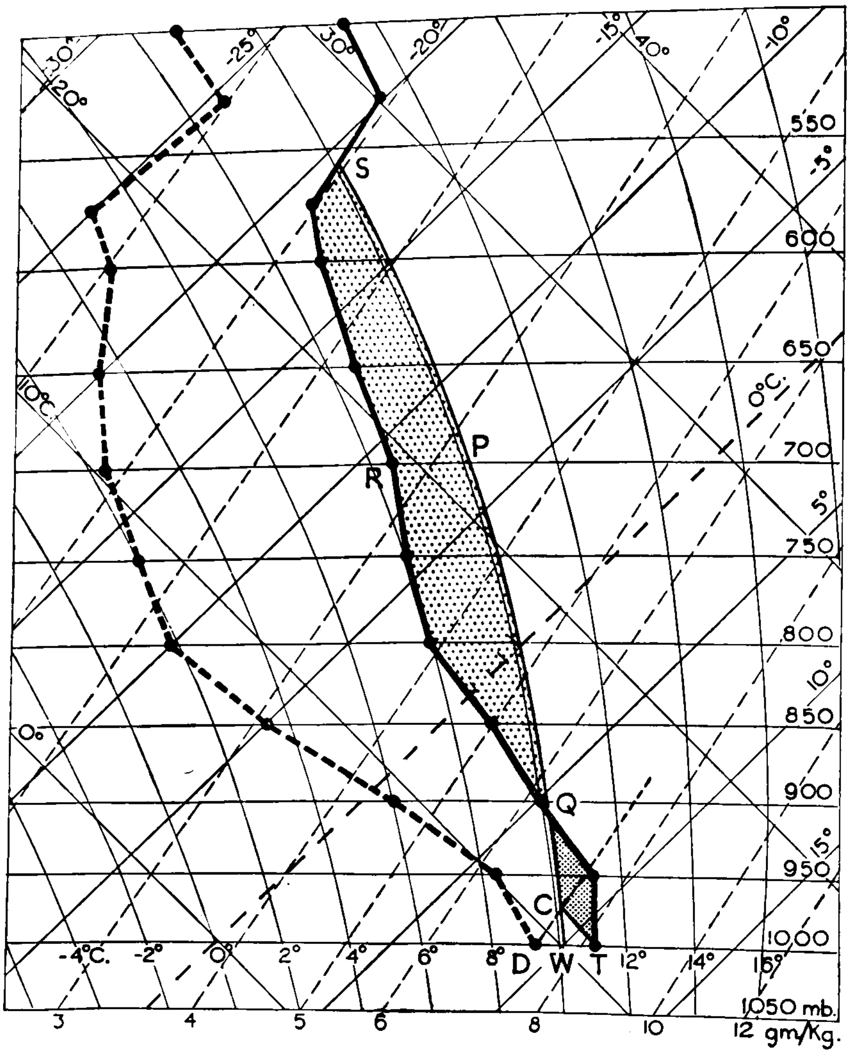

FIG. 49. *Latent instability*

released in this stage accounts for the kinetic energy of the upcurrent and also any electrical energy that may be developed; the area QRSPQ accordingly counts as positive. In this way, the difference between the positive and negative areas between the two curves gives an indication of the intensity of the instability which develops; the larger the net positive area, the more vigorous is the system which would be expected to develop.

Convective instability

Hitherto attention has been restricted to ascent of isolated parcels of air, but we have now to consider the lifting of a whole layer of air. Fig. 50 shows a stable lapse rate AA′ with the associated dew-point curve DD′ of a layer 100 millibars thick. The result of adiabatic lifting of this layer through a further 100 millibars is found by considering what happens to individual levels. Thus air at A (1000 millibars) ascends along the dry adiabatic to C and then along the saturated adiabatic to B where the pressure is 900 millibars. Similarly air at A′ reaches its condensation level at C′ and the 800-millibar level at B′. The resulting lapse rate BB′ is evidently greater than the saturated adiabatic and as the lifted air is everywhere saturated, instability is present. The effect of lifting the layer of air has been to develop instability from what was originally a stable lapse rate. Since for instability the saturated adiabatic through B′ must lie to the left of that through B, it follows that the wet-bulb potential temperature (Section 32) of the air at A′ is less than that of the air at A; in other words, the condition for convective instability is that the wet-bulb potential temperature should decrease with height.

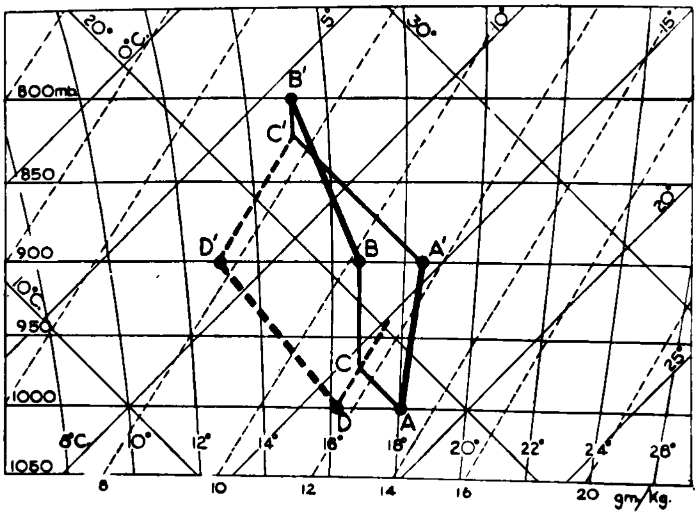

FIG. 50. *Development of convective instability*

Instability due to colder air flowing over warmer air

When the direction of the wind changes with height the upper air may have a very different origin from that of the lower air. For example over north-western Europe southerly or south-easterly winds often have south-westerlies above them. In certain situations, while the lower stream of air comes from warm latitudes the upper stream of essentially cooler air originates in higher latitudes, although at the time it is moving north-eastwards. As the temperature contrast increases, the lapse rate passes the saturated adiabatic and a high degree of instability may become latent; this can be released by any process which gives cloud at an appropriate level and severe thunderstorms may result. The initiating process may be surface heating and development occurs in much the same way as in simple convection systems during the heat of the day, but frequently the warm surface air is too stable for convection currents from the ground to penetrate to the conditionally unstable layer above. Then the ascent necessary to produce condensation must first begin by some other process, possibly orographically or by a general ascending motion associated with a fall in pressure. This effect, which may be only slight, has been referred to in Section 27; it is however worth recording here that locally falling pressure is to be regarded as very significant when the upper levels are known to be potentially unstable. Such cases may contribute by generating convective instability. Many severe summer thunderstorms of the British Isles originate in this way; the first sign in the sky is frequently a cumulus type of cloud in the altocumulus level. Altocumulus is usually flat, but if it begins to develop a turreted structure (castellanus, see Plate XXI, Chapter 12) it is a certain sign of high-level instability which may pass further to the thunderstorm stage. Being independent of surface heating, such storms may break out during the night.

Movement of showers

Instability systems generally drift along in the air current in which they are formed. When they result from surface heating, the geostrophic wind is a fair measure of their speed and direction of travel although the change of wind with height should be considered and the wind at, say, 10000 feet may be a better guide. On the other hand, high-level systems travel with the upper winds and often move at variance with, even contrary to, the wind near the surface when this is light.

Instability at a cold front

An important case of instability is that associated with the displacement of warm air by colder air at a cold front. The study of fronts is undertaken later; it is sufficient to note here that a showery or even thundery type of rain often, but by no means invariably, accompanies the cold front.

36. OROGRAPHIC CLOUD AND PRECIPITATION

The effect of a barrier of hills in the path of the wind is to force the air to rise, as has been mentioned already in the theory of the föhn wind (Section 25) and in connection with vertical motion generally. The rate of cooling by ascent of unsaturated air is 3 degC per 1000 feet, and with adequate lifting the condensation point must ultimately be reached, giving cloud on the hills perhaps with drizzle, or even rain or snow according to the temperature. If there is no precipitation, the condensation level is the same on the lee side of the hill as on the windward side; with precipitation there is less water to be re-evaporated in the downcurrent, and a higher cloud base would be expected on the lee side. This is illustrated in Fig. 51 where the shallow cloud is presumed to give precipitation in the form of drizzle. It is evident that the amount of ascent required to produce cloud depends

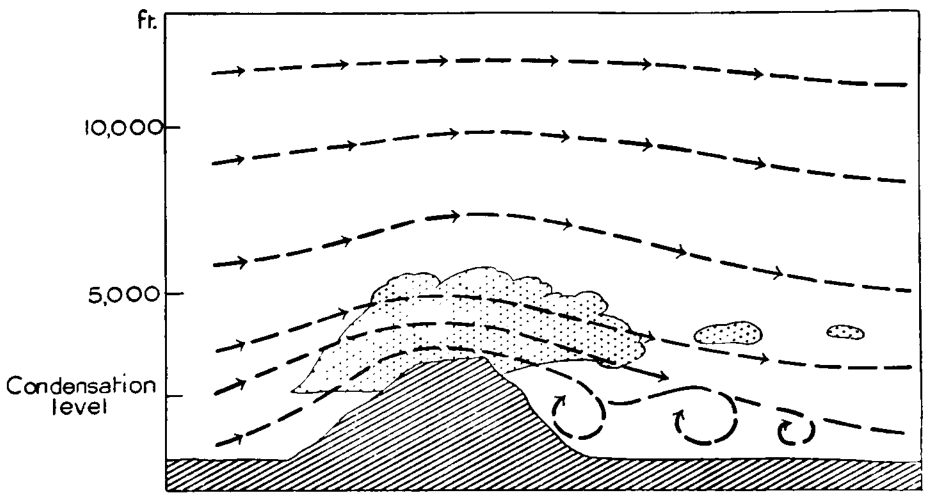

FIG. 51. *Formation of orographic cloud with drizzle*

on the humidity of the air and if this is near saturation quite a low range of hills will give cloud, as is well known in the British Isles where orographic cloud is common with moist south-westerlies on all the exposed high ground and even at times on hills as low as 200 feet. Moist winds from the North Sea give the same effect on the low hills in eastern England. On higher ground, condensation often passes to the drizzle stage but whether more intense precipitation occurs depends on the lapse rate and humidity through a deep layer. When there is widespread ascent continuous rain does not necessarily require an unstable lapse rate, but saturation is necessary through a depth of some 10000 feet or more. In other cases the result of lifting may be to develop convective instability. Such effects often produce a local intensification of precipitation in circumstances where rain is already more or less general for other reasons, as in a depression. In certain parts of the world, purely orographic rain occurs in considerable amounts and has a profound effect on the climate of those areas.

The simplest form of orographic cloud is stratus with a flat base and generally of no great vertical thickness; it forms a sheet covering the higher ground with breaks over lower-lying parts. The descent on the lee side causes warming and the cloud dissolves rapidly. Very marked clearances produced in this way are common and often provide clear flying weather when the windward side of the hills is enveloped in low cloud. For example, the better route between London and Scotland, as regards low cloud, is usually along the eastern coasts with westerly winds and along the western coasts with easterlies; similar local variations are well known to pilots operating on air routes abroad. Orographic cloud, it will be noticed, is continually forming on the windward side of the high ground and clearing on the lee, the cloud as a whole remaining stationary while the wind passes through it.

In some conditions when an extensive cloud sheet is present and the air below is nearly saturated, the slight lifting of the air which occurs in passing over high ground may cause condensation at a lower height above sea level, so that there is a definite dip in the cloud base over even the smaller hills; this should be borne in mind when flying over hilly country in cloudy weather.

When there is a layer of almost saturated air aloft, orographic lifting may cause a persistent cloud cap to form above the high ground (Fig. 52). Like low orographic cloud, this type is stationary with the cloud particles streaming through it, and being little disturbed by turbulence it often presents clear-cut margins; viewed from below it has thin and sometimes pointed ends with a thicker and broader centre, in shape suggesting a lens – it is called 'lenticular' in the international classification. Such cloud may form at great heights in suitable conditions, even at cirrus levels. Its formation in standing lee waves was mentioned in Section 27.

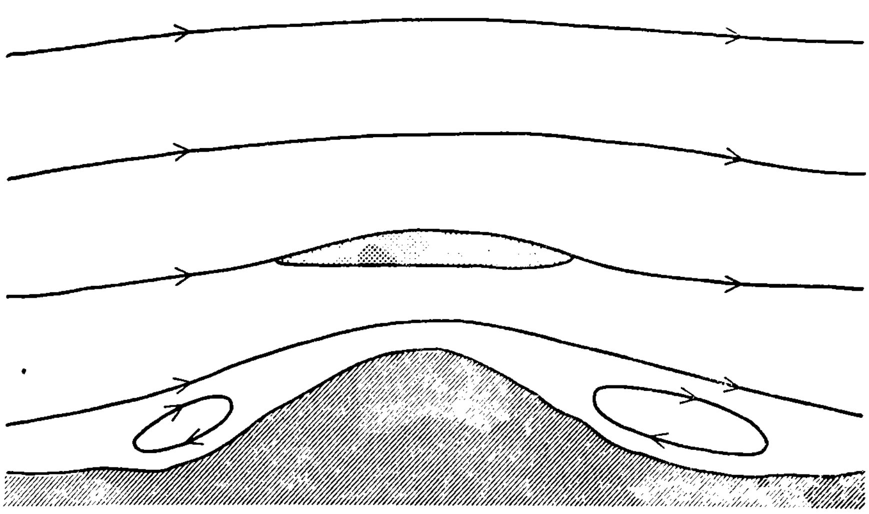

FIG. 52. *Lenticular cloud formed over high ground*

Orographic cloud gives no particular bumpiness, apart from that due to turbulence over the irregular ground, except when complicated by vertical instability when the cloud assumes a cumuliform structure and may give showers or thunderstorms; once formed these may break away from the hills and travel down wind. The presence of orographic cloud suggests the probability of strong downdraughts on the lee side of the mountain range.

37. CLOUD AND PRECIPITATION FORMED BY WIDESPREAD ASCENT

Some characteristics of widespread vertical motion have been considered in Section 27. If the relative humidity is high, condensation occurs throughout the whole depth of rising air and gives extensive cloud masses many thousands of feet thick; on other occasions variations of humidity may result in the formation of separate cloud layers. In such cases the lapse rate of temperature may be stable and remain so after lifting. If the lapse rate is such as to favour convective instability, then the cloud masses take on a more convective character; precipitation becomes heavy and of a showery nature, perhaps with thunder, although individual showers may be prolonged or succeed one another with little intermission. Widespread ascent occurs mainly within the area covered by a depression and for that reason the connection between low pressure and bad weather is brought about – but these relationships form the subject matter of later chapters.

38. PRECIPITATION IN THE FORM OF ICE

Snow

Snow is precipitation in the form of crystals of white ice, apparently opaque, generally in flakes of light feathery structure. Small flakes, up to 4 or 5 millimetres diameter, often show a six-rayed star-like structure of great beauty. Larger flakes usually consist of aggregates of such crystals, the general structure being no longer perceptible. Large flakes are usually found only when the temperature is near 0°C; at low temperatures the snow may fall as a fine dry powder. For precipitation at ground level in the form of snow, the surface temperature must be less than about 4°C.

Ice needles are crystals in the form of single rods about 2 millimetres long. Because of their slow rate of fall they sometimes appear as if suspended in the air and in sunshine they then give rise to a variety of optical phenomena.

Granular Snow consists of white opaque snow-like grains, generally less than 1 millimetre diameter. They do not readily rebound or burst when falling on hard ground.

Sleet in British usage denotes rain and snow falling together, or snow melting as it falls. In the United States the term is used for precipitation of transparent grains or pellets of ice formed when raindrops from warmer air aloft become frozen on falling through a cold layer nearer the ground.

Hail

Soft hail consists of white opaque pellets rarely exceeding a few millimetres in diameter. It falls from shower clouds in cold weather and is small and easily compressed. True hailstones are hard pellets of various sizes and shapes, frequently with a structure of concentric layers of clear and opaque ice alternately. Though usually only a few millimetres in diameter, very much larger stones occur occasionally and some as big as grapefruit, and weighing a kilogramme or more, have been observed.

After the initial formation of ice particles at temperatures well below 0°C, growth is mainly due to collision and coalescence with supercooled water drops. The opaque layers are caused by air trapped within the ice which forms on the particle when the liquid water content of the cloud is small; the clear layers are the result of slower freezing associated with high liquid water content of the cloud.

Hailstones are supported within the cumulonimbus clouds by strong and possibly increasing updraughts. In conditions favourable for the formation of large hail, those stones which are of such a size that their fall-speed is just less than the updraught will remain in the cloud and thus in time grow larger. Some occasions of large hail are described in Section 42.

Sometimes the hailstones melt before reaching the ground; this in part accounts for the rarity of hail at low levels in equatorial regions.

39. ARTIFICIAL RAIN

In Section 32 the theory was described that the release of precipitation from a cloud does not in general take place until after ice particles have formed in its upper part. This has led to attempts at rain-making by the introduction of material into a non-precipitating cloud intended to facilitate the production of ice crystals. The method most often used has been to 'seed' a well-developed convection cloud with solid carbon dioxide dropped from an aircraft. The low temperature of this substance, less than —65°C, cools the air coming into contact with it so that ice crystals are formed and precipitation may be initiated. Other attempts have been made to release precipitation by introducing large water drops into the lower part of a convection cloud; according to the coalescence theory the drops might be expected to start the rain process when carried by the updraught into the higher parts of the cloud. Clouds have also been seeded with silver iodide crystals produced by burning the substance upwind of the area to be seeded.

While some success has been achieved with these and similar methods, the results so far have been by no means spectacular; limited falls of rain have been initiated only in favourable situations which were near to the point of giving rain naturally. The experiments have not yet become of economic importance.

CHAPTER 7

THUNDERSTORMS

40. CONDITIONS FAVOURABLE FOR THUNDERSTORMS

For a thunderstorm to develop, the conditions required for the development of instability clouds (described in Section 35) must be present to a marked degree. Briefly, these conditions are:

(i) Lapse rate greater than the saturated adiabatic throughout a layer of considerable depth and extending several thousand feet above the 0°C level.

(ii) Adequate supply of moisture from below.

(iii) A process which produces saturation in the region of high lapse rate.

The height reached by instability cloud is usually limited by mixing with and evaporation into the drier environment, and is thus less than the height limit set by the buoyancy alone. If the relative humidity of the surrounding air is very low, the evaporation process may become dominant; the upward growth of the cloud then ceases at a height well below that which would be expected from consideration of environment lapse rate alone. It is to be noted that evaporation into the entrained clear air tends to reduce the temperature and to increase the lapse rate of the ascending air to a value greater than saturated adiabatic, depending on the relative humidity in the clear air.

Condition (i) implies that the environment lapse rate must be conditionally unstable through a deep layer, since lapse rates exceeding the dry adiabatic are not found in the free atmosphere. Before instability can be realized, saturation must be reached and this is usually brought about by ascent of air as a result of one or more of the processes already described: (*a*) insolation over land, (*b*) advective heating, (*c*) orographic lifting, (*d*) frontal lifting, (*e*) lifting resulting from convergence usually in association with falling pressure at the surface. For details of these processes, reference should be made to Section 35 and, for (*d*), to Chapter 16.

When the temperature within a cumulus cloud falls below —10°C supercooled water drops begin to change to ice crystals. The onset of glaciation starts precipitation and as this increases in volume a stage is reached where its containing air acquires negative buoyancy due to cooling by evaporation. The descent of cool air improves the ventilation of the storm and thus accelerates the increase in cloud volume. At the same time turrets or towers grow rapidly and, if not sheared off by rapid changes of wind with height, penetrate the upper troposphere in succession.

Types of storms

Thunderstorms are sometimes classified as either 'heat' or 'frontal' storms according to the method of formation. Heat storms are perhaps better referred to as 'air-mass storms', the name implying that the meteorological conditions are (to start with, at least) more or less uniform over a large area in any horizontal plane. Frontal storms, on the other hand, usually occur where a cold air mass undercuts a warm mass, that is, at a cold front, although occasionally they break out also at a warm front; discussion of the formation and movement of storms of this type is deferred to Chapter 16.

An air-mass storm is often associated with intense insolation over land but other causes, acting either independently or concurrently, include the convergence associated with falling pressure at the surface and the progressive steepening of the lapse rate aloft. Particularly favourable pressure situations in the neighbourhood of the British Isles are a col or a weak depression, since in these the requisite conditions of lapse rate and humidity are often fulfilled. Air-mass storms are also likely in a mass of relatively cold air made unstable by its passage over a warmer sea, particularly if the air is later subjected to orographic lifting. Generally, when conditions become favourable over a considerable area, there is always a doubt as to exactly where storms will break out or what path they will follow, since these details depend on factors which are not easily assessed.

41. THE CELLULAR STRUCTURE OF THUNDERSTORMS

A thunderstorm is a collection or complex of cloud cells in different stages of development. The diameter of individual cells varies from one to five miles, while between neighbouring cells there are rather narrower cloud-filled lanes.

It is convenient to consider the development of a cumulonimbus cell in three stages.

Cumulus or building stage

This occurs when one or more small cumulus clouds begin to grow into a large cumulus with a base up to five miles across. A general updraught prevails throughout the cell at this stage with perhaps extreme vertical velocities of 30 metres per second.

Inflow to the cell takes place through the sides at all levels as well as through the base of the cloud. The average life of this stage is short, perhaps 15 to 20 minutes.

Mature stage

This stage (Fig. 53) begins when some of the droplets are large enough to fall from the cloud as precipitation. Cloud tops extend thousands of feet above the 0°C level and are still rising. The downdraught associated with the increasing volume of precipitation warms at the saturated adiabatic lapse rate. The surrounding cloudy air has a lapse rate greater than the saturated adiabatic because during ascent it will have entrained some cloud-free air. Thus the downdraught soon acquires a temperature lower than that of its immediate surroundings and will continue moving downwards. Below the cloud base evaporation of the rain assists the cooling process and accelerates the downdraught which on reaching the ground forms a cool gust or squall that spreads outwards from the storm.

Third or dissipating stage

This is reached when the storm has used up the local supply of moist air and the updraught is possibly cut off by the spreading downdraught. The precipitation diminishes until the last drops have fallen. So long as the updraught is vigorous, as in the first and second stages, the cloud maintains its typical cauliflower structure, but in the third stage equilibrium with the environment is attained by the rising air which then begins to spread laterally. If at this stage of maximum development the cloud has extended to the region where large numbers of ice crystals are formed, the spread of the updraught is apparent in the development of the cumulonimbus anvil (Plate XIV).

Although the life of a cell is about two or three hours, the most active period, comprising the first and second stages, is confined to the first 30 to 40 minutes. After this period the cell, though at first containing much cloud and precipitation, is for

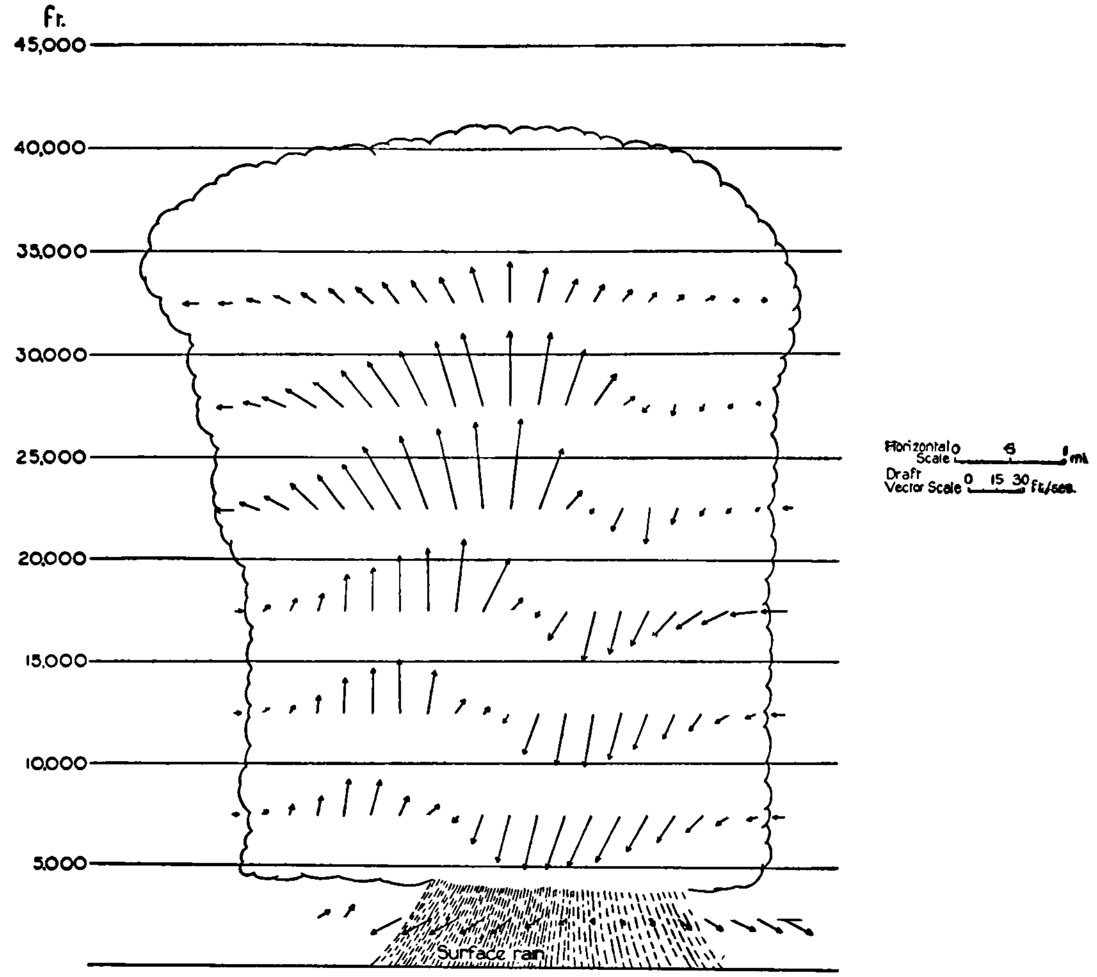

FIG. 53. *Circulation within the mature stage of a cumulonimbus cell* (after Byers and Braham*)

(The heights shown correspond to a well-developed subtropical thunderstorm, and should be suitably reduced for a storm in temperate latitudes.)

the most part subsiding and taking little part in the thunderstorm as a whole. A large storm consists of an agglomeration of several cells but at any one time most of them will be in the comparatively prolonged dissipating stage. As the cold downdraughts from the mature and dissipating cells spread out above the ground, they cause convergence in adjacent regions and so may set off new convective cells near by. Even an existing cell in a late stage of decay may be regenerated in this way, a new active cell forming within the stratified cloud of the old one.

Fig. 54 is taken from a study of thunderstorms occurring on 18 June 1958. The estimated tops of the cloud by the parcel method would be about 49000 feet. In practice the limit set by buoyancy alone is not reached because of the mixing of the cloud with the drier environment. On this occasion the observed highest tops were 43500 feet while some tops were no more than 29000 feet.

42. SELF-PROPAGATING STORMS

Effects of wind shear

On occasions a wind shear with height may be adverse to the development of large cumulus clouds, but if the instability is great enough for development to take

*BYERS, H. R. and BRAHAM, R. R.; Thunderstorm structure and circulation. *Jnl. Met.*, *Lancaster, Pa.*, 5, 1948, pp. 71–86.

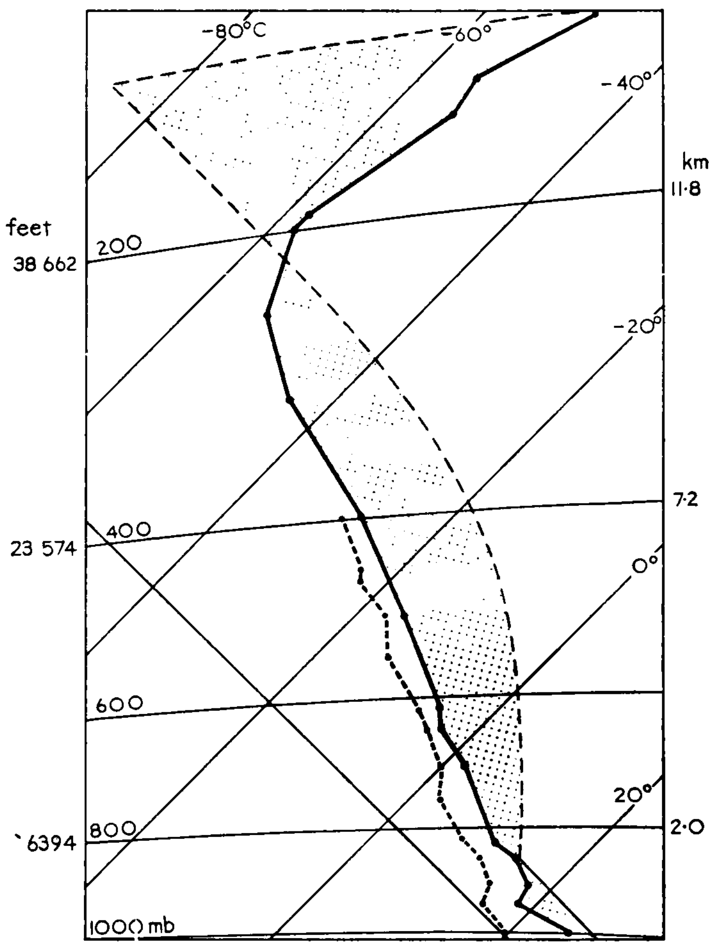

FIG. 54. *Typical tephigram for thunderstorm development* (after Ludlam*)

——— Temperature .- - -. Wet-bulb temperature

place despite the shear then the updraught and the axis of the cloud will be tilted out of the vertical. The falling precipitation passes through only a small section of the rising air and the updraught can continue until its supply of energy is exhausted.

When a wind shear with height is associated with a great depth of instability and moist surface layers, cumulonimbus clouds become especially well organized. If there is a layer of drier air at medium levels, above the moist surface layer, particularly propitious conditions exist for the development of severe thunderstorms. The downdraught, commencing by precipitation, is then enhanced by evaporation of the precipitation into the dry medium-level air and becomes invigorated and persistent. In the presence of this cold downdraught the tilted updraught becomes part of an overturning in a deep layer of the lower troposphere. Individual cloud towers can no longer be recognized in the summits; the ascending air may attain speeds of 50 metres per second or more and penetrate a few thousand feet into the stratosphere before settling back into the anvil.

Storms developing to this stage become persistent and self-propagating; they move slowly over long paths and produce heavy rain, hail and possibly tornadoes for periods extending up to several hours.

In this large-scale flow mixing between environment and ascending air is reduced and in the centre of the cloud cooling is probably near adiabatic. In the tephigram

*LUDLAM, F. H.; A case-study of thunderstorms on an occasion of small wind shear. London, Imperial College of Science and Technology, Department of Meteorology, Tech. Note No. 11, 1962.

in Fig. 54 the shading shows the positive and negative areas corresponding to the adiabatic ascent of an unmixed parcel of saturated air; the positive and negative areas are equal. The positive area is proportional to the energy liberated by the ascending parcel and this could be dissipated by the further ascent of the parcel to the top of the negative area.

A self-propagating storm is illustrated in Plate XXIX. The photograph was taken over the mid-west of the U.S.A. from an aircraft flying at 65000 feet. The turbulent area surrounding the main dome was about 40 miles in diameter and was produced by a family of storm cells. The summit, at about 60000 feet, of the giant cell (near the visible horizon in the plate) was not far beneath the aircraft and protruded some 10000 feet above the surrounding turbulent cloud. The top of the anvil sheet was near the tropopause at about 40000 feet.

This type of storm is responsible for the devastating tornadoes and giant hail often reported from the mid-west of the U.S.A. Severe storms occasionally occur over Europe and southern England. Large hail and tornadoes developed in a storm at Horsham, Sussex, in September 1958. A storm which gave falls of large hail at Wokingham, Berkshire, in July 1959 is well documented and Fig. 55 shows a model,

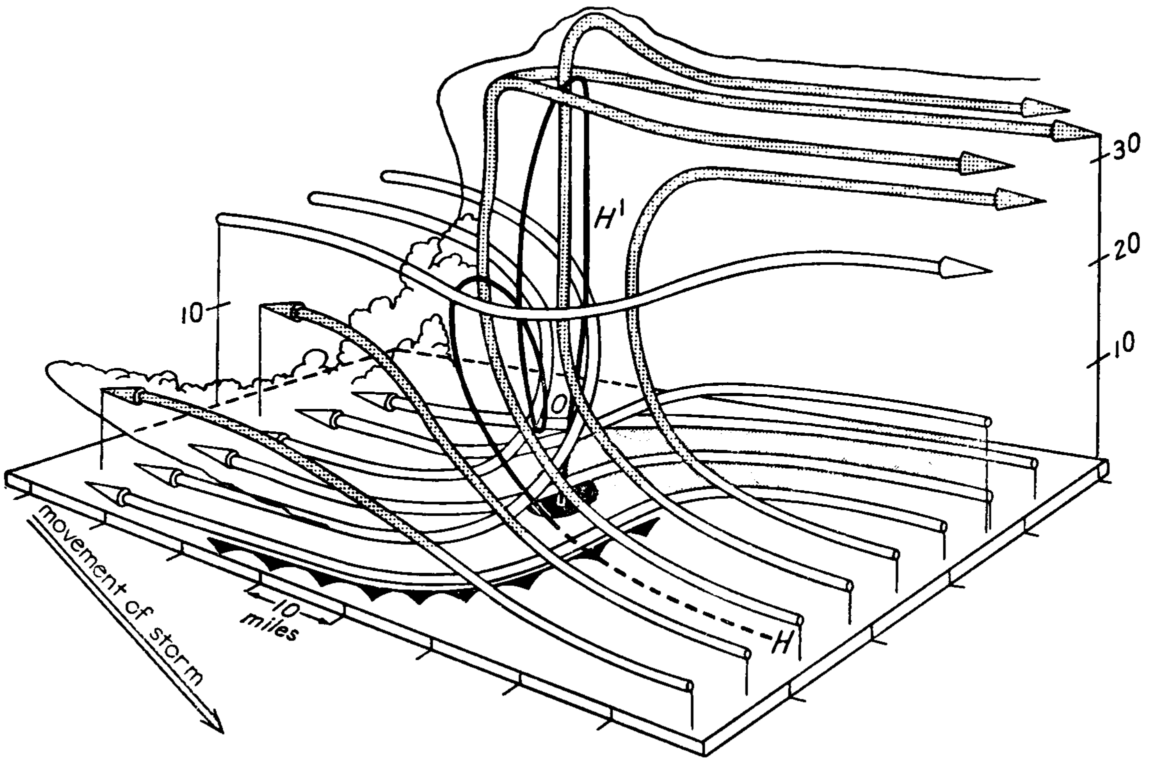

FIG. 55. *Model of airflow in a severe storm cumulonimbus* (by courtesy of Dr F. H. Ludlam*)

HOH′ Path of particle which eventually becomes a large hailstone; it re-enters the updraught at O, falls out at H′ and finally reaches the ground in the black hail area.

Streamlines are of motion relative to the storm and are shaded where condensation has occurred. The anvil cloud and cumulus belt over the trailing edge of the squall front (marked as a cold front) are drawn in schematic outline. Rain reaches the ground in the stippled area ahead of the hail and to the left of the direction of advance of the storm.

Heights are in thousands of feet. The model was constructed for the Wokingham hailstorm of 9 July 1959.

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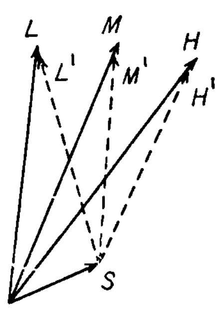
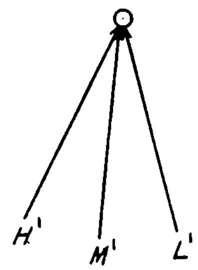
*London, Ministry of Aviation. Atmospheric turbulence and its relation to aircraft. Proc. Symp. R.A.E. Farnborough, November 1961. London, HMSO, 1963.

devised by F. H. Ludlam, of the airflow in the storm. The updraught enters the front of the storm and rises to the anvil dome before subsiding and spreading out into anvil cloud under the tropopause. In this model, small hail, formed in the first ascent, can fall again into the updraught and be recycled to grow into large hail. The upper winds on this particular occasion were almost the reciprocal of the lower winds, but the process is similar when an increase of wind with height is associated with any change of wind direction.

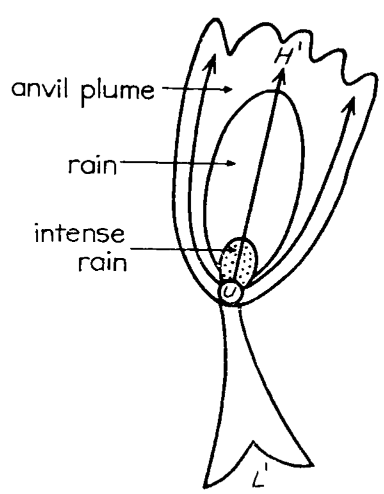
Fig. 56 shows vector winds and a plan of the flow of the updraught and downdraught devised by D. E. Pedgley for a severe storm which passed over Dover in October 1964.



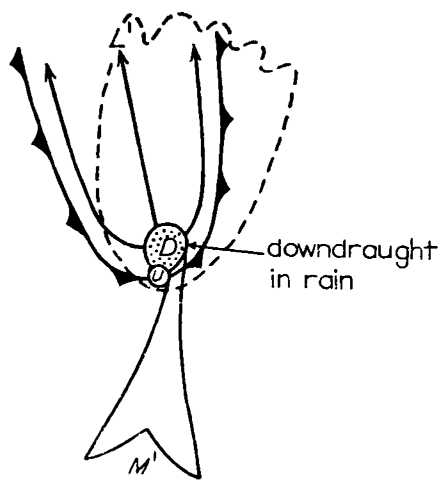
(*a*) Lines showing vector winds over Kent during the afternoon of the 14th at low, *L*, medium, *M*, and high, *H*, levels within the convection layer. *S* shows the velocity of the storm.



(*b*) Variation of wind with height relative to the storm at low, *L'*, medium, *M'*, and high, *H'*, levels.



(*c*) Schematic flow of updraught air as seen from vertically above the storm. *U* is core of updraught. Low-level air enters the storm from *L'* and spreads out at high level as indicated.



(*d*) Schematic flow of downdraught air as seen from vertically above the storm. *D* is core of downdraught and its edge at the ground is a mesoscale cold front.

FIG. 56. *Dover storm of 14 October 1964* (after Pedgley* and reproduced by courtesy of *Weather*)

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*PEDGLEY, D. E.; The Dover storm of 14 October 1964. *Weather, London*, 10, 1965, p. 351.

43. FURTHER CHARACTERISTICS OF THUNDERSTORMS

Diurnal and seasonal variation

In so far as insolation over land is a cause of thunderstorms, they are most likely to occur during the afternoon and to die out in the evening. However, a large convective system once developed is apt to maintain itself for several hours, so that storms may continue well into the night; on other occasions, drifting with the upper wind, they may break out at night or even during the following morning over areas not previously affected. Storms started by advective heating can occur at any time, although they are more frequent by day over land since the effects of diurnal heating are then also present.

In middle latitudes, storms over the land are most frequent in summer. However, the frontal type of thunderstorm is more frequent in winter, because of the greater frequency of cold fronts in that season. In some tropical regions the effects of high humidity, steep lapse rate, intense insolation and convergence combine to make thunderstorms or showers of almost daily occurrence. Over land such storms occur mostly in the afternoons, but over the sea they are often more frequent at night.

First gust

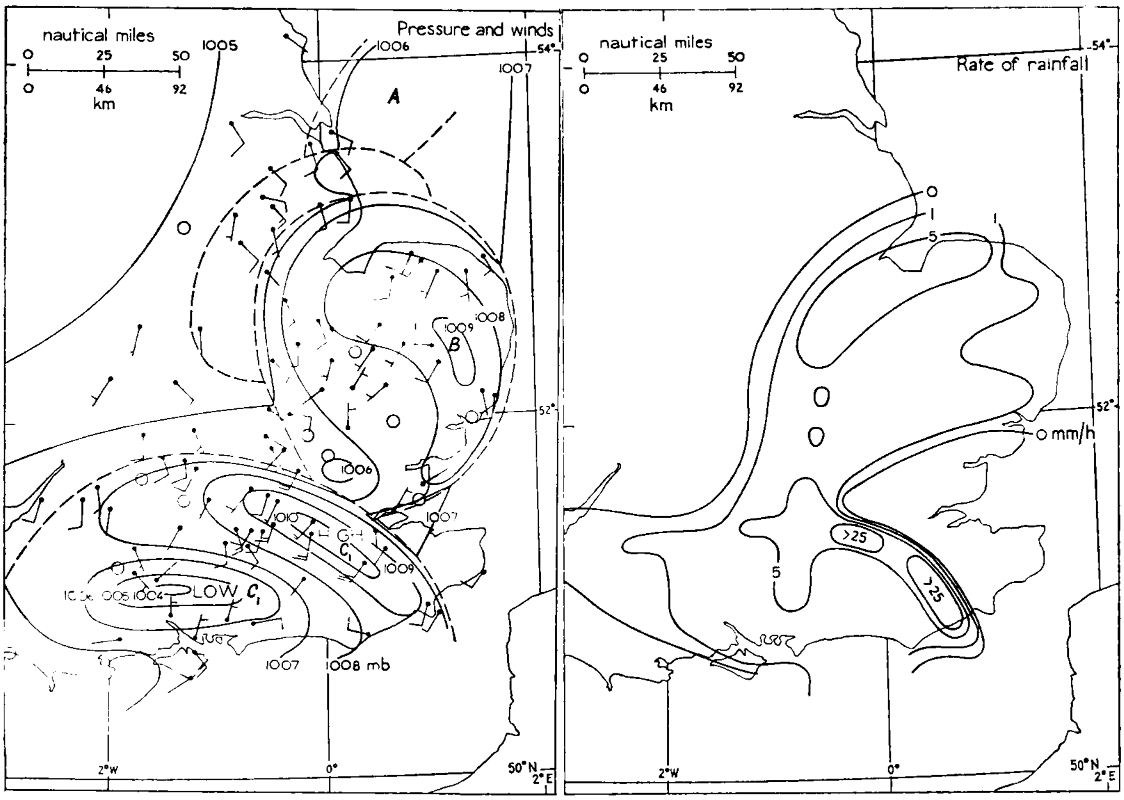
As the downdraught from a cell nears the ground, it spreads out horizontally and its leading edge has been called the 'first gust'. The downdraught has been cooled relatively to its surroundings (see Section 41) so that the outflow above the ground takes on the character of a miniature cold front (see Chapter 16), often giving a severe squall with a marked increase of wind speed, probably accompanied by a change in direction. The cold air of the downdraught spreads out all round the cell so that the first gust is usually directed away from the storm; its direction is influenced by the distribution of the several cells of which the storm is composed, but the motion of the storm as a whole tends to make the first gust stronger ahead of the storm than behind it.

Thunderstorm high and pressure jumps

Marked fluctuations of pressure often accompany severe thunderstorms and the first gust frequently produces a sudden pressure rise. On these occasions local rises in pressure during the growing stage of the cumulonimbus develop into a region of high pressure which expands as the rain area increases. When a thunderstorm high has become intense its leading edge is marked by a large pressure gradient and its centre corresponds to the position of the most violent storms. Fig. 57 shows the small-scale (mesoscale) pressure field together with the boundaries of the meso-systems, the rate of rainfall, and the anemogram and barogram for Croydon on 28 August 1958. The data are taken from a detailed analysis of thunderstorms on that date (see Appendix III).

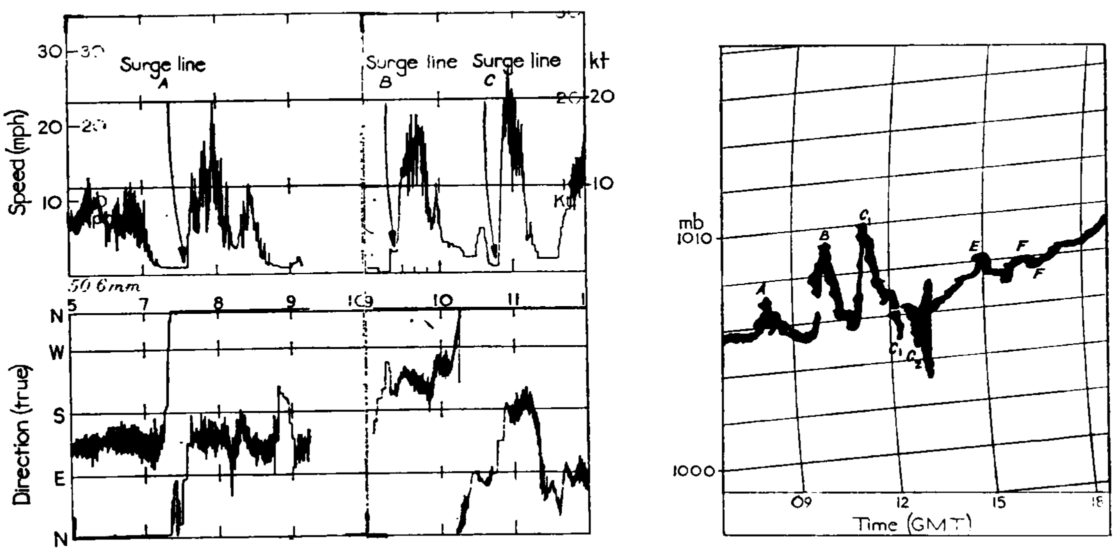
Turbulence

Turbulence refers to a sequence of local variations of air motion which may be regarded as superimposed on the general horizontal flow in the area. In a thunderstorm there are strong updraughts and downdraughts, and gusts occur both inside and outside these main draughts. Although many flights through thunderstorms were undertaken during an intensive thunderstorm project in the U.S.A. in the late 1940s, it now seems probable that the most severe storms were not encountered and the findings should be interpreted with caution.



(*a*) Pressure and winds at 1100 GMT
—— Isobars at 1-mb intervals
– – – Boundaries of meso-systems

(*b*) Rate of rainfall at 1100 GMT
Isohyets for rates of 0, 1, 5, and 25 mm/h.



(*c*) Anemogram for Croydon on 28th
The letters correspond with the systems identified in (*a*).

(*d*) Barogram for Croydon on 28th
Systems C_2, E and F developed after 1100 GMT.

FIG. 57. *Thunderstorm high and associated rainfall and changes in pressure and wind on 28 August 1958* (after Pedgley*)

It is possible to avoid cumulonimbus by avoiding the radar echoes (see Section 87) which are received from them, but severe turbulence has been encountered, even near the level of the cloud tops, as much as 10 miles from the radar echo. Until there

*See Appendix III.

is some way of distinguishing the most severe storms from the less severe ones, the safest course is to avoid flight through active thunderstorm cloud.

Hail

Heavy hail is rare, and when it occurs only a small region of the cloud is affected, so that it is encountered relatively infrequently in flight. Small hailstones cause only superficial damage but there have been reports where large stones have holed and splintered windscreens, shattered astrodomes, ripped off de-icer boots, and bent radiator fins. Damage by hail usually increases with the speed of the aircraft. There is no reliable method of recognizing in advance thunderstorms which will produce large hailstones. It is thought that large stones grow because small hailstones, formed on the first ascent, fall into the updraught of the storm and are recycled. As stones grow in size they travel more slowly in the updraught and thus have time to increase into large hail.

Clear layers of ice form on the stones in those parts of the cloud where supercooled drops predominate and the stones grow with a wet surface. Higher in the cloud, where ice crystals predominate or supercooled drops are small, opaque layers of ice form (see Section 38).

Lightning

A lightning flash is a large-scale example of an electric spark. A spark or discharge occurs between two points when the difference in electric potential reaches a certain value depending on the conductivity of the air and on the distance between the points. In clear air of normal density this critical 'field strength' is about 3 000 000 volts per metre, but in cloud it is reduced to about 1 000 000 volts per metre. Such intense fields must be present somewhere in a thunderstorm before lightning can occur, but it appears that they are very localized, the field strength generally within a thunder cloud being of the order of 100 000 volts per metre. In any case, these differences of potential imply the existence of electrically charged areas within the cloud. Investigations by sounding balloons and by other means reveal the existence of a positively charged area in the top of the cloud and a negatively charged area lower down, the cloud being said to have positive polarity. Below the area of negative charge there is often a second but more localized area of positive charge. A number of theories have been put forward to account for the separation of positive and negative charges into different parts of the cloud, but none of these has yet satisfactorily explained all the known facts, and even the relative importance of the many possible processes of charge separation is unknown. It is unnecessary to describe any of these theories, but among the suggested processes may be mentioned the movement of water drops in an existing electric field, breaking and coalescence of drops, friction between ice crystals, evaporation and melting of ice particles, freezing of water drops, and sublimation of vapour on to ice particles. All these processes are found to produce a separation of electric charge in the laboratory.

Once the critical field strength has been built up, a lightning discharge occurs and temporarily neutralizes the field, but if the cell is still active the building-up process is immediately resumed. Discharges may take place from cloud to earth, between two different clouds or two parts of the same cloud, or from a cloud to the surrounding air. An active storm consisting of several cells may produce lightning flashes at an average rate of four a minute; of these, the number entirely within the cloud much exceeds the number passing from cloud to earth.

The visible lightning stroke is a channel of incandescent air not more than an inch or two in diameter. Thunder is simply the accompanying explosive report caused by the sudden expansion of the air heated by the lighning flash. Since the flash and the noise originate simultaneously, the distance of the lightning from the observer may be estimated from the interval which elapses between the times when the lightning is seen and the thunder is heard. The flash is seen immediately, for light travels 186 000 miles per second, but sound takes five seconds to travel one mile and the thunder is therefore heard an appreciable time after the lightning is seen. As the flash itself may be a mile or more in length, the sound which reaches the observer will not all have travelled the same distance so that the noise may continue for several seconds. Ordinary echoes often add to this effect.

A lightning strike can be very unpleasant to the occupants of an aircraft; it may occur in or beneath cloud, or between two clouds, and even before a thunderstorm has developed (Section 44). The brilliant flash, the smell of burning and the explosive noise may be alarming and distracting. As an extended trailing aerial increases the likelihood of a strike, such aerials should be earthed and wound in. However, if the aerial winch has to be wound by hand, there is a risk of injury to the operator if a strike occurs at this time. In such cases, if the aerial has not been already wound in, it should be earthed and left trailing. While many aircraft have been struck by lightning, there is little positive evidence of serious damage to metal aircraft by the strike itself, and the occupants are safeguarded by the aircraft bonding requirements which effectively prevent any electrical discharge from penetrating to the interior. Nevertheless there is a danger that in the turbulence of a storm the disconcerting effects may lead to loss of control unless pilots are fully prepared. When a flight occurs in a thunderstorm at night, cockpit lighting should be turned on fully to reduce dazzle. Where two pilots are carried, a further protection is for one of them to wear dark glasses. This may not be practical where there is only one pilot because of the difficulty of reading instruments through such glasses.

Static electricity

This refers to the spark discharges which occur when the potential difference between the aircraft and the ambient air exceeds a certain value; it is usually due to a charge accumulated from cloud particles, from precipitation, from contact with dust or sand, or to the aircraft's not immediately taking up the potential of its surroundings after a rapid change in the level of flight. The phenomenon is usually noticed first as radio noise, particularly on high and medium frequencies; VHF reception is affected to a much lesser extent. As the static electricity increases in severity, the noise increases and in extreme cases a visible discharge – known variously as St Elmo's fire, brush discharge or corona – is observed around some parts of the aircraft. Static electricity is not normally dangerous although there have been rare occasions where a discharge has caused the breakage of windscreens and plastic panels. The onset of static discharge is likely to be delayed in aircraft fitted with discharge wicks, but even then discharges are still likely to be experienced in thundery conditions. (See also Section 44).

Precipitation static. This refers to the static electricity produced when an aircraft encounters precipitation. In this case the charge accumulates from the electricity carried by precipitation elements which strike the aircraft, from the electricity generated by the breaking of particles (especially snow crystals) against the aircraft, from friction with solid particles, or in other ways. In particular, charges imparted to the aerial contribute to radio noise.

Noise. This is caused by electrical effects on the radio. In combination with the effect of hail and heavy rain striking the aircraft the noise may build up to an alarming extent, submerging the normal engine noise and preventing speech on the intercommunication system. The explosive noise of a lightning strike has already been mentioned.

Instrument errors

During a thunderstorm, rapid variations of surface pressure of the order of a millibar can occur, particularly during periods of heavy rain. These affect the indications of the pressure altimeter although it appears that differences between readings of the radio and pressure altimeters in such circumstances are too small to be operationally significant. Both the pressure altimeter and the rate-of-climb indicator may give faulty readings because of localized turbulence. Partial blocking of the pitot tube by heavy rain may cause the airspeed indicator to read low. If that power has been selected which gives the safest speed before entering the storm, then any fluctuations in airspeed readings should be disregarded provided a reasonably level attitude is maintained. A lightning strike may seriously affect magnetic compasses, which should be checked as soon afterwards as possible. When approaching the storm, the auto-pilot should be disengaged and the flight instruments checked.

Ice accretion

This is discussed in the next chapter.

44. ATMOSPHERIC ELECTRICITY

There are various electrical phenomena which occur naturally in the atmosphere, the thunderstorm being the most spectacular. Investigation shows that the atmosphere even in fine weather carries a net positive charge, implying a corresponding negative charge on the ground. Measurements of the electric potential show that it increases with height; the rate of change with height is called the potential gradient. Near the ground in fine weather the potential gradient is of the order 150 volts per metre but fluctuations occur continually and there are also regular diurnal and annual changes. The potential may reach hundreds of thousands of volts by the middle troposphere, the earth being assumed to be at zero potential.

The potential gradient is increased in haze, fog or cloud. The value at the surface may reach 2000 volts per metre in fog, sufficient to cause static discharges from the extremities of metallic conductors connected to earth. When there is precipitation in any form, the electric field becomes more seriously modified. With steady slight rain or drizzle, the electric field may be either increased or decreased compared with the fine-weather value, but no large electrical forces are produced either inside or outside the cloud. With showery rain, especially from detached clouds, strong electrical forces are brought into existence both in and near the clouds. These forces increase with the intensity of the showers and reach their climax in thunderstorms when a spark discharge of lightning occurs.

The occurrence of static discharges from parts of an aircraft has already been noticed (Section 43); they can occur also at the surface of objects attached to the earth. The immediate cause is a sufficient difference in electric potential between a projecting object and the air. It should be appreciated that although thunderstorms provide the most favourable conditions, the phenomenon of static discharge is not confined to those occasions but may occur whenever the electrical field is sufficiently intensified in any way, usually in association with disturbed weather of the types mentioned. An extreme instance concerns cases where aircraft have actually been struck by 'lightning' when no thunderstorms were observed in the vicinity and when radio noise was not enough to suggest the likelihood of a discharge. This comes about through the intensification of an electric field in the vicinity of a conductor, which in this case is the aeroplane itself. If the general field strength has increased in convective conditions, even though these have not developed to the stage of thunderstorms, then in a critical case the further local intensification in the neighbourhood of an aircraft is sufficient to cause the sparking value to be reached, thereby initiating a lightning strike.

CHAPTER 8

ICE ACCRETION ON AIRCRAFT

45. INTRODUCTION

The possibility of ice accretion on the airframe of an aircraft has always to be considered when flight takes place through cloud or rain at temperatures below 0°C, while engine icing may occur also in clear air and at temperatures above 0°C. While various methods are available for the prevention of accretion or for removing a deposit of ice after it has formed, there is none that can provide more than partial protection. It is therefore important that a pilot should be aware of the conditions favourable for ice formation so that he can avoid the risk as far as possible and may know what action to take should ice begin to form. In addition to a description of the physics of ice accretion as affecting both airframe and engine and of the meteorological conditions in which it occurs, this chapter includes an account of the effect of ice formation on performance and of the procedures to be followed to reduce the risks. Reference should be made elsewhere for information on methods of prevention and de-icing. It should, however, be noted that the design and application of such methods requires to be properly related to the various types of ice liable to be encountered and to the rate of accumulation. A method which is perhaps effective in one set of icing conditions may be of little use in another.

46. FORMS OF AIRFRAME ICING

Two physical processes may cause a deposit of ice on objects exposed to the atmosphere. Ice may form directly from water vapour, i.e. by sublimation, or ice may form by the freezing of liquid water drops. On the ground these processes produce two familiar forms of ice deposit known as hoar frost and rime. Glazed frost (alternatively 'glaze ice' or 'clear ice'), which is in the second category, forms when raindrops, which may themselves be supercooled, freeze on striking cold ground.

The same forms of ice may affect aircraft in flight and similar names are used to describe them. Aircraft encounter a greater variety of icing conditions in flight than on the ground, resulting in a wider variety of forms of ice deposit.

Hoar frost

This type of icing occurs in clear air on a surface whose temperature is reduced below the frost-point of the air in contact with it and of course below 0°C (Section 31). The water vapour in excess of that necessary to saturate the air with respect to ice condenses into a white crystalline coating of ice, normally of a feathery nature. It may occur on a parked aircraft in the same circumstances that lead to hoar frost on the ground, that is during a clear night when there is a fall in temperature to a value below 0°C. The weight of the deposit is unlikely to be serious but it can interfere with the airflow and the attainment of flying speed during take-off, with vision through the windscreen, with the free working of such moving parts as ailerons and with the efficiency of wireless reception. Any hoar frost should therefore be carefully removed from the aircraft before taking off.

Hoar frost may occur in flight if the aircraft, after flying in a region where the temperature is well below 0°C, moves rapidly into a warmer and damp layer of air; this may result for example from descent to a lower level, or from ascent into an inversion. If the air passing over the cold aircraft is chilled to a temperature below the frost-point, hoar frost is deposited but this soon disappears as the aircraft warms up. The effects are similar to those of hoar frost on the ground. There may be some loss of radio facilities, frost on the windscreen just before landing may cause much inconvenience, and frost on the airframe may increase the stalling speed. In a rapid descent, hoar frost can also form inside the aircraft, obscuring the instruments and the view through the windscreen unless these parts are fitted with protective heating.

Rime ice

Rime occurs when small supercooled water drops freeze on contact with a surface at a temperature below 0°C to produce tiny ice particles between which air is entrapped to give a rough crystalline deposit. At ground level it forms in freezing fog and the white crystalline deposit grows out on the windward sides of exposed objects.

In flight an aircraft may encounter clouds of low water content composed of small drops comparable with those of freezing fog, and it is subject to icing which in appearance and method of formation is similar to that formed on ground objects in freezing fog. This type of ice accretion is most liable to occur at low temperatures, where the unfrozen cloud droplets tend to be small and therefore freeze almost instantaneously. Rime ice forms and accumulates on leading edges with no spreading back. The air entrapped between the particles gives the ice a white opaque appearance and it usually breaks away easily. Ice of this type usually has little weight but it alters the aerodynamic characteristics of the wings and it may block the air intakes.

Clear ice (glaze ice)

This consists of a transparent or translucent coating of ice with a glassy surface appearance. It results from water flowing over an airframe before freezing. The drops unite while in the liquid state and very little air is enclosed between them. Although the ice surface is smooth it is not always even and bumps and undulations occur. Ice formed in this way is tough, sticks closely to the surface of the aircraft, and cannot be broken away easily. If it breaks at all it comes away in large pieces which sometimes are of a dangerous size. The danger of clear ice is primarily aerodynamic but it is increased by the weight of the accumulation and by the vibration set up by unequal loading of wings, struts and propeller blades.

Clear ice forms when large water drops at temperatures not far below 0°C are encountered in flight. The freezing process is comparatively slow. The water spreads back and freezes in contact with the cold surface. It occurs in dense cloud of convective or orographic type where large liquid water drops may be carried up in vigorous vertical currents to levels where the temperature is below 0°C.

Clear ice may occur when a rapid descent is made; the aircraft temperature lags behind the ambient air temperature, and if rain is encountered whilst the temperature of the aircraft is still below 0°C the relatively large drops form clear ice over a large part of the aircraft with considerably spreading.

A further example of clear-ice formation in rain occurs when there is an inversion of temperature and rain falls from a level where the temperature is above 0°C to a layer where the temperature is below 0°C. In the lower layer impact on an aircraft results in clear glassy ice formation. These conditions, which occur very rarely, are found in association with warm fronts when the icing layer occupies a narrow range of altitude below the frontal surface (Chapter 16).

Cloudy ice or mixed ice

Rime ice and clear ice are the extreme forms of ice accretion experienced by aircraft flying in cloud and rain, but as a large range of drop sizes may be encountered at temperatures from 0°C to —40°C a wide range of forms of icing exist between these two extremes. These varieties are usually described as cloudy ice or mixed ice. The smaller the drops and the lower the temperature the rougher and more cloudy will be the build-up on the leading edges, whilst the larger the drops and the nearer the temperature to 0°C the greater will be the tendency for a smoother and more glassy ice formation with spread-back over the airframe.

In clouds of liquid droplets where temperatures are below 0°C, ice crystals may be present. These tend to stick if they strike the wet surface of an aircraft and become frozen along with the cloud drops to give a formation of rough cloudy ice. If snowflakes are present they are similarly imprisoned within the ice as it forms, producing an opaque deposit with the appearance of tightly packed snow which is called pack snow.

47. FACTORS AFFECTING THE FORM OF AIRFRAME ICING

Freezing of supercooled drops

The most important factor for ice accretion on aircraft is the freezing of supercooled drops – either cloud particles or raindrops – following impact with the aircraft. It was seen in Section 31 that supercooling of the water droplets in cloud is of common occurrence. A certain amount of heat is required to melt a given mass of ice without change of temperature and the same amount is liberated when freezing takes place. This is the latent heat of fusion; its value is approximately 80 calories per gramme of water or ice, a calorie being the amount of heat required to raise the temperature of 1 gramme of water by 1 degC. When a supercooled drop freezes, the latent heat liberated tends to raise the temperature of the drop. If the whole drop were to freeze instantaneously, the heat released would, unless the initial temperature were below —80°C, suffice to bring the temperature above 0°C; this would be contradictory, since ice cannot exist with a temperature above 0°C. In fact only a portion of the drop freezes instantaneously, not more than enough to raise the temperature to 0°C. Thus if a drop has a mass m grammes and the mass m' freezes, the latent heat released is 80 m' calories. If the initial temperature is —t°C, the heat absorbed when the temperature rises to 0°C is approximately mt calories, the specific heat being taken as unity. If these two quantities of heat balance, then

$$80\, m' = mt,$$

whence $m'/m = t/80$. Therefore not more than 1/80 of the drop can freeze instantaneously for every degree Celsius by which the temperature is below 0°C. Once the temperature of the partly frozen drop is raised, it begins to lose heat by evaporation and conduction to the air or objects in contact with it, so that the remainder of the drop freezes more gradually while assuming the temperature of its surroundings.

The higher the temperature of the supercooled drop, the smaller the fraction which will freeze instantaneously and the greater the amount of liquid which will freeze progressively.

Temperature

Spontaneous freezing of supercooled drops in the free atmosphere is determined partly by the temperature and size of the drop and partly by other factors; experiment shows that the average temperature of spontaneous freezing decreases with the size of the drop. Consequently as the temperature falls the larger drops are likely to freeze first, while at lower temperatures only the smallest drops will remain liquid until freezing becomes general at about —40°C. Further, we have seen that the higher the temperature of a supercooled drop, the greater the fraction which remains liquid for a time after impact with an aircraft. The liquid portion starts to flow over the airframe and so increases the available area from which the latent heat of fusion can be dissipated to the environment. Thus the higher the (sub-freezing) temperature, the more it favours the formation of clear ice and permits the accretion to spread back from the leading edges, while lower temperatures tend to the formation of rime concentrated near the leading edges. The form of ice accretion cannot however be simply related to temperature, since much depends on other factors, particularly the concentration of liquid water and the size of the drops encountered.

Kinetic heating

This arises in two ways. When an aircraft is in motion, the air follows the streamlines round the component parts of the airframe. The air pressure on the surface of the aircraft varies from place to place, being greatest at stagnation points such as the leading edges where it exceeds the static pressure, and least on the upper surface of the wing where it is less than the static pressure. The local airflow is least where the pressure is greatest, and greatest where the pressure is least. At the stagnation points the airstream is compressed and heated adiabatically, but there is expansion and fall of temperature where the pressure is less than the static pressure. There is another effect due to friction between the airframe and the air which generates heat at all parts of the airframe except the stagnation points. This frictional heating is greatest where the relative motion is greatest and so where the heating due to compression is least. The compressional and frictional heating together are known as kinetic heating. The amount of the heating varies over the surface of the aircraft; it is a minimum on the upper surface of the wing and a maximum at the stagnation points. Conduction of heat through the airframe, particularly if metal, tends to smooth out the temperature differences. The increase of temperature in clear air is small on slow aircraft – about 1 degC at 100 knots on a leading edge – but increases in proportion to the square of the speed of the aircraft. For a true airspeed of 500 knots the effect is in the neighbourhood of 25 degC. But in icing conditions, relevant surfaces of an aircraft may be wet and warming above ambient temperature results in evaporation. This in turn leads to cooling as latent heat of evaporation (600 calories per gramme of water evaporated) is taken away, mostly from the wet surfaces. In this way the effects of kinetic heating in icing conditions may be partly or even largely, offset. It may be noted that while a rise of airframe temperature to above 0°C would prevent ice accretion, a rise to a value below 0°C would be likely to increase the chance and the severity of icing. In any case, the indicated air temperature, which includes the effects of kinetic heating and latent heat of vaporization at the thermometer element, is probably a better guide to the likelihood of ice accretion than is the true air temperature (see also Section 51).

Concentration of liquid water

The effect of high concentration of liquid water is similar to that of high temperature (but still below 0°C): there is more latent heat to be dissipated to the air before freezing can take place of all the water impinging on the airframe. In this process the liquid water spreads over a large area with the formation of clear ice.

Airspeed

An increase of airspeed implies an increased rate of catch of supercooled drops and so has much the same effect as an increased concentration of liquid water in the cloud.

Size of supercooled drops

The smallest supercooled drops tend to freeze immediately on striking the airframe; the latent heat of fusion is quickly removed by the airstream and there is little or no spreading of the drop before freezing is complete. At the same time air is enclosed between the ice particles, so that the accretion takes the form of rime concentrated near the leading edges. On the other hand, large drops are accompanied by a spreading of water over the airframe while the latent heat is being dissipated, so that freezing takes place more slowly and tends to be in the form of clear ice. Drops of moderate size can produce results intermediate between these two, although the effect of drop size is not wholly separable from that of concentration of liquid water.

Severity of ice accretion

The severity of icing is defined as the rate of accumulation of ice by weight per unit area per unit time. Among the meteorological factors determining this rate are the amount of liquid water present and the size of the droplets, These characteristics are not the same throughout a particular cloud, even at one level. A cloud containing both liquid drops and ice crystals may have patches in which water drops predominate and others in which ice particles predominate. Subject to these variations from cloud to cloud and within clouds, icing will tend to be severe when the temperature is not far below 0°C, when the cloud droplets are large, and, in convective cloud, when the cloud-base temperature (Section 48) is high. Though the likelihood of heavy accretion falls off as temperature decreases this must not be taken to imply that the intensity diminishes with height in any one cloud.

An analysis of reports of ice accretion shows a preponderance of occasions at temperatures above about —10°C and indicates that the frequency diminishes rapidly when temperature falls below about —20°C; but on occasions icing has been reported at temperatures below —40°C.

48. ICE ACCRETION AND CLOUD TYPE

Convective cloud

Observations show that at temperatures down to about —20°C cumulus clouds nearly always consist of water drops. There have been very few observations at lower temperatures but the occurrence of ice accretion indicates that water droplets must exist in cumulus clouds at lower temperatures. Cumulonimbus clouds have a cellular structure and while some cells are in the growing stage others may be decaying, so the composition of the cloud varies considerably at the same level. As the growing cells mature the liquid water content is continuously diminishing with time because ice crystals grow more rapidly than water droplets in a mixed cloud. Thus there is more liquid water in newly developed parts of the cloud than in mature parts. In general, liquid droplets predominate down to about —15°C and either liquid drops or ice crystals may predominate between —15°C and —30°C.

In convective cloud the following rules may be generally accepted:

(i) At temperatures below —40°C the chance of icing is small.

(ii) At heights where temperatures are between about —20°C and —40°C the chance of moderate or severe icing is small except in newly developed convective cloud, but light icing is possible.

(iii) At heights where the temperature is between —20°C and 0°C the rate of icing may be severe over a substantial depth of cloud for a wide range of cloud-base temperatures.

Effect of temperature of cloud base. The cloud-base temperature has an important effect on the risk and severity of ice accretion in convective cloud because of its influence on the free water content throughout the whole cloud. At the cloud base the air is just saturated, and the higher the temperature, the higher is the vapour content. As the air ascends, vapour condenses and much of the condensed water is carried upwards. It is found that the content of liquid water, when expressed as mass per unit volume of air, shows little variation with height over the greater part of the depth of the cloud, the increasing condensation as the air cools being for the most part offset by the expanding volume through which the water drops are distributed. Consequently the free water content at any given level in a convection cloud increases with the temperature of the cloud base. Further, with increased concentration of supercooled water, ice accretion is likely to be more severe and as already seen is more likely to be in the form of glaze ice. The base of convection cloud occurs at much the same height in all latitudes, so that the temperature at the base is much greater in the tropics than it is in temperate latitudes, with the result that the liquid water content of convection clouds in the tropics is often about double that in temperate latitudes. Ice accretion at a given height above the level of 0°C in convection clouds is therefore likely to be noticeably more severe in tropical than in temperate latitudes. Similarly, in temperate latitudes the accretion in such cloud is likely to be more severe in summer than in winter.

Layer cloud

At temperatures down to about —15°C stratocumulus cloud usually consists entirely or predominantly of liquid drops. Altocumulus cloud usually consists entirely of liquid drops in temperature ranges down to —10°C; at lower temperatures, down to —30°C, ice crystals may be present but are normally outnumbered by liquid water drops. Stratus cloud usually contains only water drops.

In general the severity of icing in layer clouds of, say, 3000 feet thickness with tops at 850 millibars is moderate when the temperature at the top of the cloud is between 0°C and —10°C and light when the temperature at the top is less than —10°C, but occasional high rates of icing may be encountered in layer clouds at lower temperatures. Stratocumulus layers, especially over the sea in winter, are often formed by convection and in this case, and in stratocumulus formed by the spreading out of cumulus, the liquid water content will probably be greater than in stratocumulus formed by turbulent mixing and the severity of icing may be greater.

Altostratus and nimbostratus are usually formed by slow ascent of a large mass of air over an extensive area. The vertical extent of such clouds may be many thousands of feet. Some part of extensive clouds of this type is likely to contain supercooled water drops, and is thus a potential icing region if within the temperature range 0°C to —15°C. If the clouds are associated with active fronts, and particularly if there is an orographic effect due to the proximity of hills or mountains, the chance of severe icing is much increased, and icing may be encountered at temperatures lower than usual. Severe icing in these conditions has been reported at temperatures as low as —20°C to —25°C.

Cirrus cloud

Cirrus clouds are usually composed of ice crystals which do not constitute an icing hazard to aircraft.

Orographic cloud

In clouds formed when air is forced to rise over hills and mountains, entrainment of dry air is unlikely and the continued forced ascent may lead to further condensation. The continuous upward motion will generally mean a greater retention of water in the cloud, and because of this icing is likely to be more severe in clouds over hills and mountains than in similar clouds away from high ground.

Stable air may become unstable when lifted or a weak front may easily become more active when passing over high ground. When stable air is lifted orographically the 0°C isotherm is lowered and icing may be experienced at a lower level than in the same air mass over level ground.

The importance of the increase in the severity of icing in cloud subject to orographic lift cannot be over emphasized.

Precipitation

Raindrops and drizzle drops lying in the path of an aircraft are practically all captured and in freezing conditions the ice deposit will consist of clear ice spreading over a large area of the aircraft surface. Freezing rain occurs occasionally ahead of a warm front.

Raindrops carried up within a cumulonimbus cloud in conditions of high instability may produce an area of abnormally high water content which, in temperatures below 0°C, could be the principal cause of possible severe clear-ice formation.

49. EFFECTS OF AIRFRAME ICING ON PERFORMANCE

The various parts of the airframe are affected in different ways by ice formation, both with regard to the types of ice likely to form and with regard to the effect of the accretion. The effects on performance are described below.

Aerodynamic effects. When ice formation occurs on the leading edges of the airframes, the pattern of the airflow becomes modified round the affected part. This leads to an increase in drag, a decrease in lift and perhaps to buffeting of the tail. Ice accretion on the leading edges of the fin and rudder and other movable parts may interfere with the airflow to such an extent that control is seriously affected.

As an object moves through air containing water drops, it catches only a fraction of the water which is present in the path swept out; this fraction varies with the shape of the object and is found for example to be greater for a thin wing than for a thick wing, other things being equal. It does not follow that a greater total weight of ice is collected by the thin wing, since the path swept out has a smaller cross-section. On the other hand, a small deposit on a thin wing may cause greater aerodynamic disturbance than a similar deposit on a thick wing. This dependence on shape explains why thin objects such as aerials, struts, the leading edges of propellers, etc., are more liable to icing than are the more bluff parts of the airframe such as the nose of the fuselage.

The extremities of the propeller blades have a much higher speed than other parts of the aircraft and for this reason too one would expect this component to be susceptible to icing, but there is some protection from icing by kinetic heating.

The aerodynamic effects of ice accretion are of course not confined to disturbance at the leading edges. Ice forming on other parts of the wing or fuselage may lead to a considerable increase in drag. Ice formation under the wing may be particularly dangerous in that it is normally out of sight and its existence may be inferred only from a change in the performance of the aircraft.

Effect of weight of ice. The effect of the accumulated weight of ice is not generally of primary importance. An unequal distribution of ice may have serious effects, however, particularly when it occurs on the propeller, for with this component the lack of balance when part of the ice breaks away may lead to serious vibration. This type of hazard may also occur in connection with aerial masts, exposed balance weights and the arms actuating control surfaces, and may in extreme cases lead to fracture.

Effect on instruments. Any small excrescence on an aircraft is liable to gather ice; the pitot static system and aerial masts are examples. Apart from the risk of vibration and fracture already mentioned, the effect may be to reduce the efficiency of the part affected, leading to serious errors in measurement of airspeed and to loss of communication facilities.

Effect on control surfaces. Normally there is a gap between the forward edge of a control surface and the fixed surface ahead of it. In some positions of the controls, the air flows through the gap and ice may accumulate not only on the leading edge of the movable surface but in the gap itself, possibly sufficient to jam the control. The risk is greater on a small aircraft than on a large one, since on the former the gap is smaller and the movable part thinner, leading to a greater rate of accumulation.

Miscellaneous effects. Other common effects include the formation of an ice coating on windscreens, canopies and astrodomes so that vision is obstructed. This may be caused by flight in icing cloud, or may be due to hoar-frost formation in clear air when the very sudden occurrence may be disconcerting.

50. ENGINE ICING

Piston engines

Ice formation in the air intake and induction system of a reciprocating engine results in loss of engine power due to obstruction of the air passages and to disturbance of the fuel metering, while movable parts may become inoperative. Engine icing is not only a low temperature phenomenon but may occur at air temperatures well above 0°C and in clear air as well as in cloud and precipitation.

Today the hazard has been considerably reduced by the provision of sufficient heating devices, sheltered air intakes, and improved carburettor design.

The ice accretion may be produced in the same circumstances and manner in which airframe icing is produced. Ice formation in the carburettor can also occur when flight is taking place in clear air even when the ambient temperature is well above 0°C. Considerable cooling can occur in the carburettor, resulting in a temperature below the dew-point of the air and below 0°C causing ice formation on the walls and internal components. The cooling is the result of two factors, one the evaporation of petrol from surfaces wet with petrol, the other the acceleration of air through the carburettor, which causes a local reduction of pressure and results in adiabatic cooling. The combined effect can reduce the temperature of the air within the carburettor by as much as 25 degC. The effect varies with the throttle settings. The intensity of ice will be closely related to the water content of the air and will be greatest if liquid drops are initially present, as when flying in rain or cloud or when running the engine in fog. To counter carburettor icing, heat is applied in some way to maintain the internal temperature above 0°C.

The direct-injection type of carburettor fitted in many modern piston engines is rarely subject to icing.

Turbine and jet engines

The intakes of turbine and jet engines are subject to icing in the same way as the airframe when flight is taking place in supercooled droplet cloud.

The susceptible parts are the rim of the intake where the radius of curvature may be small, any struts across the intake, and the vanes in the early stages of the compressor. Thereafter temperatures are usually too high for icing to be a problem, although ice breaking away from parts near to the entrance may cause damage within the engine.

Generally speaking, engine icing will be directly proportional to the rate of airflow through the engine and thus to the number of engine revolutions per minute. It is frequently found that the rate of icing may be reduced by decreasing the revolutions per minute.

When the jet engine is operating at high revolutions during flight at low speeds, as when taking off and landing, or whilst stationary, as in running up, the pressure within the intake is much less than the pressure outside. The consequent adiabatic expansion in the intake causes a drop in temperature of as much as 5 degC. If the clear indrawn air is moist and the temperature is near 0°C, prolonged operation may result in condensation and ice formation when this would not occur on the airframe. This effect may accentuate the icing which would normally be expected when the flight is in icing cloud, or when the aircraft is taking off or landing in freezing fog. Usually jet engines ice up in flight only under conditions which might be expected to produce airframe icing.

The intensities of icing on the airframe and in the engine may be different since the airframe icing rate depends on the airspeed, whilst engine icing depends in addition on the rate of engine revolutions. At high speeds the engine tends to be supplied with more air than it needs and there is a ram effect, whereas at lower speeds, below about 250 knots, air is sucked in. Because of the ram effect at the higher speeds some of the air is deflected round the intake, but the inertia of the water drops results in a higher water concentration within the intake and the icing rate increases markedly with increase of airspeed above 250 knots. At speeds where the air is being sucked in (below about 250 knots) the water concentration of the air entering the intake remains virtually the same as in the free air, so that the engine icing rate tends to be constant with decreasing airspeed, whereas the airframe is likely to be showing a marked decrease of icing rate with decreasing airspeed.

51. PROCEDURES IN RELATION TO ICING RISKS

Pre-flight procedure

If an aircraft is left in the open on cold nights it should be protected from hoar frost, rime, snow or rain ice. Any frozen deposit which has formed should be carefully and completely removed before take-off.

A meteorological aviation forecast for the appropriate route and time should be obtained before take-off. Route forecasts usually contain information giving the height of the 0°C isotherm and, when cloud is expected at levels where the temperature is favourable for icing, the estimated icing conditions are indicated by one of the following terms: 'light icing' or 'icing index low', 'moderate icing' or 'icing index moderate', 'severe icing' or 'icing index high'. At certain airports the routine forecast contains a series of charts and landing forecasts, and in this case the height of the 0°C isotherm is included with spot-wind and temperature data. The icing risk is shown on the significant-weather chart, and where icing is expected this is indicated by a symbol and the cloud layers are given for the levels between which icing is expected.

Even if the forecast indicates a neglible risk, the pilot should nevertheless be prepared for icing, particularly if clouds are likely to be encountered at temperatures near or below 0°C. It has been seen that the likelihood of icing depends in part upon the characteristics of the aircraft, but the forecaster cannot take account of unusual features in that respect. The pilot must therefore consider his own particular aircraft when deciding in the light of the forecast whether or not its performance is likely to be seriously impaired during flight. It must be remembered that a situation which presents no difficulty in normal conditions may become dangerous if the aircraft is handicapped by ice accretion with a possible reduction of lift and engine power and perhaps loss of control.

In-flight procedure

The fullest use should be made of any indication available in flight regarding the likelihood or type of ice accretion. The possible indications are explained below under the relevant headings.

Temperature. The indicated temperature given by the aircraft thermometer includes the effects of kinetic heating at the position of the thermometer element, but we have seen that the temperature is not uniform over the surface of the airframe. In clear air, the free-air temperature is obtained by applying a correction to the indicated temperature (Section 95), but in cloud or precipitation the required correction is not known precisely; moreover the reading may be falsified by ice on the thermometer itself. It is clear that implicit reliance should not be placed on either the indicated or the corrected temperature when in the neighbourhood of the 0°C isotherm. It should be remembered, too, that engine icing can occur with air temperature well above 0°C.

Radar echoes. Some aircraft are fitted with radar equipment which can be used to detect liquid or frozen water in the atmosphere. It appears that a radar echo of meteorological origin is likely to indicate a region where ice accretion is probable if the temperature is below 0°C. With experience it is possible to recognize the kind of cloud associated with the various types of echo which appear on the radar screen. A clear-cut echo of high intensity frequently indicates a cloud of the convective type. A line of such echoes on the screen may indicate a line of cumulus or cumulonimbus clouds which may be embedded in other cloud not producing an echo. Warm-front clouds (Chapter 16) usually give echoes which are less well defined and less intense. It must be emphasized that absence of an echo does not imply freedom from icing. The intensity of the echo is dependent upon the concentration and drop size of the liquid or frozen water in the cloud and if these are inadequate there will be no echo although ice accretion may still occur if the temperature is suitable. The aim should be to avoid the sub-freezing regions giving radar echoes and then the worst icing conditions will probably be avoided also.

Optical phenomena. It is sometimes doubtful whether a particular cloud consists of ice particles or water drops. In this connection it is useful to remember that a halo of 22° radius (Section 78) is an indication of ice crystals, though absence of a halo does not necessarily indicate that the cloud consists of water drops. On the other hand a corona indicates a cloud of water droplets. Also a water cloud may be recognized from above by a 'glory', a system of coloured rings surrounding the shadow of the aircraft.

It might be thought that the visibility in a cloud should give a guide to the amount of water per unit volume and hence to the risk of icing at appropriate temperatures. This is a very unsound criterion and should not be used.

Type of ice. If ice is seen to start forming on the airframe, some information can be obtained from the type of ice. If the deposit is opaque, appears light in texture and grows out into the airstream, the conditions are probably not severe. If clear ice is observed to be forming, the rate of growth is likely to be rapid and immediate action is desirable.

Melting and evaporation of ice. Any ice formed might be expected to melt if the aircraft is flown to a region or level where the ambient temperature is above 0°C. The critical temperature however is not necessarily 0°C, because the ice evaporates slowly and acts as a wet bulb (or ice bulb), so that before melting can occur it is necessary that the wet-bulb temperature of the air should be above 0°C. The effective wet-bulb temperature is that given by the uncorrected reading of a suitably placed wet-bulb thermometer which thus takes the kinetic heating into account. (For a similar reason, snow or hail falling through unsaturated air does not start to melt immediately on passing through the 0°C isotherm but continues to fall unchanged until the (corrected) wet-bulb temperature is above 0°C.) Although direct evaporation of airframe icing also takes place at temperatures below 0°C, this is a very slow process. Successive traverses of cloud may therefore lead to a progressive accumulation of ice, and flight through clear air at only a few degrees above 0°C provides no guarantee that the ice will disperse.

Some possible procedures in relation to various conditions of airframe icing are discussed in the following paragraphs.

Procedure in relation to clouds

Altocumulus and thin altostratus. While altostratus cloud usually consists of ice crystals, the possiblity of supercooled drops must not be ignored. Altocumulus usually has water drops predominating, at least down to —30°C. In both types the water content and the rate of ice accretion are likely to be low. Prolonged flight in such clouds may cause appreciable accretion but there should seldom be difficulty in getting out of the clouds, which are generally shallow, by change of height.

Stratocumulus. This is essentially a layer cloud in the lower part of the atmosphere. The vertical extent seldom exceeds 3000 feet so that it will usually be easy to climb above it. It is a water cloud of comparatively low water content so that the ice accretion is usually light or moderate and only very occasionally severe. It is unlikely to be serious unless flight in the cloud is prolonged. If it is known that a stratocumulus cloud has been either convectively or orographically formed, then a higher water content and a higher rate of icing should be expected than in stratocumulus formed by turbulence. There is also a possiblity that cumulus clouds may penetrate the layer of stratocumulus; icing would then be intensified during flight through the embedded cumulus.

Convection clouds. Ice accretion is liable to be severe in large cumulus and cumulonimbus. A reduction in the intensity of icing cannot be expected as a result of increasing height unless the temperature eventually reached is less than about —20°C and even then it is by no means certain that serious icing would not occur; it is not until the temperature falls to about —40°C that the icing risk becomes negligible. Hence if icing occurs during flight through large convection clouds, height should be reduced, subject to maintaining adequate ground clearance, to a level where the temperature is above 0°C, or flight should be continued without deliberate change of height or heading until emerging from the cloud. Since the lateral dimensions of such clouds seldom exceed a few miles, the time occupied in a traverse is brief. Alternatively the clouds can often be avoided altogether, which is undoubtedly the safest course not only with respect to ice accretion but also in regard to other risks (Section 43).

A proper appreciation of the remainder of this section requires some knowledge of the structure of frontal cloud systems. These are described in Chapter 16 to which reference should be made if necessary.

On occasions, large convection clouds amalgamate along fronts to form a belt possibly hundreds of miles long and in extreme cases a few tens of miles wide. An increase in the width of the belt is usually accompanied by a decrease in the intensity of convection, but icing may be moderate or severe over a considerable distance. If flight below such a belt is impracticable, the use of radar, as already described, may be of considerable assistance in avoiding the worst places. A belt of this kind should be crossed at right angles. In this connection the importance of obtaining pre-flight advice about the location and orientation of fronts is emphasized. Extensive areas of convection cloud also occur at times in unstable air over mountainous country. In such cases the practicability of flying high enough to avoid most of the clouds should be considered.

Warm-front cloud. A characteristic of warm fronts is the extensive and almost horizontal layer of cloud. The base of the cloud (nimbostratus) is low near the front, where there is frequently steady rain, snow or sleet. Ahead of the front the base of the cloud becomes higher and the whole cloud sheet thins out into a layer of altostratus or cirrostratus. The main cloud sheet near the front extends from a few hundred feet above the surface to at least 5000 feet and frequently above 10000 feet, while some part of the frontal cloud system certainly contains supercooled drops and is thus a potential icing region.

If it is known that the air temperature is well above 0°C in the lower layers, it is probably best for the low-flying aircraft to cross the front at a level where the temperature is above 0°C, if terrain permits. If there is a large area of uniform frontal precipitation, serious icing should be anticipated within cloud in the temperature range 0°C to —15°C, i.e. in a layer perhaps 10000 feet in depth. Moreover, supercooled rain may exist beneath the cloud (see below). At temperatures less than —15°C there may be snow and the cloud will consist largely of ice crystals. The proper action then is to fly at a level where the temperature is either above 0°C or below —15°C, whichever is the more practicable in the prevailing circumstances. On the warm side of the front uniform precipitation is unlikely and the cloud may consist of water drops for a considerable height above the 0°C level. Flight should then be made if possible at a height below the 0°C level unless it is certain that flight can be made above the cloud layer.

If ice formation starts during flight in a cloud layer, descent may well prove better than ascent since the performance of the aircraft will already be affected to some extent by the ice and putting the aircraft into a climbing attitude will increase the risk of icing under the wing where it cannot be seen. It should be realized too that cumulonimbus clouds are occasionally embedded in warm-front systems.

High ground near a warm front is particularly dangerous. Not only does the cloud obscure the hill tops but there is an increase in the free-water content of the air when forced to rise over the hills (Section 48).

Cold-front cloud. At cold fronts the clouds are not usually extensive or unbroken and accordingly there is a better chance of avoiding the worst icing regions. Such clouds are, however, of the convection type with strong ascending currents so that the free-water content and rate of icing are liable to be higher than in warm-front clouds; embedded cumulonimbus clouds are a fairly common occurrence at the more active cold fronts.

The activity of a front is increased on passage over high ground and the possibility of large convective clouds and serious icing is increased.

Occlusions. As the characteristics of an occlusion cover a range of conditions intermediate between those appropriate to warm and to cold fronts, the icing characteristics of an occlusion may tend towards those of either type of front.

Procedure in relation to supercooled rain

Supercooled rain occurs beneath warm fronts and occlusions and occasionally beneath cold fronts. When present, there is necessarily a warmer layer above, in which the temperature exceeds 0° C. In low flight the best procedure for avoiding icing is to climb into the warm layer which will usually be found at no great height, although this should not be attempted unless the pilot is reasonably sure of the details of the meteorological situation. Descent to a layer where temperature is above 0°C, if one exists, may be dangerous because of the low cloud base in the vicinity of the front. As with the frontal cases discussed above, the belt of frontal cloud and rain should be crossed at right angles so as to give the shortest traverse through the icing region. A particularly dangerous procedure would be to fly parallel to the front in freezing rain, since this might result in a heavy accumulation of clear ice.

CHAPTER 9

VISIBILITY

52. INTRODUCTION

Although modern developments in radio and radar have made it possible, in principle, to fly without ever looking outside the cockpit, visibility remains a factor of great importance in all ordinary aircraft operations both civil and military. A pilot's interest in visibility arises because he wants to know how far off he will be able to see various things – landmarks, targets, obstructions, beacons, both in the air and on the ground. Unfortunately a meteorologist cannot usually answer the pilot's questions fully and accurately because the range at which an object can be seen depends on many things besides the state of the atmosphere; it depends on the size, shape and colour of the object, on its illumination and background, on the pilot's keenness of vision, on the speed with which he is moving and on the transparency of his windscreen. Even if all these things were known and allowed for, there remains the fact that the state of the atmosphere often changes rapidly with height – a dense layer of fog at ground level may be associated with clear air at heights of a few hundred feet – and from place to place on the ground. It must therefore be emphasized that the meteorologist's reports and forecasts of visibility normally apply to horizontal visibility near the surface for particular localities, and in this form they often give little guidance about the visibility to be expected from aircraft in flight.

53. SIGNIFICANCE OF VISIBILITY

Visibility at ground level

By day, the visibility in a particular direction is defined as the greatest distance in that direction at which a person of normal sight can distinguish and identify an object for what it is known to be, under normal conditions of daylight illumination.

Objects for visibility points should be dark in colour and so placed as to be viewed against the sky or other light background; they should subtend an angle of not more than five degrees in width and not less than half a degree in both height and width. When the visibility varies with direction the practice is to report the lowest value, but a supplementary report of the maximum visibility and its direction may also be made. The range at which any ordinary object can be seen under normal conditions of illumination will not differ seriously from the visibility distance unless the object is very small or blends well with its background, but in special conditions of illumination the visual range may differ appreciably from the visibility distance. For example, the visual range is greatly increased if the object is reflecting sunlight strongly. Again when haze particles are producing glare in sunlight, visibility is better away from than towards the sun; on the other hand objects can be seen at night more readily if viewed towards the moon than away from it.

By night, it is much more difficult to define visibility in a way satisfactory to all concerned because different types of lights and objects have very different ranges and no one figure for 'visibility' can satisfy all users. For meteorological purposes it is convenient if a given figure for visibility corresponds to the same state of the atmosphere by night as by day; the meteorologist, therefore, determines the visibility at night by measuring the degree of transparency of the atmosphere (by means of observations on specially selected lights) and thence deducing the visibility that would be observed if it were daylight. The range at which any particular light can be seen at night can be deduced from this visibility but is not, of course, necessarily or even usually equal to it. When the visibility is one mile, for example, a light of one candle-power will be visible for half a mile, a light of 20 candle-power will be visible for one mile, and a light of 1 000 000 candle-power (such as a high-intensity approach light) will be visible at over three miles.

Runway visual range

For general information on the state of transparency of the atmosphere, visibility as defined above is the most suitable quantity to use. Moreover, as already explained, the visual range of lights by night may differ considerably from the nominal visibility, largely because of the brightness of the lights. The same is true, although to a lesser extent, of the visual range of objects by day. To determine as closely as possible the conditions which would be experienced by a pilot on landing or on take-off, observations are made of the visual range of those objects or lights provided for the guidance of the pilot at many airfields in the United Kingdom. Such observations are made only when the ordinary visibility is less than 1100 metres. In daylight the objects are markers placed alongside the runway, while by night runway lights are used. In certain circumstances the range of the lights by day is greater than that of the markers, then the range of the runway lights is reported. Since the lights are flush with the surface of the runway, it is not usually possible to determine the runway-light range from the ground; observations are therefore made by means of subsidiary lights and are finally converted into runway-light range. In any case the observations of runway visual range are made from a point close to the runway, so that the results should correspond closely with the range actually experienced by the pilot at touch-down or take-off. (See *U.K. Air Pilot.*)

Air-to-ground visibility

It has already been pointed out that normal meteorological measurements of visibility are made horizontally at ground level and thus give little information about the visibility from points above the ground. Even from heights as low as 100 feet the 'slant visibility' may be much greater (when above a shallow layer of fog) or much less (when in a low cloud layer) than the horizontal visibility at ground level. A common case is illustrated in Fig. 58 which represents a layer of fog covering an aerodrome with clear air above. From positions well above the fog, objects on the aerodrome can often be seen relatively clearly through a fairly small thickness of fog, but on descending to the level of the fog it may be impossible to see the aerodrome at all since a greater thickness of fog has to be penetrated. Moreover, glare caused by diffuse reflection of sunlight from the top of a layer of fog or haze can seriously reduce the air-to-ground visibility. For various reasons, then, there will on occasions be little relation between air-to-ground visibility and the normal visibility; in particular, it is important not to be deceived into thinking a landing easy because the aerodrome is clearly visible from above.

Fig. 59 shows that in the circumstances illustrated the greatest horizontal distance at which the ground is visible from an aircraft increases steadily as the height of the aircraft increases; this applies to all cases where the poor visibility is due to a layer of mist or haze below the aircraft. However, at higher altitudes ground objects

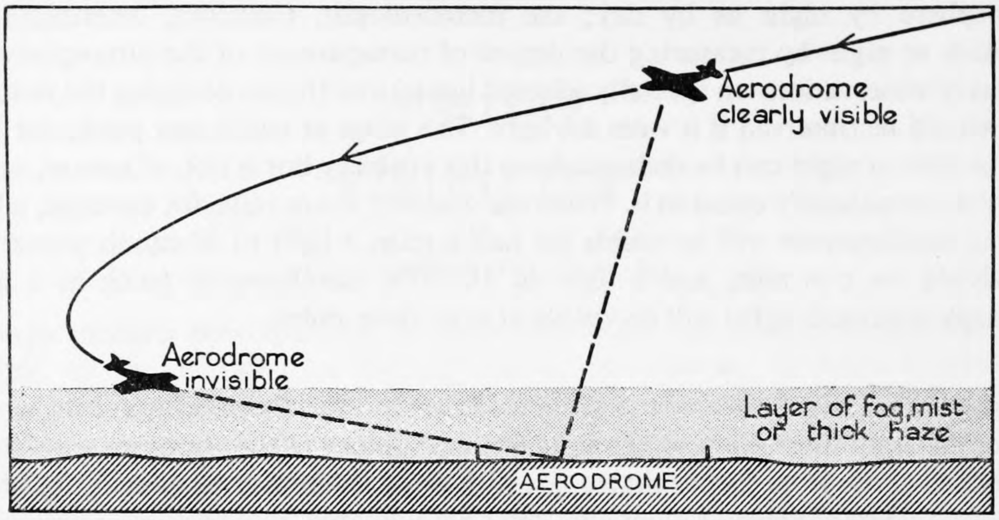


FIG. 58. *Obscuring of an aerodrome by surface fog as an aircraft descends to land*

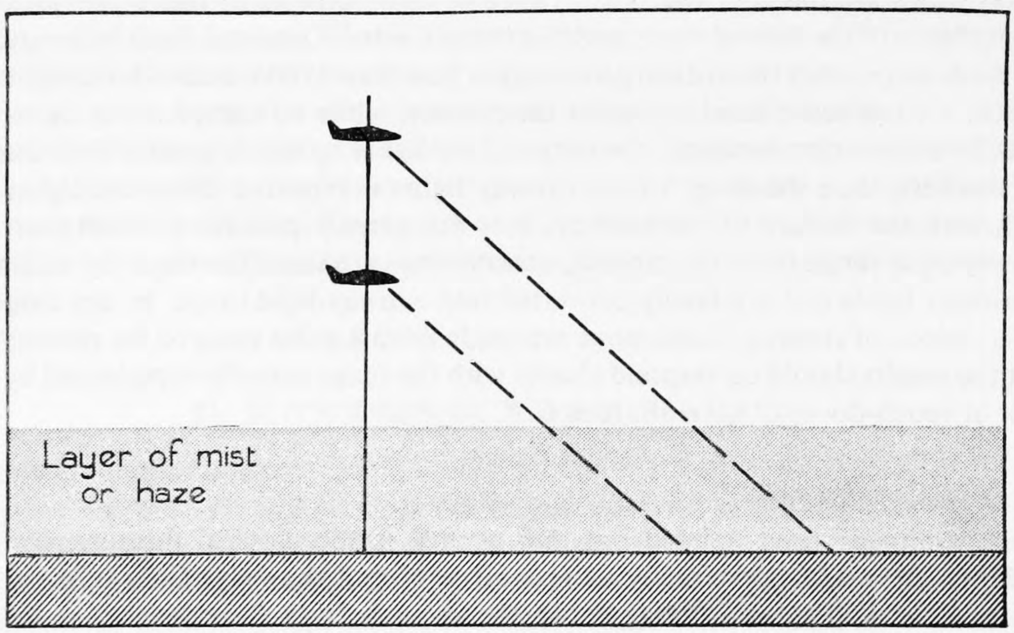


FIG. 59. *Horizontal range of visibility from an aircraft flying above a haze layer on the ground*

are less easily seen because of diminished size, and in practice on any particular occasion there is an optimum altitude which gives the greatest range concurrent with the identification of ground objects. Sometimes it is possible to estimate the depth and opacity of the layer of mist or fog from ground observations and hence to deduce the ground range from any height, but the estimates can seldom be precise or reliable.

Air-to-air visibility

As with objects at ground level, the ability to see an object at higher levels (whether in a horizontal direction or not) depends not only on the state of the atmosphere but also on the state of the object being observed and on the observer himself. Thus the ability to see one aircraft from another depends among other things on the size, colour, illumination, background and speed of the remote aircraft, on the sight of the observer, the bearing of his line of sight relative to that of the sun or moon and on the transparency of the cockpit windows. The transparency of the atmosphere depends chiefly on the presence or absence of clouds. In cloud, visibility is almost always very low (though at great heights more tenuous clouds, equivalent to a thin mist, may be encountered); outside cloud and precipitation, visibility is normally good though layers of dust or haze may reduce it to one or two kilometres. In general, however, factors such as the size and speed of the object and the position of the sun in relation to object and observer play the chief part in deciding the range at which the object can be located visually.

At great heights, as in the stratosphere, cloud is rare and dust and haze are almost unknown but it does not follow that conditions for seeing an object are always good. The fact that the apparent brightness of the sun increases with the height of the observer while that of the sky decreases leads to greater dazzle than that which occurs at lower heights. Moreover, in the absence of objects on which to focus, the eye adjusts itself to a focus midway between distant and near vision and the location of an object at great heights is less easy than at small heights.

54. CAUSES OF ATMOSPHERIC OBSCURITY

The previous sections have shown how the distance at which an object can be seen depends on the position and characteristics of the observer and the object, as well as on the obscurity of the atmosphere between the two. In the remaining sections of this chapter we shall deal with the strictly meteorological aspects of visibility – the conditions determining the obscurity of the atmosphere at a given place and time.

The main causes of atmospheric obscurity may be classified as follows:

(i) fog and mist,
(ii) cloud and precipitation,
(iii) wind-blown spray from the sea,
(iv) smoke, and
(v) dust and sand.

The obscurity in (i) and (iii) is caused usually by water droplets, in (ii) by water droplets or ice particles or both together, and in (iv) and (v) by solid impurities. The invisible water vapour does not contribute to atmospheric obscurity, nor do the microscopic condensation nuclei referred to in Section 31.

55. FOG AND MIST

By definition, fog is composed of a suspension in the air of very small water droplets (ice crystals in ice fog) reducing the horizontal visibility at the earth's surface to less than 1000 metres. In mist, which is otherwise similarly defined, the visibility does not fall below 1000 metres. Relative humidity is generally near 100 per cent in fog and is at least 95 per cent but generally less than 100 per cent in mist.

Fog composed of water droplets (or possibly of ice crystals) may be described as a cloud on the surface. Over high ground fog may indeed be merely one or other of the usual cloud types, requiring adiabatic ascent for its formation; the hills may protrude into a sheet of cloud or may themselves be the cause of local orographic cloud. Apart from such cases, condensation in the great majority of fogs is produced as a result of a relatively cold underlying surface. Two distinct types come within this category:

(i) Radiation fog caused primarily by loss of heat from the ground by radiation at night.
(ii) Advection fog caused by the transport of moist air over a relatively colder surface.

There is no essential difference between these types of fog except in the mode of cooling – in (i) the air remains stationary or slow moving while the ground cools, in (ii) the air is transported to a place of lower surface temperature. In addition, there are two other types of water-drop fog:

(i) Steaming fog caused by evaporation into cold air lying over warmer water.
(ii) Frontal fog caused by precipitation or low cloud.

While fog particles are usually water drops, they can be ice crystals. Such 'ice fogs' are liable to occur in the Arctic (and no doubt also in the Antarctic) when the temperature is less than about —20°C, the wind is light and other conditions are favourable.

56. RADIATION FOG

The development of radiation fog depends on the cooling of the ground at night (Section 12). It is therefore exclusively a land fog, although having formed it may drift over coastal waters. There are occasions when radiation fog occurs before saturation is reached, even with relative humidity as low as 80 per cent. These cases are explained in part by the presence of hygroscopic nuclei and in part by smoke pollution. Generally the air needs to cool at least to the dew-point before fog will form. The immediate cause of the cooling is the colder ground; the air in contact with it is cooled by conduction and the cooling is spread upwards mainly by turbulent mixing and to some extent perhaps by radiation. The initial cooling taking place at the ground, an inversion tends to develop with the lowest temperatures on the ground; if the dew-point is reached first on the surface itself, then moisture extracted from the air is deposited as dew. In quiet conditions with little turbulence the cooling extends slowly upwards and the ground becomes much colder than the air at the height of a few feet. Just as condensation of water takes place on the cold walls of a room when the air is humid without any visible moisture in the air, so a copious deposit of dew may be formed without fog. In such cases the air above the surface is progressively dried so that its dew-point remains below the air temperature.

Take now the other extreme, strong winds and well-developed turbulence; the loss of heat from the ground by radiation may be equally rapid but the cooling is spread through a deep layer of air and temperature falls but slowly. Since turbulence ensures a lapse rate approaching the dry adiabatic, the lowest temperatures are not near the ground but towards the top of the friction layer. If therefore the dew-point is reached at all it is likely to be at a height well above the surface, giving stratus or stratocumulus cloud but not fog.

If, then, the air contains ample moisture a rather delicate adjustment between the rate of cooling and the degree of turbulence is required to ensure that condensation shall take place in the air near the surface and not only on the ground as dew or above the surface as low cloud. It is hardly to be wondered at that radiation fog is erratic in its development, affecting one locality while leaving another clear when the differences in general conditions are otherwise hardly noticeable.

The conditions favourable for radiation fog are therefore:

(i) A high relative humidity so that little cooling is required to reach saturation.

(ii) Little or no cloud so that heat is lost by radiation from the surface.

(iii) Little wind so that cooling is confined to the surface layers, but sufficient to give some turbulence. A wind of 2–8 knots near the surface is considered to be the most favourable, although in practice fog should be expected even if the wind is calm.

An inversion of temperature is sometimes stated to be a further favourable factor; this is true of smoke fog but as regards water fog it is rather an accompanying condition than a cause, both condensation and inversion being caused by surface cooling. The inversion usually appears near the ground before condensation begins, but as the fog forms the lapse becomes roughly isothermal from the ground upwards until the displaced inversion is found across the top of the fog. Loss of heat by radiation from the ground is almost completely prevented by a layer of fog, but further cooling by radiation takes place mainly from the top of the fog itself.

Dispersal of radiation fog

Generally radiation fog clears during the day as a result of incoming radiation. Morning insolation heats the fog layer, turbulent mixing assists the process, and a point is reached where the visibility improves above the fog limit. Further heating may lead to a breakdown of the inversion, and mixing with the drier air above results in a further improvement in visibility.

Dispersal of fog is complicated by changes in cloud cover. A layer of cloud spreading over early in the day may prevent fog clearance by reducing insolation, but on the other hand the fog may clear even more quickly if the loss of insolation is accompanied by a sufficient gain in downward long-wave radiation from the cloud base (see Section 12). The arrival of a cloud sheet during the night also modifies the radiation balance. Radiation from the top of the fog is reduced and the flux of heat upwards in the soil may raise the temperature of the air in contact with the ground enough to clear the fog.

An increase in wind speed aids fog dispersal either by lifting the fog into low stratus, or by mixing the surface layers with drier layers above.

Diurnal and seasonal variation

The minimum night temperature occurs on the average about dawn; the highest frequency of radiation fogs might naturally be expected at the same time, but experience shows that the maximum appears about an hour after sunrise. The slight increase in turbulence is considered to be the factor which causes existing fog to thicken or a sudden formation to occur when only dew had formed before, although the theory may not be free from objection. Whatever the cause, the fact is important, and consideration should be given to the possibility of a sudden formation of fog just after sunrise following a clear quiet night. The fog formed either during the night or in the early morning requires further heating before it will clear. In winter when the sun has little strength the clearance may be long delayed and a thick fog, by shutting out the heat of the sun, is particularly likely to persist; in summer, radiation fog is infrequent and unlikely to persist much after sunrise. Because of the long nights and the generally low land temperatures, the winter half of the year is very liable to radiation fog, although in most districts the autumn is more liable than winter, since the seasonal cooling of the land is greatest in autumn.

These broad statements may require modification before they can be taken as applying in other parts of the world. The principles are everywhere the same, but climatic differences are all-important; in some very dry climates fog is almost or entirely unknown.

Local influences

The topography and the condition of the ground are factors responsible for the local nature of many radiation fogs. When conditions are very favourable the fog becomes widespread and may cover a large area in an unbroken sheet, but usually its incidence is localized. Aggravation by smoke pollution will be considered later; it is the pure water fog with which we are now concerned. The most noticeable feature is its tendency to develop first in valleys, due partly to the katabatic drainage of cold air into the lower-lying places, and partly to the marked nocturnal fall in surface temperature in sheltered localities; the higher humidity when a stream or river flows along the valley is a further factor – although if the river is wide its higher temperature may keep the air immediately above free from fog when the surrounding banks are enveloped. Although waterlogged or moist ground is less easily cooled at night than the dry ground, the effect on humidity is generally the more important, so that fen country is notoriously foggy; similarly, fog is particularly likely when the sky clears at night after rain. Apart from moisture content there are other properties of the ground, particularly the conductivity, which account for local peculiarities.

Not infrequently observations show that fog forms first on high ground, and this is especially true of some sudden developments in the early morning. We should, however, always remember that the conditions favourable for fog and for low stratus cloud are much the same, the difference being a matter of turbulence; the fog which forms on higher ground may generally be better regarded as low cloud. When low fog clears it often passes through the stage of stratus which is sometimes then described as 'lifted fog'.

Vertical thickness

There is no precise lower limit to the vertical extent of a fog; it may be but a few feet in depth forming a ground fog, while many valley fogs are also quite shallow. In general, however, the fog is co-extensive with the layer of frictional influence; although this is usually taken as 1500 feet in average circumstances it is much less in the light winds and stable lapse rate of foggy conditions. Most radiation fogs have a depth of only a few hundred feet. The upper surface is usually sharp with clear air above; seen from above, there is little or nothing in the appearance to distinguish between a thick fog and low stratus cloud.

In meteorological reports, distinction is made between fogs through which the sky is visible and those through which it is not. The depth of fog necessary to prevent the sky being observed in daylight is roughly 300 feet.

Pressure types associated with radiation fog

Several types of pressure distribution may be associated with radiation fog but all have one feature in common, a slack pressure gradient, that is, with the isobars widely separated, so giving little wind. This condition eliminates most depressions, which are also unfavourable because of the clouded skies, but on occasions in a weak or decaying depression the slack pressure gradient and scarcity of cloud allow fog to form readily. In general, however, one associates radiation fog with quiet anticyclonic conditions, an indefinite pressure distribution, or a col. Long foggy periods in winter are invariably associated with a persistent anticyclone.

57. ADVECTION FOG

Advection fog occurs when air moves over a cooler surface of land or sea, the surface temperature being below the dew-point of the moving air.

Over the land, advection fog is particularly likely in winter after a cold spell when a supply of milder air arrives from the sea. The humidity is often already high, for the air may have become almost saturated over the sea, and further cooling by contact with the land readily produces fog and perhaps even drizzle. The conditions for fog are especially favourable during a thaw produced in this way, since melting snow maintains the temperature of the ground at 0°C, while evaporation ensures an ample supply of water vapour. In other cases the ground itself becomes gradually warmer until after a few hours its temperature rises above the dew-point and the formation of fog then ceases; thus a broad belt of fog advances across country in the forward position of the mild air. If the wind is more than moderate, turbulence lifts the condensation level above the surface and low cloud is formed rather than fog, although if drizzle is present it may keep the surface visibility below 1000 metres, the defining limit of fog. Even if there should be no fog at low levels, the moist air very readily becomes saturated by orographic lifting forming 'fog' over quite low hills, but this has been discussed already as a form of low cloud; the frequent winter fogs over southern and south-west England with mild south-westerly winds are a combination of both effects, except when caused by fog drifting from the sea.

Over the sea, advection fog may form in three different ways. In the first place, the transport of relatively warm air from the land to a colder sea produces fog mostly in spring and summer, since the temperature contrast is then most favourable. Even if the air from the land is not particularly moist, evaporation will accompany the surface cooling and sooner or later the dew-point will be reached. Secondly air may flow from a warmer part of the sea over a much cooler current, a well-known example being the fog of the Newfoundland Banks caused by the cold waters of the Labrador current.

Thirdly, advection fog may develop in an extensive mass of air as it becomes cooled on moving into higher latitudes. For example, if air from the subtropical oceans moves polewards it easily becomes saturated on arrival in temperate regions over the sea, for the cooling may amount to some 10 or 15 degC. Sea fog, in contrast with fog over the land, can form and persist with moderate or even strong wind; often the drizzle stage is reached, reducing visibility below the fog limit even if the fog itself should be lifted off the surface. Fog formed in this way is often widespread and occurs in any season but is much more frequent in summer than in winter. Coastal regions with an on-shore wind may be similarly affected by sea fog at any time of the year, but on a warm day the fog dissipates on passing inland. An example of advection fog drifting over the coast is the 'haar' of eastern Scotland.

There is little or no diurnal variation in advection fog over the sea. Clearance occurs with a change of air mass, or less commonly with an increase of surface wind.

Advection fog is not confined to any particular pressure distribution, since the only general requirement is that the air should move towards a cooler surface and this is as likely to happen in a cyclonic as in an anticyclonic circulation. Some further remarks about advection fog will be found in Chapters 15 and 16.

58. STEAMING FOG

This is named by analogy with the condensed vapour or steam appearing above water which is heated; invisible vapour is given off from the water but is almost immediately recondensed on coming into contact with the colder air. The process requires that the air shall be much cooler than the water; in these circumstances convection currents rapidly develop and fog cannot form unless certain other conditions are also present. These are as follows:

(i) A marked surface inversion in the air before it moves over the sea or inland water, otherwise the lapse rate would quickly be rendered unstable through a deep layer.

(ii) A low air temperature usually about 0°C or below, so that a comparatively small amount of moisture will produce supersaturation. If this condition is not satisfied, the heating process will outweigh the tendency towards saturation.

Fogs of this type are in consequence confined to water surfaces near a source of cold air, such as frozen land, or ice-floes in polar regions; the steaming of roads in sunshine after rain is another example of the same process. In Icelandic and Norwegian fjords and similar regions elsewhere, the fog may attain a depth of 500 feet or more and drift inland so that nearby airfields may be rendered unserviceable. Other names for this type of fog are 'sea smoke', 'frost smoke' and 'Arctic smoke'.

59. FRONTAL FOG

Both radiation and advection fog may be described as 'air-mass' fogs since they depend on cooling taking place within an extensive and more or less uniform mass of air (see Chapter 15). In contrast, 'frontal' fog is associated with the interaction between adjacent air masses. It occurs in one of two ways, either as cloud coming down to the surface with the passage of the front, a type which is more common over hills than over low ground; or it may develop because of saturation occurring in continuous rain preceding the warm front or warm occlusion of a depression, when it forms a belt some hundreds of miles in length and perhaps 200 miles in breadth which advances with the front (see Chapter 16).

60. VISIBILITY IN CLOUD AND PRECIPITATION

Cloud

Horizontal visibility varies from less than 10 metres to over 1000 metres according to the cloud type. In cirrus visibility usually exceeds 1000 metres, in medium cloud it ranges according to circumstances from about 1000 metres down to 20 metres, while in all types of low cloud the range is generally below 30 metres and may fall below 10 metres in cumulus and cumulonimbus.

Precipitation

The visibility in rain depends both on the size of the drops and on their number in a given volume. Light rain has little effect, moderate rain usually gives a visibility of 3–10 kilometres, while heavy rain (as the term is used in temperate latitudes) reduces visibility below 1000 metres. The heavy rain of tropical regions is associated with a visibility of 50–500 metres.

In drizzle the visibility ranges according to intensity from about 3 kilometres down to 500 metres but on some occasions the simultaneous presence of fog droplets reduces the visibility to less than 500 metres.

Snow has more effect than rain, the visibility commonly falling below 1000 metres even in moderate snow, while in heavy snow it varies from about 200 metres to less than 50 metres. Visibility can also be reduced, at times seriously, by drifting snow, that is by snow blown off the ground by the wind. This is most likely to occur when the snow is dry and powdery, i.e. when the temperatures are low and in high latitudes.

Ice crystals

These are sometimes found in otherwise clear air in polar regions. The crystals are small enough to remain in a state of suspension and may produce 'ice-crystal haze'. On one occasion in north-east Greenland these conditions extended from the surface to 8000 feet, reducing visibility to 'very low limits' and almost obscuring the sun at times.

61. WIND-BLOWN SPRAY

When the surface wind speed over the sea reaches force 5, many 'white horses' are formed and spray begins to form from the breaking waves. With stronger winds, foam is blown in streaks along the wind but it is not until the strength reaches force 9 – a strong gale – that spray begins to reduce surface visibility. The effect intensifies with further increase of wind and in a hurricane (force 12) the visibility is said to be 'very seriously affected'. The deterioration is confined to levels within the friction layer. Coasts with on-shore winds are similarly affected by spray.

62. REDUCTION OF VISIBILITY BY SMOKE

In quiet weather the air is often polluted with a think haze caused by the smoke from industrial and domestic fires. The larger particles settle easily under gravity and do not drift far from the populated areas, but much of the pollution is in the form of finely divided particles, comparable in size with the water droplets in a cloud, which may remain suspended in the air until eventually removed by rain or snow.

The obscurity produced by smoke depends on:

(i) the rate at which smoke is being produced,
(ii) the rate at which it is dispersed by wind and turbulence, both horizontally and vertically, and
(iii) the distance from the source of smoke.

In light winds and calms, the smoke remains near the source and local visibility is seriously reduced. The smoke may be dissipated upwards by convection during the day but at night, and by day also if there is an inversion of temperature, it is generally confined near the surface and horizontal visibility remains low. The stronger the wind, the more rapidly is the smoke carried away and dissipated through the atmosphere. Not only does the plume of smoke broaden as distance from the source increases, but it also tends to spread vertically unless confined by an inversion. In conditions of moderate wind and small turbulence, the smoke haze from London and other large towns may reduce surface visibility to below 2 kilometres as far as 50 miles from the suburbs and occasionally as far as 70 miles. In the air, the smoke plume may extend to even greater distances, that from London for example, with suitable winds, being traceable to the south coast and across the English Channel. Smoke from the Black Country and from the industrial areas of northern England and southern Scotland is similarly effective in reducing visibility far from the source. Glasgow smoke provides a good example of the effect of topography on the concentration. In drifting north-eastwards through the Clyde-Forth Valley, the lateral spread is restricted and the concentration of smoke on arrival at the east coast is greater than it might have been had the flow been across level country. A mountain mass may protect its lee slopes from pollution produced on its windward side. It may be observed that the clear atmosphere of many coastal regions is due not so much to the absence of a source of pollution as to the predominance of sea breezes, notably in the summer season. Even a large residential resort such as Blackpool, with an enormous industrial area as its hinterland, has a valuable proportion of brilliantly clear weather; owing to the prevailing wind direction the western coasts generally have an advantage in this respect. However, much depends on the actual winds at any one time; irregular distributions of smoke, often found 50 miles or more from the source, provide many sudden deteriorations in visibility.

When smoke accumulates beneath an inversion layer, visibility may be reduced well below 1000 metres; if, however, a water-drop fog is already present, then the obscurity is substantially intensified by the addition of smoke and the atmosphere takes on an unpleasantness which amply justifies the designation 'pea-soup' fog. Moreover, since combustion of fuels produces hygroscopic nuclei (Section 31) the presence of smoke often causes fog to form before the relative humidity has reached 100 per cent. Because of the serious effects of smoke pollution, the foggiest parts of the British Isles are the industrial areas; there the fog is not only more frequent, but thicker and more persistent than it is elsewhere. Country fogs are generally white and clean and the sun can usually break through during the day and cause at least a partial clearance even in winter; town fogs may almost completely shut off the light of the sun and clearance takes place only slowly, if at all. Frequently visibility deteriorates during the day after domestic fires have been lit.

The names 'high fog' and 'anticylonic gloom' have been given to conditions which sometimes arise when an inversion of temperature at some little height causes smoke to collect in a layer above the surface while vertical currents prevent undue accumulation nearer the ground. The smoke combines with any cloud layer (stratus or stratocumulus) below the inversion and produces a thick black pall which more or less completely darkens the sky and necessitates artificial lighting, even at midday.

63. REDUCTION OF VISIBILITY BY DUST AND SAND

One source of atmospheric dust is the solid matter thrown out from active volcanoes; it is carried as haze to great distances and is indeed regarded as an important climatic factor since it reduces the intensity of the solar radiation reaching the ground; as an element in aviation meteorology its significance is, however, negligible. The only sources of dust which need concern us are the sandy deserts of the world and the semi-desert regions in the dry season.

Dust or sand is raised from the ground by the wind and carried upwards to a height depending on the meteorological conditions and the size of the particles. Whenever the deterioration in visibility is serious the phenomenon is generally known as a duststorm or sandstorm. The terminology is not however everywhere the same and the conditions associated with a 'duststorm' vary widely from place to place. We are concerned here only with general principles, but the pilot flying in regions subject to such storms should make himself familiar with the local types and the special names applied to them.

Wind-blown sand or dust

Whenever the wind exceeds a moderate value, 15 or 20 knots, dust is picked up and whirled along with the wind and carried upwards by turbulence. Large sand particles are too heavy to be lifted far and some sandstorms consist merely of this surface sand, no more than three or four feet deep. To an aircraft coming in to land, the surface of the ground may be blanketed but hangars and buildings stand out. Small dust particles, however, are carried up through the whole turbulent layer to some 3000 feet and remain in suspension as long as the wind blows with sufficient strength. This type of storm can be prolonged and widespread, or it may occur in patches in association with minor squalls of wind. The diurnal variation of wind speed is usually marked in the regions affected and such storms rarely occur at night. In any case the existence of duststorms and sandstorms requires the presence of fine sand or dust in or upwind of the area affected.

Haboobs

Among the more severe types of duststorm are the haboobs of the Sudan. This term has been loosely used in the past in connection with any duststorm but among meteorologists it is now restricted to summer duststorms originating with cumulonimbus clouds. Because of the gusty wind and the strong convection in and below the clouds, the dust is carried high into the air, sometimes even to above 10000 feet, especially when the cloud base is this high. The storms can be seen approaching at distances of 50 miles or more as an apparently solid wall of dust and their arrival is characterized by a sudden increase of wind and a rapid deterioration of visibility. It may not always be practicable to fly over them, but in most cases they can be avoided. In local flying a careful watch should be kept; if a landing is contemplated it should be made in good time and the aircraft placed if possible under cover as a protection from the penetrating dust and the possibly violent squalls. Although the storms form initially during the day-time, the time of occurrence at any one place depends on the locality, as they frequently form in association with a line-squall and travel along with it.

Other types of duststorm

A more widespread type of duststorm occurs with strong winds of long fetch over desert. Both the strength of the wind and instability of the lower layers contribute to the dust-laden atmosphere. Such storms may continue for many hours and can cover a considerable area, visibility being reduced to a few hundred metres or even to tens of metres, while they may extend upwards a few thousand feet. They may continue well into the night, although they are then generally less severe. They are not only extremely unpleasant on the ground but constitute a serious hazard to flying on account of their vertical and horizontal extent and their duration.

Dust-devils or sand-pillars

During the heat of the day over any hot dusty region small whirlwinds develop by convection; these are known as dust-devils or sand-pillars. Their diameter is quite small, a matter of yards, but the wind may easily exceed gale force in the whirl and surface dust is picked up and carried in a dense column to 1000 or 2000 feet. Minor structural damage is not uncommon but as a factor in visibility they are of no particular consequence owing to the small area covered. They should of course be avoided in flight, particularly when taking off or landing.

Dust haze

The fine dust carried upwards in duststorms of various sorts is gradually dissipated through the atmosphere and gives a general haziness to all air of desert origin. Polluted air may remain hazy for days or even indefinitely because the particles are too small to fall at any measurable speed by gravity. Haze commonly extends to 10000 feet; it seldom reduces visibility to less than 1000 metres but the air-to-ground visibility is invariably less than that at the ground. The final clearance may not occur until the particles are washed out by rain or snow, perhaps thousands of miles from the place of origin. This process gives rise to 'coloured rain', sometimes seen in the British Isles, which has a tint characteristic of the dust being washed out.

Haze layers

Sharply defined layers of haze may occur at any level in the troposphere and give a haze horizon when viewed from above. The most common occurrence is when the surface air, polluted with smoke or dust, is carried upwards by turbulence and convection until it meets a reduced lapse rate or inversion which the upcurrents cannot penetrate. Largely on account of the diffuse reflection of light from the upper surface, the layer may completely obscure the ground in any direction far from the vertical; on some occasions the ground may be entirely invisible although the aircraft may be seen clearly by an observer on the ground. Occasionally a haze layer is found in a position which would be occupied by cloud if the air were sufficiently humid, or it indicates a layer previously occupied by cloud which has been evaporated by descent and adiabatic warming. Anticyclonic regions, in which subsidence is characteristic, have usually one or more haze layers in the higher levels. Dust would not be expected to penetrate to any observable degree into the stratosphere, but the tropopause when free from cloud may be revealed as a well-marked haze layer to an observer above.

64. ARTIFICIAL DISPERSAL OF FOG

The problem of the artificial dispersal of fog on an airfield is one of some importance in aviation. It became a practical proposition during the Second World War 1939–45 but since then its use has been severely limited because of the expense.

The method used was to produce heat by petrol burners laid in lines forming a rectangle enclosing the runway but at a distance of at least 50 yards from the boundaries. When air containing fog droplets is heated, some of the heat is absorbed in evaporating the drops and the rest in raising the temperature of the air. For radiation fogs it was found that the amount of heat required to disperse a dense fog was equal to that required to produce a rise of 4 degC in clear conditions; smaller amounts of heat were found sufficient to disperse thick and moderate fogs. Experiments also showed that the larger fog droplets take an appreciable time to evaporate, even though ample heat is available. The amount of heat required depends among other factors on the speed of the wind and on its direction relative to the runway. The artificial dispersal of advection fog is more difficult because of the greater wind speeds which are usually present.

Fogs in sub-zero temperatures have been dispersed in the U.S.A. and U.S.S.R. by seeding with dry ice (solid carbon dioxide). As the dry ice falls it produces many tiny ice crystals which grow into snow at the expense of the supercooled drops. The visibility usually improves within about 30 minutes and the improvement is maintained for about an hour.

CHAPTER 10

SOME CHARACTERISTICS OF HIGH-ALTITUDE FLIGHT

65. INTRODUCTION

With the earliest aircraft, flight was necessarily restricted to the lowest levels but with later developments flight at increasingly high levels has become not only possible but desirable on account of certain advantages it brings with it. First among these is the fact that fast aircraft are more efficient, and so are operated more economically, in high-altitude flight than at lower levels, a tendency which has been increased by the development of jet aircraft. Also, flight at the highest practicable level is required for military aircraft if only because it may help to reduce the chance of detection and interference by enemy forces. A further advantage of high-altitude flight is that it often takes place 'above the weather'. On the other hand there are certain disadvantages which are not encountered, or are much less severe, at lower altitudes; these include the possibility of extremely strong winds and of severe turbulence, even in clear air.

There is no sharp line of demarcation between high-altitude flight and flight at lower levels, but perhaps a limit in the neighbourhood of 25000 feet would be fairly representative on most occasions. Thus the cruising stage of a high-altitude flight may take place either in the upper troposphere or in the stratosphere or possibly partly in one and partly in the other, as for example would be the case with an aircraft flying from middle to low latitudes at say 40000 feet. The nature of the changes in the meteorological characteristics between troposphere and stratosphere has been explained in earlier chapters; they are such that the height of the tropopause is a matter of primary importance for high-altitude flight. For information on the variation in the height of the tropopause with latitude, reference should be made to Chapter 3 and Fig. 13, while further details will be found in Chapter 22, Section 157. Details of the large changes of temperature and wind which occur in the stratosphere both in middle and high latitudes in the winter months are given for the northern hemisphere. Similar changes are expected to occur over the southern hemisphere but the scarcity of observations makes the drawing of charts less reliable (Section 72).

66. AIR DENSITY AND AIRCRAFT PERFORMANCE

Both the lift and drag of an aircraft are directly proportional to the air density; other things being equal, an aircraft must fly faster to maintain height when the density is reduced. Since density decreases rapidly with increase of height in the atmosphere, flight at high levels necessitates an increase in either wing area or airspeed; the latter is made possible with modern engines and the greater airspeed is moreover facilitated by the reduced drag at these levels. However, at about two-thirds the speed of sound (574 knots in the stratosphere of the International Standard Atmosphere) a marked increase in drag begins to develop, connected with an increase of density caused by compression as the air impinges on the aircraft. The same effect operates on the tips of propeller blades even at much lower aircraft speeds; as the thrust of propellers also falls off as density decreases or as height increases, it will be appreciated that this type of propulsion becomes less and less efficient at speeds over about 400 knots and so is more suited to low-level flight. On the other hand, the jet-driven aircraft approaches its maximum efficiency at speeds of this order, but it should be mentioned that with these aircraft the performance does not depend on density alone, since temperature itself has a direct effect (Section 18).

With the internal combustion engine, the amount of oxygen drawn into the cylinders at each piston-stroke is determined by the air density; with increase of height the oxygen intake sooner or later becomes inadequate to maintain the required power and supercharging is then necessary; similarly with jet engines the air intake must ultimately decrease with height, in spite of increased airspeed. For a similar reason, additional oxygen must be provided for aircrew and passengers; on some aircraft this is arranged by means of cabin pressurization. In many ways therefore aircraft need to be specially designed for high-altitude flight; moreover high-speed aircraft can be operated efficiently only at high altitudes.

In Section 22 it was seen that the value of density becomes practically uniform over the earth at heights of about 26000 and 80000 feet. Between these two levels density in general increases from the poles to the equator and, in other words, the true height corresponding with a given density is greater in low latitudes than in high latitudes.

67. JET STREAMS

These are exceptionally fast streams of air confined to relatively narrow bands of the atmosphere. The velocity at the centre of the stream is commonly about 100 knots and may reach 200 knots at times over Europe and the North Atlantic, while speeds approaching 300 knots have been recorded over south-east Asia. These high speeds extend for many hundreds of miles along the track of the wind, but to either side and above and below the axis they fall off sharply. The strong-wind core is usually about 200 miles wide but only about 2 miles deep. The wind velocity in jet streams mainly results from large temperature gradients which develop in horizontal bands through a great depth of the atmosphere.

Two major areas of jet-stream activity in the upper troposphere are the subtropical jet streams near latitudes 30° and the jet streams associated from time to time with the polar front in middle latitudes in both hemispheres. These jet streams have a generally westerly flow, but marked deviation from westerly can occur on any particular occasion. There is also evidence of a jet stream with a predominantly easterly flow in equatorial latitudes, mainly in longitudes where the upper easterlies are displaced some 10° or so from the equator. All these jet streams usually reach their maximum speeds near the level of the tropopause, i.e. the equatorial jet at about 100 millibars, the subtropical jet at about 200 millibars and the polar-front jet at about 300 to 250 millibars.

There is a belt of very strong winds in the stratosphere of the winter pole with the jet core well above the 50-millibar level (see Section 72).

Jet streams of mid latitudes

The mid-latitude jet streams, like many other features of the atmosphere, show great variation between one jet and another, and any one jet stream displays variations from day to day and from place to place.

The polar-front jet stream of the northern hemisphere is so variable that it does not show clearly on the chart of mean wind speeds (Section 160). The subtropical jet stream, being more constant in time and space, is clearly marked on the mean charts of both hemispheres. The maximum values of mean winds over Asia are probably a result of the combination of effects of adjacent polar-front and sub-tropical jet streams.

Though jet streams result invariably from the existence of powerful thermal gradients, the corresponding surface fronts are not always easily identified, and the location of the jet axis with respect to fronts varies from case to case. Usually the jet lies more or less parallel with the surface fronts and over the surface cold air. Fig. 60 shows the jet-stream position with respect to surface fronts on a number of occasions over the British Isles.

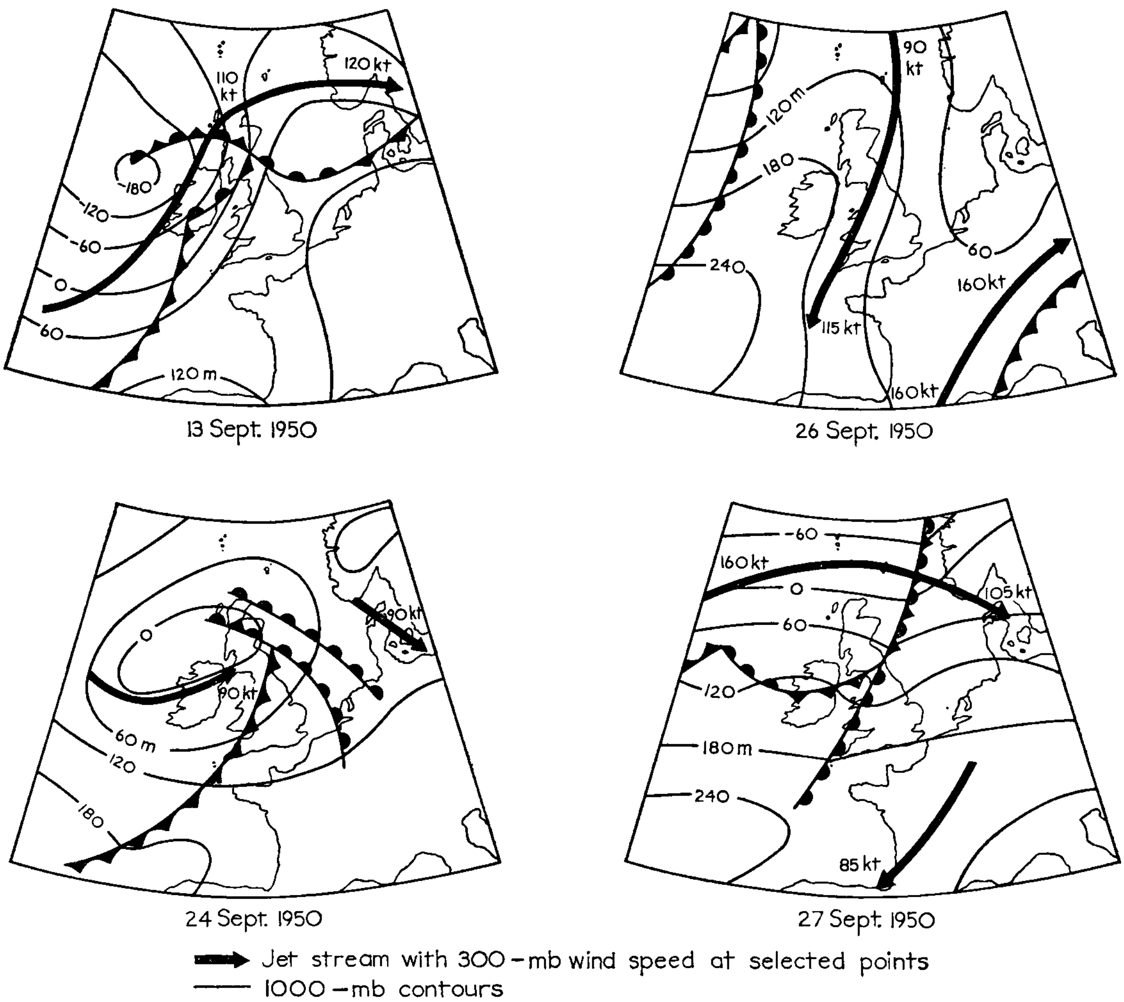


FIG. 60. *Examples of association between jet streams and surface fronts*

A cross-section of a polar-front jet stream is shown in Fig. 61 (the vertical scale is greatly exaggerated in relation to the horizontal scale). The horizontal temperature gradient throughout the troposphere can be clearly seen. The tropopause is higher above the warm air than above the cold air and the core of strong winds lies under the warm tropopause. Above the jet core the horizontal gradient of temperature, and consequently the thermal wind, is reversed and the wind speeds diminish with increasing height. It will be noted that the decrease of speed towards the colder tropospheric side of the jet is very well marked.

Large variations of wind speed occur along the jet axis. A wavy jet as a whole moves generally with the wave patterns, but at the same time the individual wind maxima tend to move downstream, i.e. with a component along the jet axis.

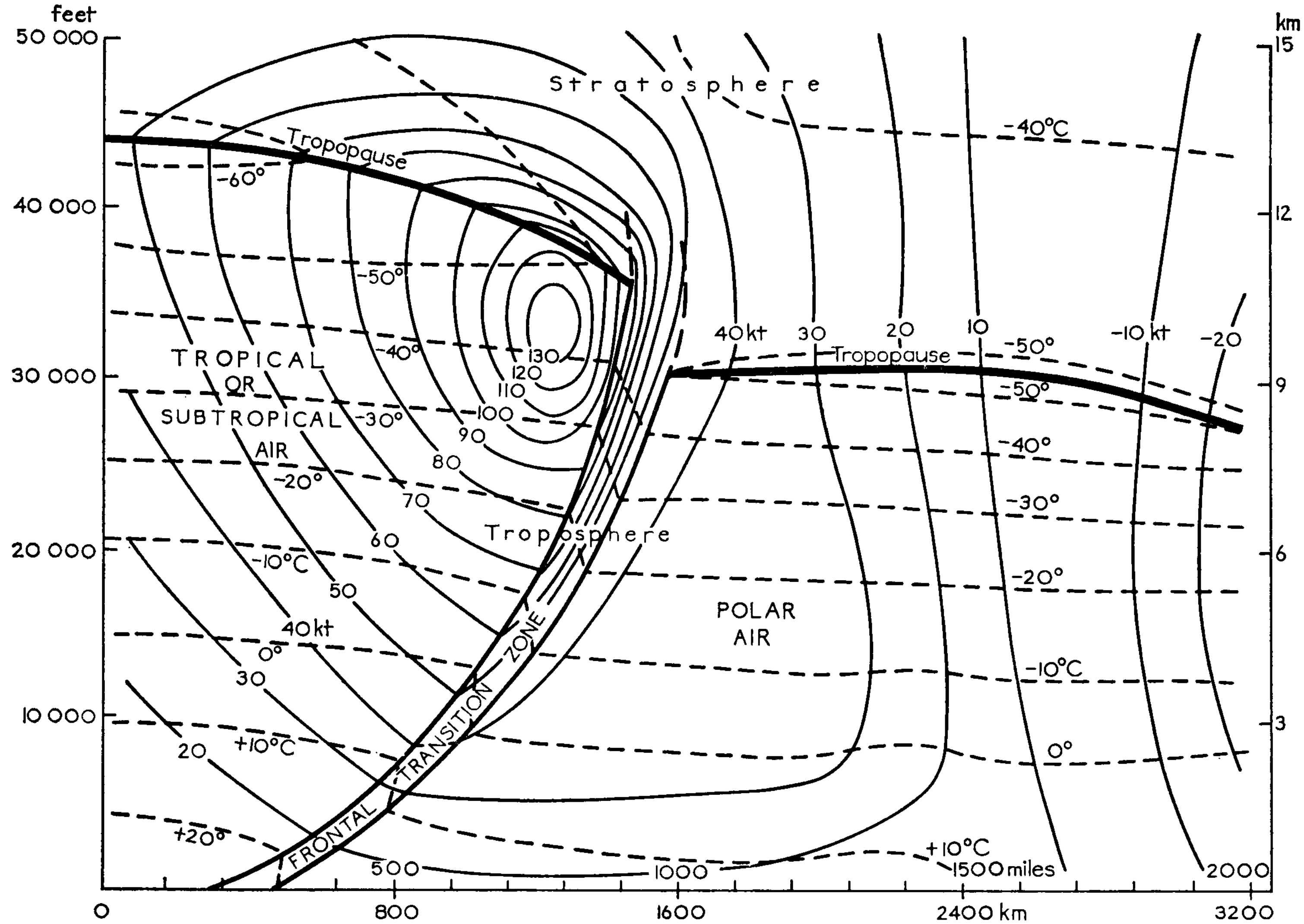


FIG. 61. *Vertical cross-section across a jet stream in summer*

——— Isotachs - - - Streamlines

Positive wind speeds have a component out of the plane of the page, negative speeds indicate a component into the page.

Further examples of cross-sections showing the jet core of a polar front will be found in Sections 106, 107 and 143.

Subtropical jet stream

This is located entirely within the tropical air mass and is not associated with any surface fronts. The core lies near the 200-millibar level below the tropical tropopause and often in the vicinity of the break between this and the tropopause of middle latitudes.

The location of the subtropical jet stream is less variable than that of the polar-front jet, and the strong westerly flow shows on the mean charts (Section 160) from 500 to 100 millibars over Eurasia in January, with the strongest mean wind over Japan on the 200-millibar chart.

In July a westerly jet stream meanders across the southern hemisphere with a maximum speed on the mean charts of 90 knots over central Australia at 200 millibars.

68. EFFECT OF WIND CHANGES ON SEXTANT OBSERVATIONS

When a bubble sextant is subjected to an acceleration such as that produced by the turning of an aircraft or by variation of airspeed, the bubble suffers a displacement. The adjustment of the setting required to re-establish coincidence introduces an error α into the indicated altitude, of which the magnitude is given by

$$\tan \alpha = \frac{f}{g},$$

where f is the component of the acceleration directed horizontally in the plane of the vertical circle through the observed body and g is the acceleration due to gravity. For this reason among others, observations with a sextant should preferably be made only when the aircraft is in steady flight on a fixed course. Now when the wind is varying along the track, the aircraft is subjected to an acceleration on this account. The rate of change of wind at the position of the aircraft depends not only on the change of wind with distance but also on the ground speed of the aircraft. With fast aircraft a vector change of wind amounting to 100 knots per hour would be not uncommon at high levels; this is equal to 1·4 centimetres per second per second and by the above formula the corresponding maximum sextant error would be 5 minutes of arc. This is the same as the maximum error that would be produced by Coriolis acceleration (the product of $2\Omega \sin \phi$ and the ground speed) in latitude 40° with a ground speed of 300 knots. The direction and magnitude of the accelerations and the consequent sextant errors could be estimated if adequate forecast wind information were available, but without wind data the errors would remain unknown except for the order of magnitude which in most cases is likely to be comparable with that of the errors produced by Coriolis acceleration. Much greater acceleration than that just mentioned may be experienced when flying in the vicinity of jet streams, while a fast aircraft would also be subjected to rapid changes of wind when flying through frontal zones (see Chapter 16). With the aid of forecast data supplemented by observations of the state of sky it should in general be possible to avoid using the sextant in such conditions.

69. CLEAR-AIR TURBULENCE

The occurrence of turbulence within the first few thousand feet of the earth's surface has already been seen to have a dual origin, frictional and thermal (Section 25). The former type depends on friction with the ground and is confined to the friction layer. The thermal type on the other hand, while often originating in convection currents at the surface, can develop upwards to a height depending on the lapse rate of temperature and, when extending to a high level, is always associated with convective clouds. Turbulence of a convective nature may also originate occasionally from thermal instability in the upper air, but again its presence is revealed by cloud. Nevertheless flying conditions in the upper air may be bumpy even in the absence of clouds. This type of turbulence appears to arise from internal friction of the atmosphere which leads to a breakdown of smooth flow either when the rate of change of wind with height surpasses a certain limit depending on the lapse rate of temperature, or when the rate of change of wind with horizontal distance becomes sufficiently great.

Attention was first drawn to the occurrence of this clear-air turbulence at great heights, though it seems clear that it can occur at any height, merging towards the surface layers into the more familiar types. It is however most important at high altitudes because there, in clear air, it is relatively unexpected, being encountered without any visual warning such as would be provided by the presence of cloud. Turbulence (of whatever origin) is important to aviation for several reasons. It may cause fracture of some part of the aircraft, due either to particularly severe gusts or to more moderate gustiness continued over a long period; also at high altitudes a severe gust may be the cause of a stall, and the resulting loss of control can be dangerous, especially for a large aircraft; finally, air travel is uncomfortable in turbulent conditions.

Observations of clear-air turbulence are made either qualitatively by an observer in an aircraft or by use of an accelerometer. The severity of the bumps affecting an aircraft is proportional to the airspeed but depends also to some extent on other characteristics of the aircraft as well as on the nature of the turbulence itself. The accelerometer records only the vertical acceleration of the aircraft; conversely the record is interpreted in terms of a vertical gust. More precisely it is expressed in terms of the equivalent velocity of a sharp-edged gust; this is defined as that instantaneous vertical velocity which superimposed on a steady horizontal wind would produce the measured acceleration of the aircraft. The acceleration is then directly proportional to the equivalent gust velocity and to the airspeed.

Much more needs to be known about the cause and incidence of clear-air turbulence. Although it decreases with height above the tropopause, turbulence does not become negligible in the stratosphere.

Turbulence is associated with strong vertical and horizontal wind shear. High-level turbulence appears to occur in patches with horizontal dimensions of about 50 kilometres or more and an average vertical extent around 600 metres, although it is sometimes as little as 20 to 30 metres.

Clear-air turbulence is frequently associated with a jet stream; a preferred region appears to be near or below the jet axis on the low-pressure side with a secondary maximum above the axis on the anticyclonic side. Not all jet streams are turbulent. Clear-air turbulence may also occur at high levels in standing waves over mountains and hills. Jet streams passing over mountains usually produce waves of considerable amplitude and this increases the risk of turbulence. Turbulence is also associated with sharp upper troughs and occasionally with upper ridges. It is associated with cumulonimbus clouds and is to be expected above cumulonimbus tops.

Most of the reports of clear-air turbulence are of light or moderate character, but 5 to 10 per cent are of heavy turbulence and 1 to 3 per cent of cases are reported as violent.

70. CONDENSATION TRAILS

Condensation trails – contrails, for short – are elongated streaks of cloud formed by the passage of an aircraft (see Plate XXVIII). Although commonly associated with the upper troposphere, they can form at any level, even at the ground, according to the temperature, humidity and type of aircraft. They are formed in various ways, the more important of which are described below.

Exhaust trails

One of the combustion products of petrol and other aircraft fuels is water; this is ejected through the exhaust and tends to raise the relative humidity of the air in the wake of the engines. On the other hand, the heat generated by the engines tends to lower the relative humidity by raising the temperature in the wake. In certain conditions the net result is to increase the humidity to the saturation point so that a cloud is formed trailing behind the aircraft. This type of trail can ordinarily occur only if the air temperature is below a critical value which varies almost linearly from about —24°C at sea level to about —45°C at 50 000 feet. The critical temperatures, which are only slightly affected by the type of aircraft, are indicated on the tephigram (Metform 2810) by a line marked MINTRA and apply to an aircraft flying at cruising speed in an atmosphere just saturated with respect to ice; the corresponding temperatures for saturation with respect to water are lower by about two or three degrees Celsius. It should be mentioned that trails can occur exceptionally at temperatures above the MINTRA value, for example when the free air is supersaturated with respect to ice, or when fuel consumption per unit air distance is greater than it is under normal cruising conditions, as when operating with throttle fully open. Conversely, trail formation may at times be lessened or even avoided by reducing the throttle setting.

Once a trail is formed, it broadens by diffusion. If the surrounding air is at or near saturation, the trail evaporates only slowly or not at all and is then long and persistent; if the relative humidity is low, the trail appears only as a short plume behind the aircraft. It was remarked earlier that supersaturation with respect to ice is not uncommon. Since the exhaust gases contain sublimation nuclei, any trail formed in these conditions is persistent and may thicken until the ice particles fall out as snow. In the stratosphere, because of the extremely low relative humidity normally found in this region, exhaust trails are usually evanescent. If however the temperature is very low, then only a small increase in vapour content is required for saturation so that persistent trails may be formed; these conditions are most likely to occur in the tropical stratosphere where the lowest temperatures are to be found.

The MINTRA temperature at any level is such that condensation trails cannot ordinarily form at a higher ambient temperature. The converse that trails should form when the air temperature is less than MINTRA is not true. The attainment of saturation is not sufficient for the trail to become visible; condensed water or ice particles must be present in a sufficient concentration to be seen, and this depends upon illumination, background contrast, distance, and other conditions of viewing.

MINTRA is that temperature above which a trail will not form even if the ambient air is already saturated. A DRYTRA would be a line joining temperatures at which trails must occur even if the ambient air is completely dry. Between these two limits, corresponding with saturation and complete dryness, trail formation may be expected to depend on the relative humidity of the environment.

It is generally accepted that at temperatures less than 11 degC below MINTRA, trails are not expected; at temperatures 11 to 14 degC below MINTRA, non-persistent short trails are probable; at temperatures more than 14 degC below MINTRA, persistent long trails are expected.

Trails may form at any height attained by an aircraft provided that the temperature is suitable. At 1000 millibars the critical temperature is about —30°C, so trails can occur at ground level in high latitudes. Over southern England the average height at which trails would be expected to form in the winter ranges from about 26 000 to 70 000 feet; in summer, when the stratosphere is warmer, the range is from about 30 000 to 45 000 feet.

Wing-tip trails

These very thin transient trails are formed aerodynamically by the reduction of pressure at the extremities of the wings of an aircraft, the expansion causing a reduction of temperature to below the dew-point. If the air temperature is already low, insufficient water may be condensed to produce a visible trail; also if the air is very dry the condensation point may not be reached. Accordingly these trails are usually seen in mild damp weather at low altitudes. Similar effects may occur at the tips of the propeller blades and over the upper surface of the wings.

Avoidance of condensation trails

In clear weather, a condensation trail reveals both the position and heading of an aircraft, so that for military operations it may be important to avoid the formation of trails. Some suggestions for evasive action are given below.

(i) Formation of trails can usually be avoided by flying at a height where the air temperature is above the MINTRA value.

(ii) If possible, fly within the stratosphere. If the trails do not cease immediately, increase height still further. This action may not always be effective when the temperature is particularly low, as in the high and cold stratosphere of the tropics, but then the height at which the air temperature first falls to the MINTRA value may be so great that flight below that level may be operationally practicable.

(iii) Try to find a dry layer at some other level. A likely level would be just above a thin layer of cloud such as altocumulus, especially if the cloudlets are flat-topped, implying the existence of a temperature inversion and probably dry air. (Flight above, or within, a continuous layer of cloud naturally provides concealment from the ground.) Alternatively reduce height to a level which during the climb had been observed as free from trails.

(iv) A layer of air in which cloud is forming, such as wisps of cirrus, is likely to have a high humidity and should be avoided.

(v) Reduce the throttle setting as far as that is practicable.

71. MACH NUMBER

At high airspeeds compressibility of the air as it impinges on the airframe assumes great importance but the significant factor is not the actual speed of the aircraft but the ratio of its speed to the local velocity of sound; this ratio is known as the Mach number. During flight, the disturbance caused by the aircraft sets up pressure waves which spread out in all directions with the speed of sound. At airspeeds much less than the speed of sound, the pressure waves move ahead of the aircraft so that when the aircraft reaches any point the air is already modified and flow over the airframe is reasonably smooth. When on the other hand the airspeed is greater than that of sound, the pressure waves set up by the motion cannot reach the air ahead; the aircraft is then flying into undisturbed air; there is intense compression on the leading edges and the flow pattern is much less smooth than the one which prevails at low speeds. These changes have an adverse effect on lift and drag, on the stability of the aircraft and on its reaction to the controls, and begin to become noticeable at about two-thirds the speed of sound. Knowledge of the Mach number is therefore vital for pilots and is given directly by a meter on the instrument panel.

The speed of sound in dry air is proportional to the square root of the absolute temperature. At sea level (temperature 15°C) in the International Standard Atmosphere the speed of sound is 663 knots; in the stratosphere (—56·5°C) the speed is 574 knots. Thus the machmeter has to be designed to take account of the dependence of the speed of sound on temperature. The meter however makes use only of pressures, as the following argument shows. If p' is the dynamic pressure on a leading edge as measured by a pitot head, and p the static pressure, then $p'-p$ is proportional to ρA^2 where ρ is the air density and A the airspeed. Thus A is proportional to $\sqrt{\{(p'-p)/\rho\}}$ or to $\sqrt{\{(p'-p)T/p\}}$, where T is the temperature in degrees Kelvin. Also the speed of sound a is proportional to $\sqrt{T}$. Therefore the Mach number A/a is proportional to $\sqrt{\{(p'-p)/p\}}$ and so depends only on the dynamic and static pressures.

72. WINDS AND TEMPERATURES IN THE STRATOSPHERE

Data about the earth's higher atmosphere is steadily increasing but knowledge of conditions in the stratosphere is still far from complete. However, it is evident that conditions in the stratosphere change very considerably with the seasons. The linear dimensions of the wind systems are in general greater than those encountered in the troposphere.

In summer the winds and temperatures usually change little with time and distance. In winter, disturbances of the stratosphere can be of such great amplitude that in middle and high latitudes in the layer from 50 millibars to at least 10 millibars the lowest and highest temperatures of the year may both occur. The life cycle of these disturbances usually extends over a few weeks rather than over a few days The incidence of such disturbances is variable from winter to winter although January and February appear more prone to disturbance than other months. The amplitude of these phenomena also varies considerably from winter to winter so that although disturbances occur each winter, usually sometime between December and April, really large-amplitude disturbances occur less than once a winter. These disturbances are centres of low temperature with the lowest value possibly at about 20 millibars; winds of great strength, increasing with altitude, circulate round the low-temperature areas. The centres at 10, 30 and 50 millibars are not always vertically above each other but lie on a slope which changes with time.

Examples of 5-year means of temperatures and of winds at 50 millibars (Figs. 62, 63, 66, 69 and 70) illustrate the great seasonal changes occurring at this level. In general, in summer the actual winds and temperatures in the lower stratosphere are similar to the means; but in winter and spring the warmings which occur are very variable and actual conditions on any particular day are frequently far from the mean.

Temperatures and winds at 50 millibars in January are shown in Figs. 62 and 63. Fig. 62 shows that a cold vortex is situated near the winter pole. In January, on average, the temperatures at this level are near their winter minimum in the northern hemisphere. The main features of the chart are a cold region over north Greenland and the north pole, a warm region over the Aleutians, a warm belt around the hemisphere near latitudes 45° to 50°N, and a cold region over the tropics. As

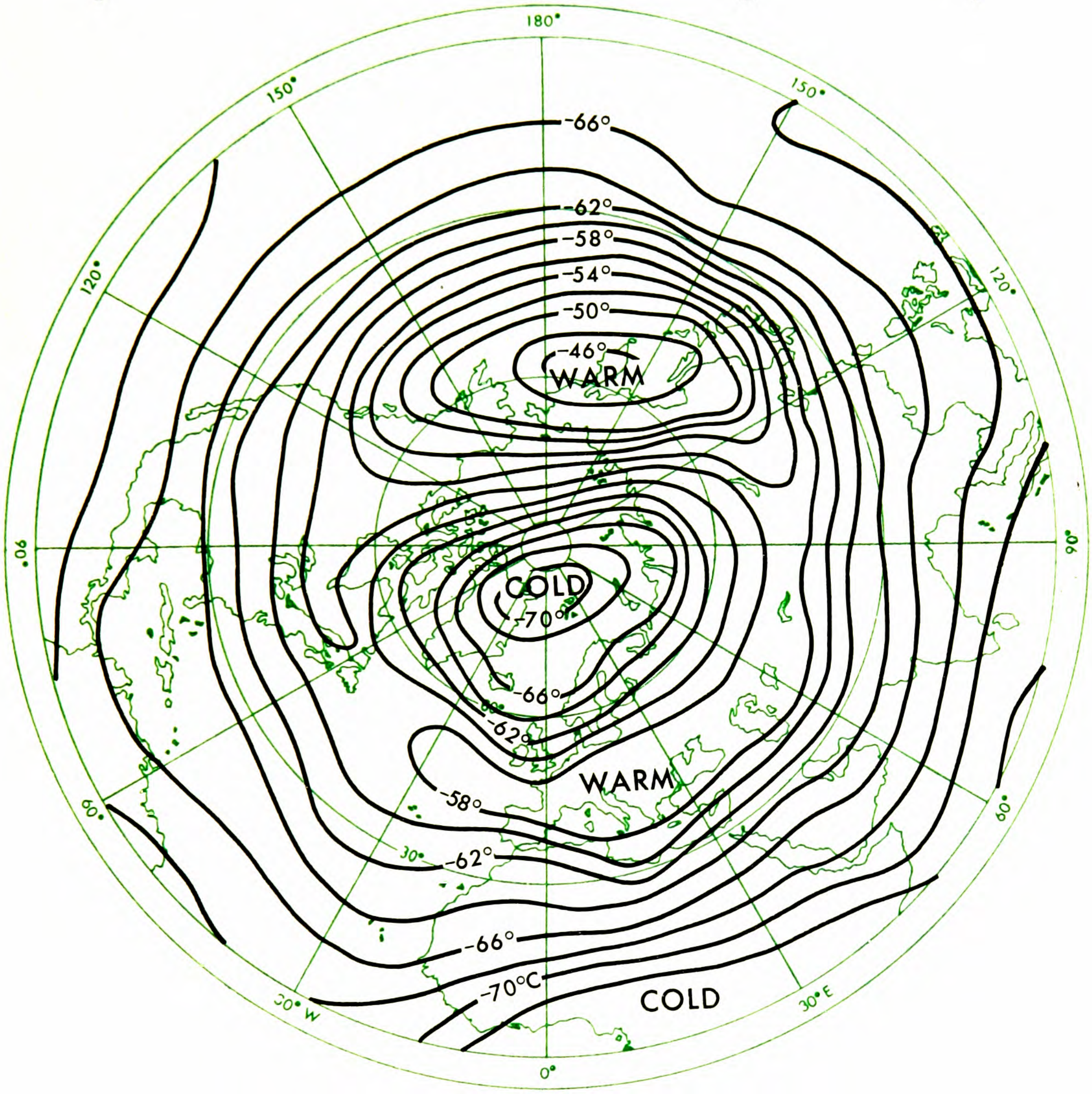


FIG. 62. *Average 50-millibar temperature, January, 1957–61*

would be expected with the thermal distribution, a well-defined belt of westerly winds with average speeds exceeding 50 knots blows around the pole. These westerly winds reach a maximum speed well above the 50-millibar level. Observations of between 100 and 150 knots are not unusual, and occasional values between 150 and 200 knots are reported in soundings which reach the levels of 20 and 10 millibars.

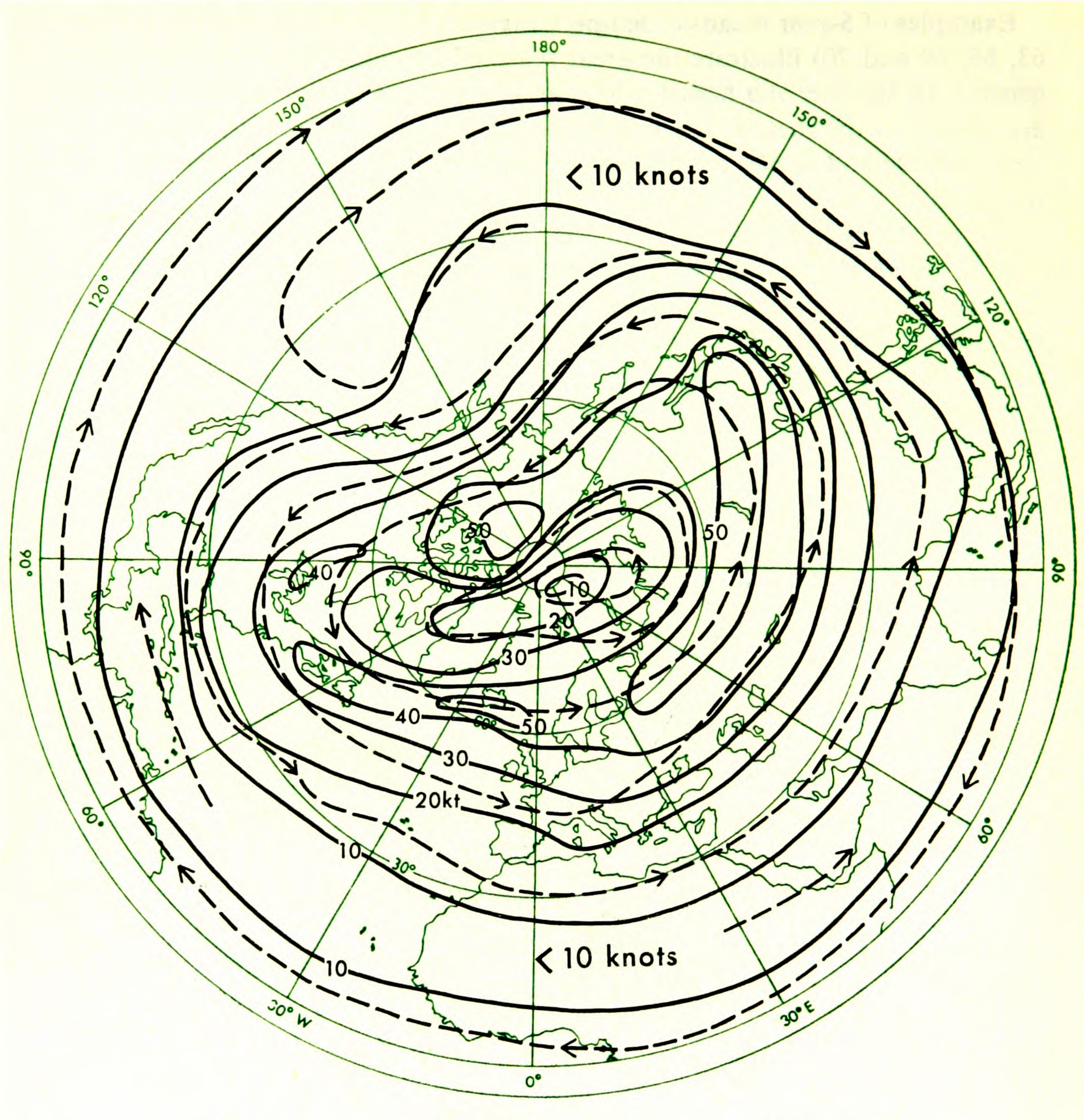


FIG. 63. *Average 50-millibar wind, January, 1957–61*

——— Isotachs — — — Streamlines

In low latitudes, according with the reversed thermal gradient from 45°N to the equator, the average wind at this time of the year at 50 millibars is light easterly.

In the tropics, winds are more complex than the mean picture suggests. In places there is a fluctuation between easterlies and westerlies which has a period of between 22 and 35 months. Fig. 64 shows that once a predominantly easterly régime

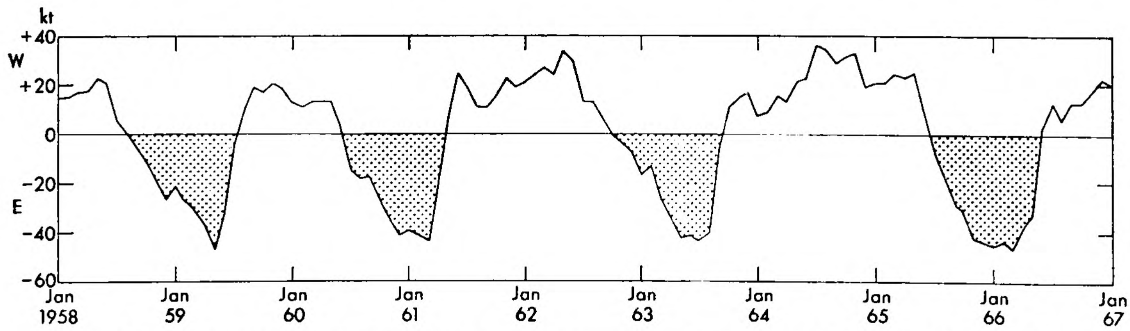


FIG. 64. *Monthly average zonal wind components at 50 millibars, Canton Island*

Westerly components are positive, easterly components are negative.

is established it is likely to persist for about 10 or 13 months whereas the westerly régimes tend to be much more variable in length and may persist for periods ranging from 11 to 24 months. A satisfactory explanation of this near-biennial tendency has not yet been advanced.

In late winter and early spring the thermal gradient towards the pole reverses. There is a considerable difference from year to year in the time and manner of this change. Some areas experience sudden changes of temperature: these occur in some years in January and in other years as late as April. Over a period of days or weeks the temperature may fluctuate between the mean values for winter and summer. The variability of the warming is indicated in Fig. 65 which gives the

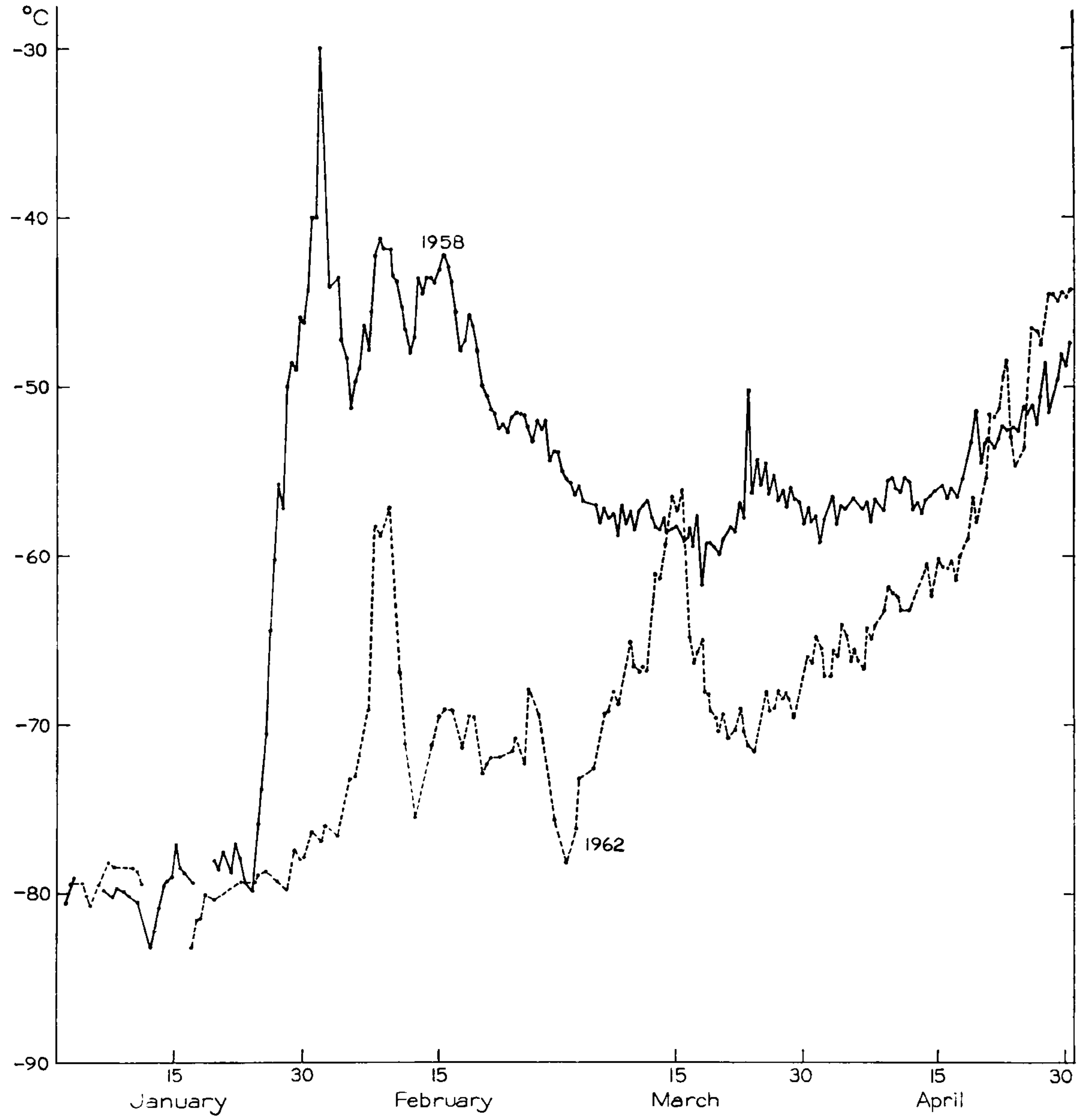


FIG. 65. *Temperature at 50 millibars at Alert, January to April, 1958 and 1962*

50-millibar temperatures at Alert (82°30′N, 62°20′W) for January to April for the years 1958 and 1962. The dramatic nature of the warming that can take place is clearly shown. In 1958 a rise of temperature of 50 degC occurred in a few days in late January and early February. Fig. 66, average temperatures at 50 millibars for April, shows the reversal of temperature gradient completed and a steady gradient established from the warm region over the pole to the cold region over the equator.

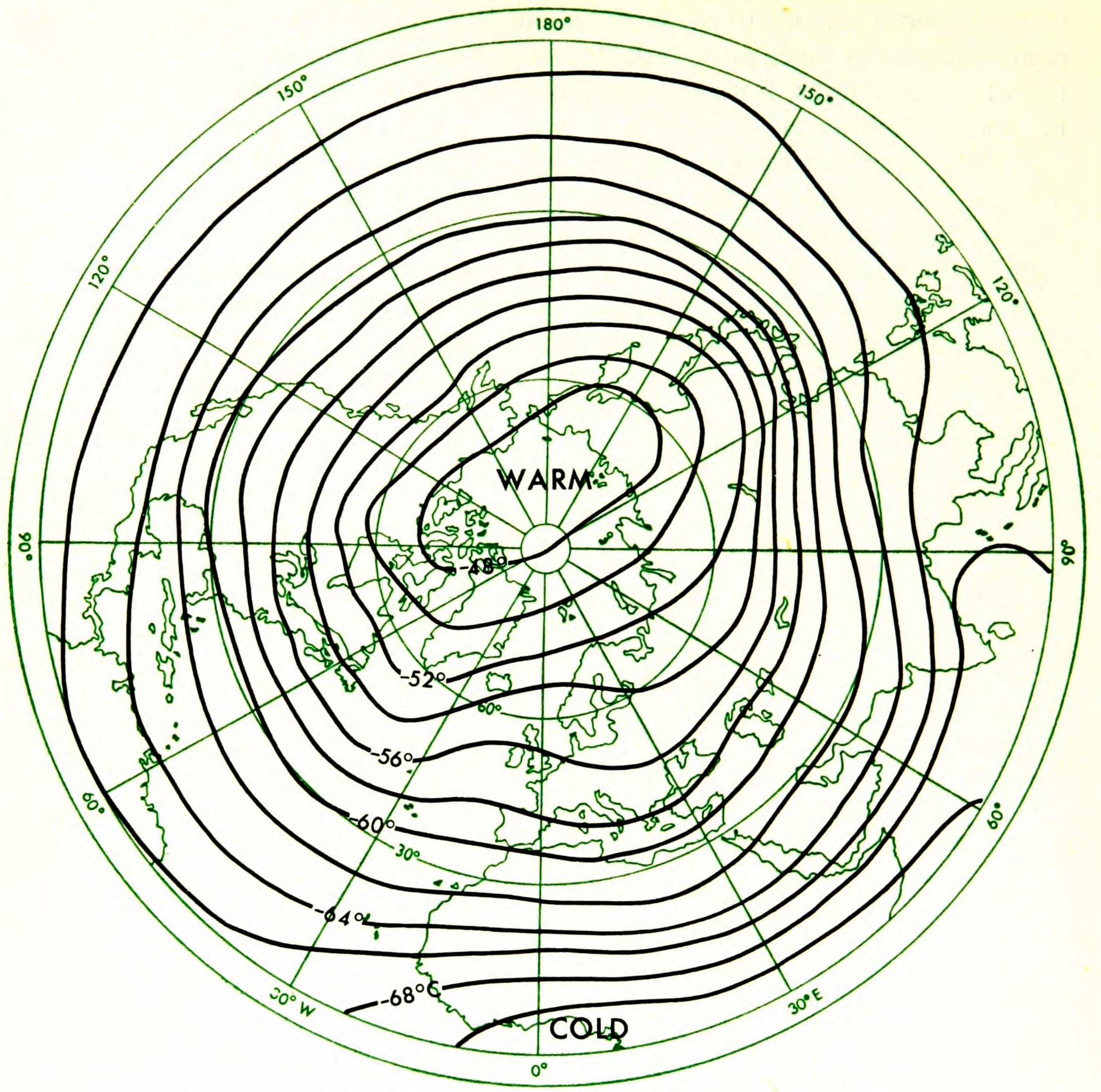


FIG. 66. *Average 50-millibar temperature, April, 1957–61*

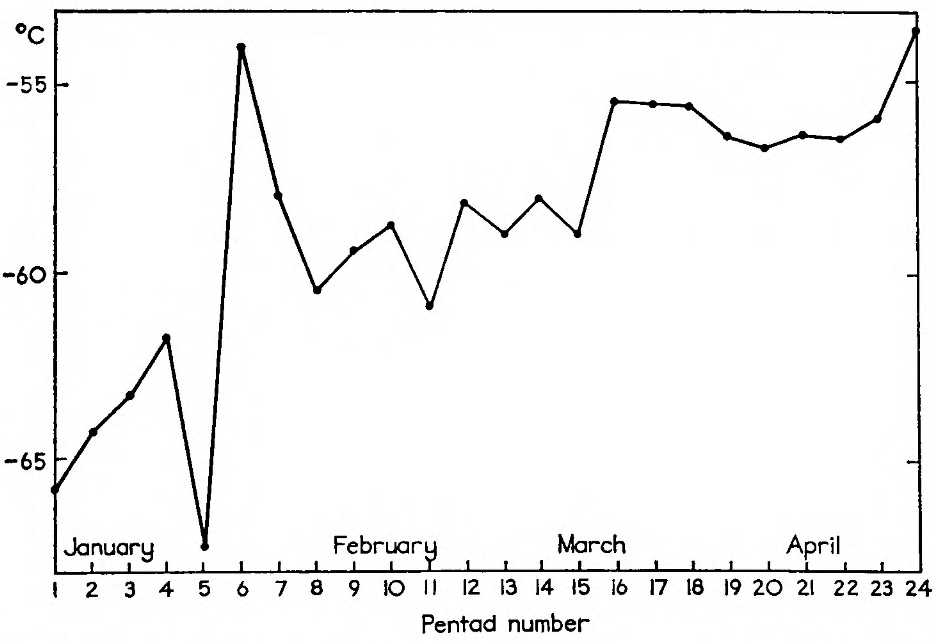


FIG. 67. *5-day average temperature at 50 millibars at Shanwell (or Leuchars), January to April, 1957–61*

The reversal of thermal gradient leads to a decrease and eventual elimination of the polar westerlies and the establishment of easterlies. The average time of onset of the easterly wind over the British Isles is late April, but this reversal can occur at widely different times from the end of January until mid-May. Records available up to the present time show that reversals occurring earlier than about mid-March revert back to westerlies before changing again to easterlies for the summer. Temperatures and winds at 50 millibars at Shanwell (56°26′N, 2°52′W) for the months January to April during the years 1957–61 are illustrated in Figs 67 and 68. Fig. 67 shows the 5-day means of temperature. Rapid warming occurs on the mean curve in late January but considerable differences in times of warming occur from year to year. In Fig. 68 the mean zonal wind component at Shanwell during this warming

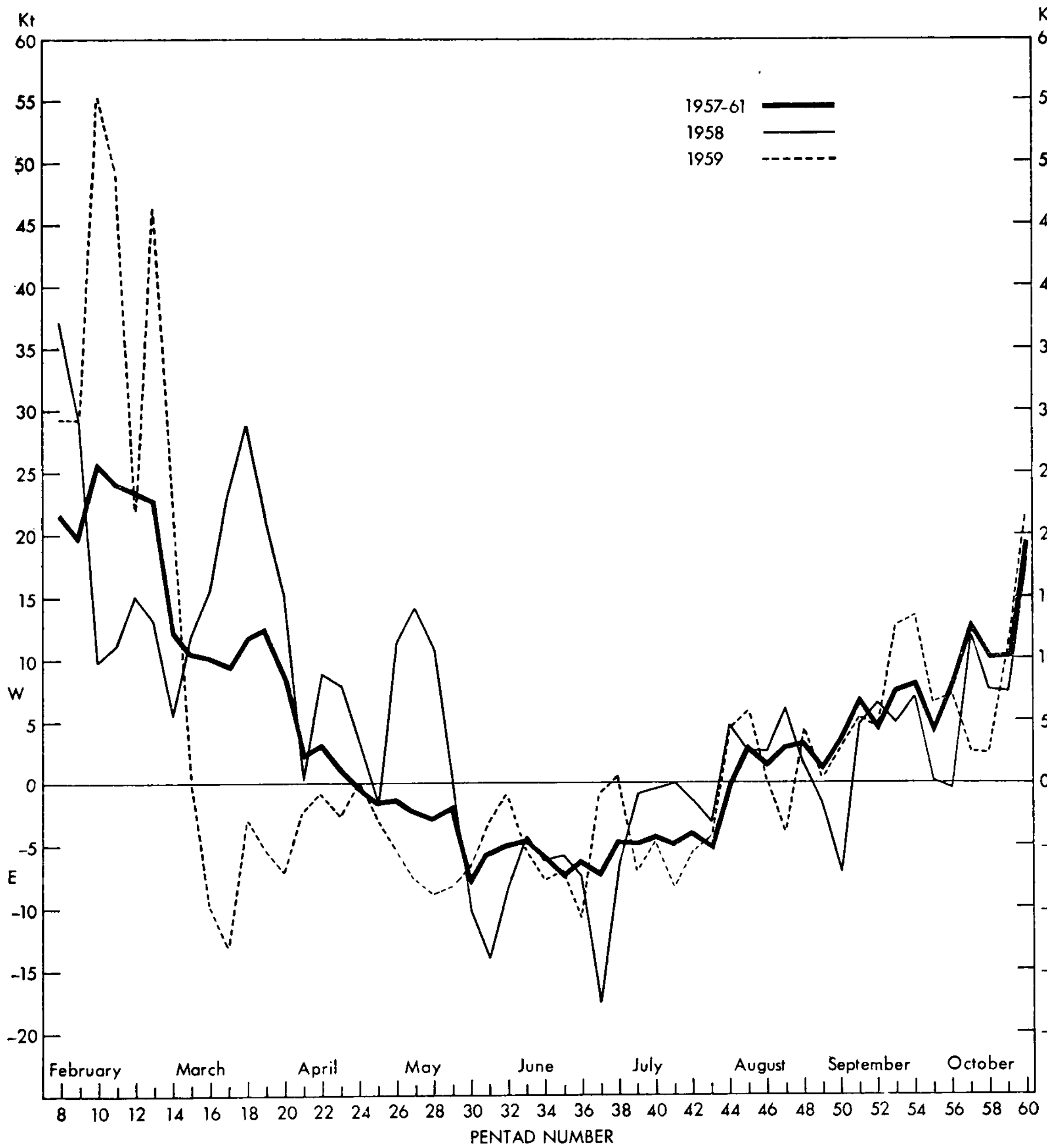


FIG. 68. *5-day average zonal wind components at 50 millibars at Shanwell (or Leuchars), February to October, 1957–61*

Westerly components are positive, easterly components are negative.
1958 and 1959 are also shown separately.

period early in the year shows a gradual slackening of the westerlies with a reversal to easterly occurring in late April; but the 5-day means for 1958 and 1959 indicate how variable the manner and time of change can be from year to year.

The average temperatures for July (Fig. 69) show a pattern which is broadly similar to that of April. The main feature is a warm area near the pole with a gradual decrease in temperature towards the equator. The predominant feature of the average winds at 50 millibars in July (Fig. 70) is an easterly flow over the whole hemisphere. The easterly flow is at a maximum near latitude 20°N with average speeds exceeding 40 knots. The tropospheric westerlies extend sufficiently high into the stratosphere to keep the mean flow westerly between 40° and 60°N almost up to 50 millibars (see Fig. 162, 100-millibar chart for July), but at 50 millibars in these latitudes the mean wind is light easterly and individual winds, although variable in direction, are also nearly always light.

As autumnal cooling commences, the westerlies replace the easterlies at 50 millibars and progress upwards and southwards, increasing in strength as winter approaches. By October the polar vortex is usually fairly well developed down to

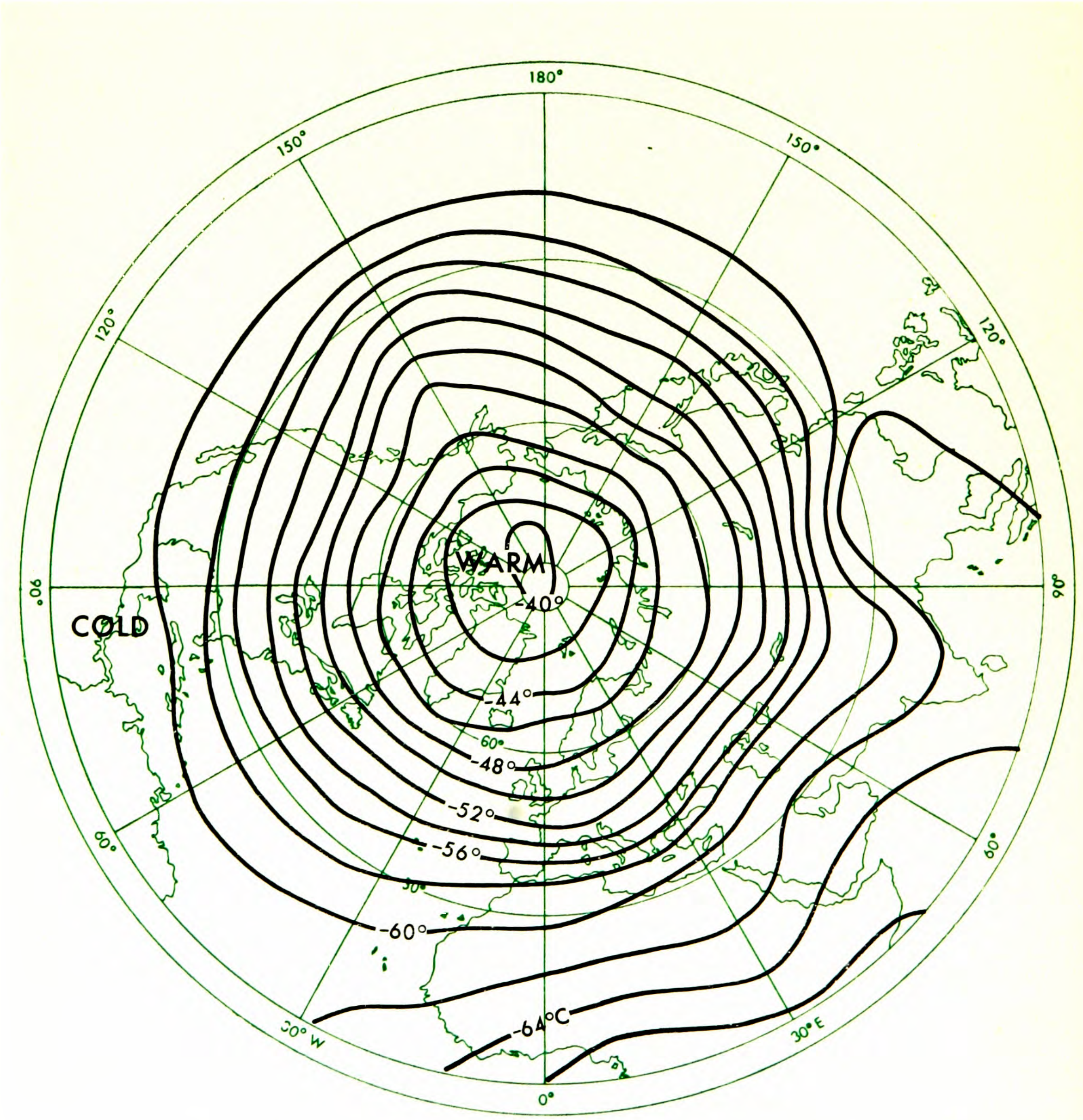


FIG. 69. *Average 50-millibar temperature, July, 1957–61*

about latitude 40°N, and a warm area is already established near the Aleutians. These features intensify steadily and, with a return to the January thermal distribution (Fig. 62), the westerly flow becomes well established north of about 35° to 40°N.

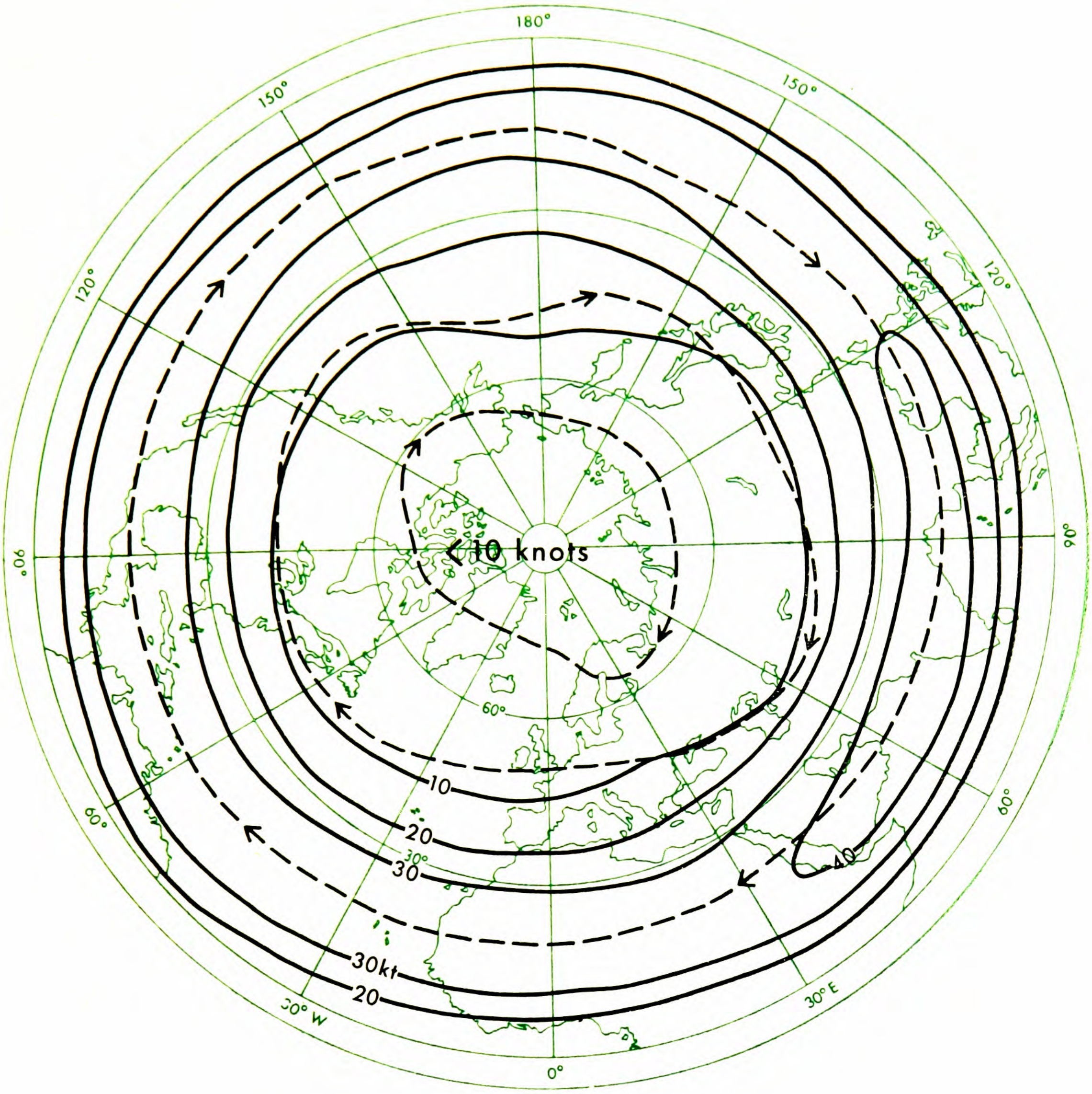


FIG 70. *Average 50-millibar wind, July, 1957–61*

——— Isotachs — — — Streamlines

Over the southern hemisphere, because of the absence of large land masses, the details of winds and temperatures vary from those of the northern hemisphere. The network of observing stations is very scanty but enough is known to suggest that there is a large cold cyclone centred over the pole during the southern winter with a westerly circulation covering most of the hemisphere. The jet-stream character of the stratospheric westerlies in winter is apparent and the breakdown of the cold winter vortex and the rapid warmings have been observed. Easterlies are established in the summer and a weak anticyclonic vortex appears to exist in this season in the southern polar stratosphere.

73. FLYING CONDITIONS AND NATURAL HAZARDS IN THE TROPOSPHERE AND LOWER STRATOSPHERE

Despite the general rule that most of the weather is confined to the troposphere, hazards from meteorological and other phenomena still exist for flight above the tropopause. Known hazards and the range of altitude in which they may be encountered are illustrated in Fig. 71.

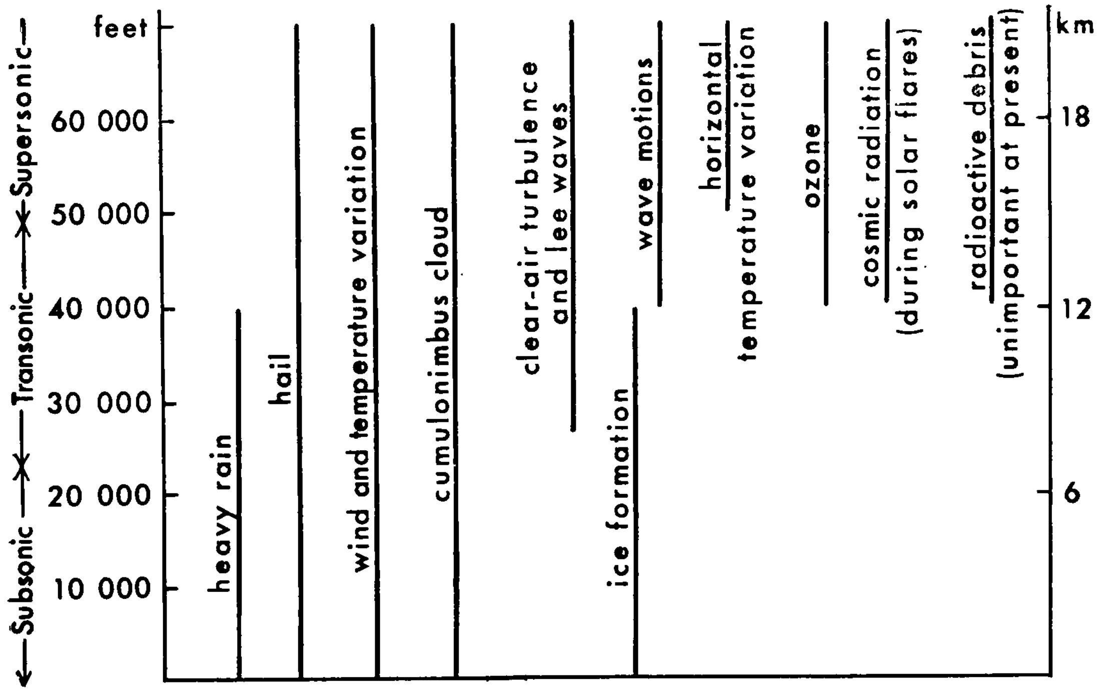


Fig. 71. *Extent of natural hazards likely to be encountered during subsonic, transonic and supersonic flight* (after R. F. Jones)

Turbulence in the stratosphere, although much less frequent than in the troposphere, is by no means negligible. It can be experienced near and above cumulonimbus clouds. Mountain lee-waves and waves above cumulonimbus tops which protrude into the stratosphere, although normally smooth, can break up into turbulent areas at times.

As aircraft speeds increase, the damage caused by heavy rain and hail increases. Hail can occur in cumulonimbus, and can also fall through clear air from the overhanging cirrus canopy. Airborne weather radar will assist in pin-pointing heavy precipitation, but a precise knowledge of the precipitation occurring within 100 miles of an airfield prior to take-off will only be available where weather radar is installed.

The thrust of a jet engine decreases with increasing temperature. Flight planning on the basis of the International Standard Atmosphere will give an optimistic estimate of the available thrust on days when the atmosphere is warmer. Large changes of temperature occur in clear air around and above cumulonimbus clouds and in the large-scale warmings of the stratosphere in winter and spring.

Intensity and location of a sonic boom, although dependent on aircraft characteristics, are affected by wind and temperature structure between the aircraft and the ground. The speed of sound is proportional to the square root of the absolute

TABLE 7. *Summary of flying conditions in cloud*

Cloud type	*Constitution*	*Continuity*	*Height of base*	*Vertical thickness*	*Horizontal visibility*	*Airframe icing*	*Turbulence*	*Remarks*
Cirrus (Ci)	ice crystals; rarely mixed*	detached	usually above 20 000 ft	usually thin, but may reach 5000–10 000 ft in low latitudes; may extend to tropopause	over 1000 m	rare, slight	nil or slight except when merging into Cb	may merge into Cs or Cb
Cirrostratus (Cs)		continuous						may merge into As
Cirrocumulus (Cc)	ice crystals, or water drops, or mixed*	layer clouds composed of detached globular masses						may merge into Cs
Altocumulus (Ac)	usually water drops to –10°C, mixed* at lower temperatures		6500–20 000 ft	usually thin	20–1000 m	rime, slight to moderate		
Altostratus (As)	usually ice crystals, occasionally mixed*	continuous, often 8 oktas cover	6500–20 000 ft but occasionally less than 6500 ft	up to 15 000 ft				often merges below into Ns
Nimbostratus (Ns)	water drops	continuous	surface to 8000 ft	merges into As	10–20 m	rime or moderate clear ice; possibly rain ice below cloud	severe near base, moderate elsewhere	envelops hills
Cumulus (Cu)	water drops	isolated, but may cover 6 oktas	usually 1500–5000 ft	up to about 15 000 ft	generally less than 20 m, and at times less than 10 m	rime or clear ice, possibly heavy; no safe lower limit to temperature	severe	large Cu may develop into Cb
Cumulonimbus (Cb)	mainly water drops to –15°C, mixed* at lower temperatures	usually isolated clouds 3–10 miles diameter; occasionally form a continuous line	usually 1500–5000 ft, but may be down to surface over water	15 000–30 000 ft or more, especially in low latitudes; may reach tropopause			severe or very severe within 10 miles horizontally and 5000–10 000 ft vertically	risk of lightning, severe 'static' and hail
Stratocumulus (Sc)	mainly water drops	layer of globular masses or rolls, often continuous	usually 1500–4000 ft	500–3000 ft	10–30 m	rime, moderate	moderate	sometimes penetrated by large Cu or Cb
Stratus (St)	water drops	continuous	surface to 2000 ft	200–1000 ft		rime, slight to moderate	nil or slight	envelops low hills

*i.e. containing both water drops and ice particles.

temperature but the track of the sound from any point is influenced by the local wind. Changes in wind and temperature as the aircraft climbs can cause a focusing of sound rays from different parts of the track, leading to an increase in the noise.

Concentrations of ozone above the tolerable limit are normal in the atmosphere from about 50000 to 100000 feet. Air fed to cabins may need to be filtered in some way to free it from excess ozone, but the high temperature in the compressor from which the cabin air is taken will assist by causing thermal decomposition of the gas.

Cosmic radiation of normal intensity is not a hazard, but at times of solar flares, outbreaks of radiation may be unacceptably high and flight at lower levels may be advisable to give the necessary screening.

Knowledge of meteorological processes is increasing steadily but much remains to be discovered, particularly about the lower stratosphere and the details of cumulonimbus clouds. It is clearly important that a pilot should have as full a knowledge as possible of when and where hazardous conditions are likely to occur. Such knowledge will be obtained partly from his own experience and from observations made during flight, and partly from forecasts and reports supplied before take-off and in flight. To some extent adverse conditions may be avoided by flight planning or by action taken during flight, but often risks have to be accepted and then fore-knowledge is important so that the necessary precautions can be taken in good time.

Flying conditions in cloud

These are summarized in Table 7. From this table some idea may also be gained of the usefulness of a knowledge of the various cloud types, particularly as flight above, within or below cloud may be demanded according to circumstances. It should be noted that it is difficult for a pilot to see cirrus cloud in which he may be flying, perhaps unknowingly.

CHAPTER 11

MISCELLANEOUS

74. SELECTION OF AERODROME SITES

Many factors come into the siting and design of an aerodrome which are independent of meteorology and are likely to be of paramount importance, but subject to these the meteorological survey falls into two categories. First, there is the broad climatological survey, as indicated by figures showing the frequency of occurrence of low cloud, poor visibility, wind and other elements, which embraces the district in which the aerodrome is to be located. It is aimed at choosing the best possible site within the given area and is limited only by the restrictions of a non-meteorological character, such as communication facilities. Secondly, when the site has been chosen, the lay-out of the aerodrome requires a survey of a more detailed character to decide how best to develop the site for flying purposes. It is not possible here to do much more than indicate broadly the nature of some of the many points to be considered.

Visibility

It is clearly desirable to choose a site as free as possible from fog and mist. Most low-lying and damp areas are particularly liable to radiation fog while coastal areas remain comparatively free, especially when backed by hills within some few miles which promote a katabatic down flow on quiet nights. On the other hand, a coastal site may be liable to fog which drifts in from the sea. The most likely region for this type is one where cold water lying off the coast promotes the formation of fog which is then brought inshore by the prevailing wind or by the diurnal sea-breeze. It may appear over the aerodrome as very low stratus, but this is almost equally adverse to flying operations. The aim in such cases is to choose a site which as far as possible is sheltered by high ground from the direction of fog-bearing winds. A similar remark, that protection will usually be provided by high ground to windward, applies to advection fog in general.

While a site elevated above the surrounding country avoids much of the radiation fog common to lower ground, too great an elevation entails the risk of frequent hill fog. This may be formed orographically, or cloud produced in other ways may envelop the high ground. Often high ground about 1500 feet or more above the general level of the surroundings is more subject to fog than is the low ground. A position particularly liable to hill fog is an isolated area of high ground where the prevailing wind blows off the sea.

The location in regard to towns and industrial areas should be considered because of the reduction of visibility brought about by smoke pollution. The aim should be to find a site which is generally on the up-wind side of the main source (or sources) of pollution in the area as far removed as other conditions permit. The canalization of smoke in valleys should also be remembered, as the concentration is there maintained over much greater distances than is the case when smoke is able to spread laterally in open country.

In arid countries, dust and sand fogs need to be considered in relation to both direction and strength of the winds. A survey of the soil conditions is necessary over a wide area around the site; in particular, a fine-textured loose soil to windward of an aerodrome causes much difficulty from dust carried by the wind. Even the movement of traffic on the windward side of such an aerodrome may cause trouble, but this can be overcome by suitable precautions.

Cloud

It is clearly desirable to avoid areas which are particularly subject to a high frequency of very low cloud. Such areas are usually those liable to the formation of orographic cloud, in particular high ground exposed to the prevailing wind, especially where this brings in air from the sea. A site which is sheltered from moist winds by high ground may be expected to have a reduced liability to low cloud (and also precipitation), largely because of the clearance associated with the föhn effect. Usually, over open country, the lower the elevation of the site the greater is the freedom from low cloud. When there is a widespread sheet of low cloud, not only is there more flying room between the cloud base and ground level over a low-lying aerodrome than over one at a slightly greater elevation, but in moist conditions the cloud base dips down over even the smaller hills (Section 36). This effect in an exaggerated form makes a tendency for cloud to form over the higher hills when there is no cloud at all at a similar level over the flatter areas. There is then an enhanced danger from cloud formation on isolated hills in the neighbourhood of an aerodrome when exposed to a current of moist air.

Wind

The effects of topography on wind have been explained in Sections 25 and 27. In the light of the information given there, it is possible to form an opinion from inspection of the area or from study of a detailed contour map whether a region is liable to locally induced strong or gusty winds which would cause difficulties to aircraft using any particular site. The local winds should be considered as modifications of the general wind distribution of the area. The main points concerned are as follows:

Valley winds. A site in a valley usually affords protection from strong winds, although gustiness may be somewhat abnormal and if the valley is long and straight a funnelling effect may cause a local increase of strength. If the valley penetrates into mountains, the diurnal winds should be considered, the katabatic blowing down the valley by night and the anabatic blowing up the valley by day.

Mountain-gap winds. Strong local winds are liable to occur opposite the mouths of mountain gaps. Their occurrence can normally be inferred from a detailed contour map, it being noticed that the gap must penetrate through the mountain barrier.

Headland winds. When a headland or the bluff of a mountain juts out into a plain, winds blowing along the mountain flank are materially increased in speed opposite the bluff.

Land- and sea-breezes. These are seldom strong enough to effect the choice of site of an aerodrome, except possibly where high mountains rise from a narrow coastal plain. Often the land- and sea-breezes constitute a main feature of the local climate and in places they have important consequences (for better or worse according to circumstances) in regard to the frequency of low cloud and bad visibility.

Vertical currents and turbulence. In this respect, regions of sharply varying contours should be avoided as far as possible. It should be remembered that turbulence generated by flow past an obstacle extends to a considerable distance down wind, and that it becomes well developed when winds are strong. With light winds, stationary eddies may form and the probable position of these should be considered.

Selection of alternates

In selecting a permanent alternate to a given aerodrome, the main consideration from the meteorological point of view is to choose a site which will rarely become unfit for landings, on account of weather, simultaneously with the original aerodrome. The only definite way of solving this problem is by comparison of a series of contemporaneous observations at the two sites, but in practice sites may have to be chosen before such data are available. It is necessary then to fall back on the following principles which emerge from the study of certain cases:

(i) The greater the distance of the alternate from the main aerodrome, the less the chance that both will be unfit together.
(ii) The simultaneous weather is often very different at two aerodromes located on opposite sides of high ground.
(iii) The weather at an inland place is often different from that at a coastal place at the same time.
(iv) Aerodromes on the opposite sides of an industrial area are less likely to be affected simultaneously or to the same degree by fog than are aerodromes on the same side.

75. DESIGN OF AERODROMES

Lay-out of runways in relation to wind

An aerodrome runway is considered to be unsafe for use by an aircraft when the wind component at right angles to it exceeds a certain critical value depending on the type of aircraft. In Fig. 72, if AOA′ is the direction of a runway and if PO

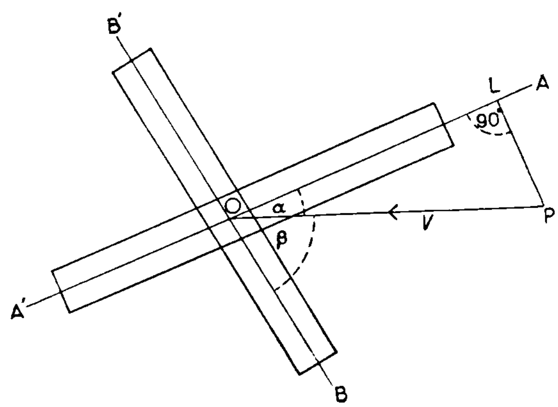


FIG. 72. *Runways in relation to wind*

represents direction and speed V of the surface wind, then the cross-component is represented by PL or $V \sin \alpha$. If the critical value of the cross-component is U, then the critical speed for a wind inclined to the runway at an angle α is given by

$$V \sin \alpha = U,$$

or

$$V = U \operatorname{cosec} \alpha.$$

If there are two or more runways to which the wind is inclined at angles α, β, . . ., then the critical wind speed is given by the smallest of $U \operatorname{cosec} \alpha$, $U \operatorname{cosec} \beta$, ...

The 'usability' of a runway in relation to wind is the percentage of time during which the cross-component is equal to or less than the critical value. To determine the usability, we require a table of frequencies of surface wind for the aerodrome, grouped into suitable ranges of direction and speed. Such frequencies should be the averages of observations made over a number of years. An example is given in Fig. 73 which shows a polar diagram divided radially into 12 sectors and by circles into zones of Beaufort force from 1 to 8. Frequencies of wind speed and direction

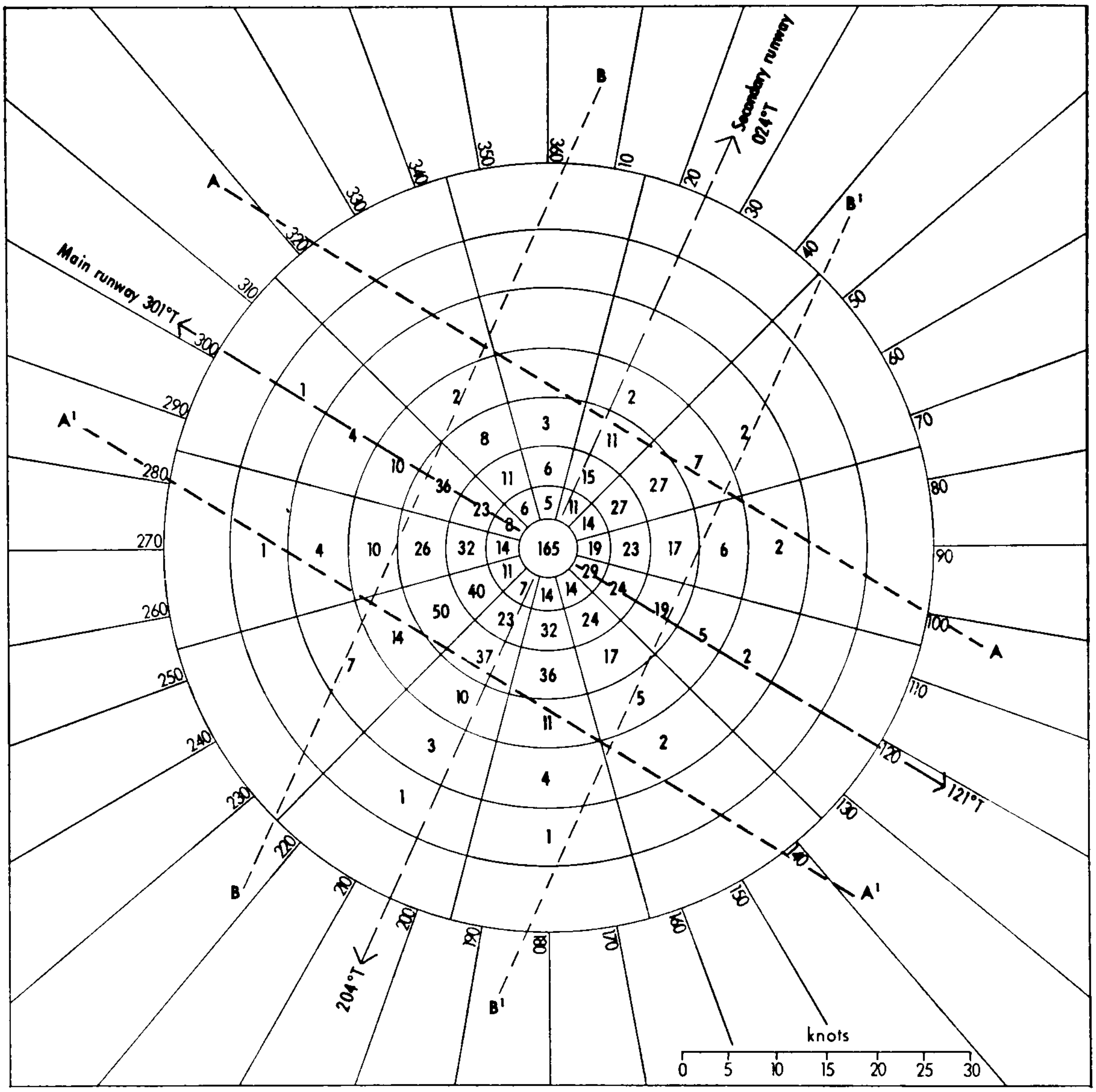


FIG. 73. *Usability of runways at Prestwick Airport*

from Prestwick Airport have been entered as the number of observations per thousand. Thus, for example, the number 26 in the 270° sector indicates that the frequency of winds of force 4 (11–16 knots) having a direction within 15° of 270° amounts to 2·6 per cent of all observations. On the diagram, two straight lines AA, A′A′ are drawn parallel to the main runway which is orientated 301°/121° true, at a distance either side of the centre corresponding to the speed of the critical cross-wind component which is taken for purposes of illustration at 15 knots. The radius vector of any point on the diagram represents a wind in direction and magnitude; if the point lies between the lines AA, A′A′, the component at right angles to the main runway is less than 15 knots while if the point lies outside the parallel lines, the cross-component is greater than 15 knots. Hence the usability of the runway is given by the total of the wind frequencies included within the parallel lines, expressed as a percentage. Interpolation should be used where necessary in determining this total which in the case illustrated amounts to about 93 per cent. The secondary runway may be dealt with similarly. Of the 7 per cent of occasions when the cross-component on the main runway exceeds the critical value, on all but about 1 per cent, the second runway would have been suitable for use. For the given critical cross-component of 15 knots, the two runways together therefore provide for use on 99 per cent of all occasions.

In designing an aerodrome, it is possible by trial and error to apply the above method to determine the orientations of runways so as to provide the maximum usability in regard to wind. However, a variation of a few degrees from the optimum orientation usually has little effect on the usability figure, and in deciding the actual layout due weight would be given to other factors such as the topography of the approach. The minimum number of runways required will be determined by the wind characteristics of the site, by the critical cross-wind component and by the coverage which the aerodrome is required to provide.

Siting of blind-approach systems

The radio installations which provide aids to blind approach to an aerodrome should be sited so as to be of use on as many occasions as possible. The placing is therefore influenced by the most frequent direction of the wind on occasions when poor visibility or low cloud produces blind-approach conditions. This direction is not usually the same as the direction of the prevailing wind. A meteorologist with experience of conditions at the aerodrome in question may be able to give a correct opinion as to which end of which runway should be chosen, but generally it is desirable to make a statistical examination of the association of bad visibility and low cloud with wind speed and direction. Such an examination will also show whether more than one blind-approach installation would be required to provide adequate coverage.

76. METEOROLOGY FOR HELICOPTER OPERATIONS

Meteorological factors of importance in the operation of fixed-wing aircraft have been emphasized from time to time in this and in the preceding chapters. The somewhat different considerations which apply (at the time of writing) to the operation of helicopters will now be described briefly under various meteorological headings.

Surface wind

Ground operations. Strong (over 20 knots) or gusty wind conditions necessitate caution in handling the helicopter on the ground. Under these conditions rotor blades are liable to undergo marked vertical oscillations (blade sailing) when rotating slowly during starting up and stopping, and to flap up and down when at rest. Both blade sailing and flapping constitute a serious hazard within the rotor diameter to personnel, equipment and aircraft fuselage. Conditions conducive to both of these phenomena are often found to the lee of hills, large buildings (hangars etc.), large trees and other similar objects. Certain types of helicopters are fitted with stops to confine these oscillations within safe limits in wind speeds up to about 40 knots. As wind speed increases, taxiing the helicopter across wind becomes progressively more difficult.

During take-off and landing. The speed and direction of the surface winds are of great importance for taking off and landing in very restricted areas. Particular care must be taken at night when marked cooling at the surface may result in the surface wind being much lighter and markedly backed in direction from the wind at a few hundred feet above the surface; the resultant shearing often gives turbulent conditions on the last stage of the descent.

Before a landing is made in a valley, the wind direction should be checked carefully, as valley winds can vary in direction to a considerable extent from the general wind.

The phenomenon of 'running snow' is encountered when a strong wind blows over frozen or partly frozen snow and 'grains' of snow are set in motion across the surface; a layer of moving snow is formed and the pilot receives the impression that he is moving forward, particularly if he is touching down in a large flat area of snow.

Snow may acquire a frozen surface layer as a result of melting in the sun and then freezing again into a thin layer of ice. Several landings will break up this surface and in strong winds the snow may 'run' on the landing area.

Turbulence

Turbulence has a serious effect on helicopter operations particularly when they are in mountainous regions, or at low levels in poor visibility, and when instrument or night flying is undertaken. Turbulence often dictates what flight path the helicopter pilot must choose to ensure the safety of his aircraft, particularly when approaching landing zones in mountainous terrain. It is possible that even entry into autorotation will not arrest a climb induced by a strong updraught, or conversely the application of full power may not prevent a helicopter from descending in a marked downdraught. Further, one of the most serious effects of turbulence is the effect on retreating-blade stall speed. An up-gust increases the blade's angle of attack and can cause a blade already operating at a high angle of attack to stall. It is on record that a helicopter with a computed stall speed of 80 knots stalled at 40 knots because of turbulence.

When operations are in mountainous regions particular attention must be paid to the possibility of standing waves (Section 27) or rotor-streaming turbulence which may make operations dangerous or even prohibitive. The probability of sharp changes in wind direction close to cliff edges makes them dangerous landing sites in strong winds.

Temperature

Increased air-intake temperatures will decrease the power output of a piston engine at a given throttle setting and for a gas turbine engine the same applies for a given compressor speed. This can limit the power available and affect the performance of the helicopter. As most helicopters on operations normally fly to maximum all-up weight, the effects of high temperatures, possibly accompanied by high moisture content, and of altitude on the air density (Chapter 4), and so on the performance of the helicopter, may have to be considered during the planning of an operation.

Low air-intake temperatures may be conducive to the formation of engine icing.

Visibility

If the horizontal visibility is less than 1500 metres, the cruising speed may be reduced, for reasons of safety, to below the economic speed with a consequent increase in fuel consumption and loss of endurance. It should be remembered that when there is smoke or haze beneath an inversion the haze is often thicker near the inversion than at the surface. Thus a pilot flying at, say, 500–1000 feet above ground and just below the inversion may experience a horizontal visibility appreciably less than that which prevails at ground level.

Cloud

As long as the ground can be seen, very low cloud bases are not in themselves a problem to the helicopter pilot, although they may force him to fly so close to the ground that turbulence may become a problem. Flying in cloud is generally avoided although there may be limited flying in stable stratified cloud. There is no planned flying in very turbulent cloud, nor in any type of cloud at temperatures below 0°C.

Diffused light conditions are very dangerous when flight is near snow-covered mountains. The light, broken up by a layer of cloud or haze, forms no shadows and this makes it difficult for the pilot to gain an impression of relief, or to judge distances to a point on the ground, which appears as a uniform grey layer.

Precipitation

If precipitation reduces the horizontal visibility below 1500 metres there may be loss of endurance as already explained.

When a flight is made at low speed in moderate or heavy snow or when the helicopter is hovering, snow may collect in appreciable amounts on the nose of the aircraft and may cover air-intake grills; this causes air starvation, loss of power and possibly engine failure. If this snow becomes compacted, possibly turning to ice, and breaks off and enters the engine, turbine damage may occur. When the ground is covered with dry powdery snow, helicopters taking off, landing, or flying very near to the ground often experience extremely poor visibility ('white-out') because of the recirculation of the snow.

Ice formation

The rotor blades are finely balanced to avoid vibration and the rotor head is subject to large centrifugal stresses. Even a comparatively small deposit of ice on the blades may have a serious effect on the performance by causing vibrations and increasing the centrifugal stress on the rotor head. Ice accretion on the head itself may interfere with the control of the blades. During running up on the ground, any vibrations set up by ice formation on the blades are liable to be transmitted to the springs of the undercarriage where sympathetic vibrations may become large enough to overturn the helicopter. There is danger also from pieces of ice leaving the blades at high speeds. Because flying in cloud is avoided, occasions of airframe icing are limited to flights at temperatures below 0°C in rain, drizzle, wet fog or snow. The deposition of ice on a helicopter on the ground may be expected to occur as for fixed-wing aircraft.

77. RADIO METEOROLOGY

The term 'radio meteorology' concerns the relationship between meteorological conditions and the transmission and reception of radio signals. As such it includes the production of radio echoes by clouds and precipitation, a subject which is dealt with in Section 87. The present section is confined to a description of the phenomenon known as 'anomalous propagation' or 'super-refraction' of radio waves.

In 'normal' atmospheric conditions, when the temperature decreases from the surface upwards roughly according to the normal lapse rate and there is no marked variation of humidity with height, the horizontal range of radio transmissions of wavelength less than about 10 metres is limited to the distance of the 'radio horizon'. Since even in these conditions the radio waves undergo a small amount of downward refraction, the radio horizon is somewhat below the optical horizon. The distance of the optical horizon from a point at height h above the earth's surface is $\sqrt{(2hR)}$ where R is the earth's radius; the distance of the radio horizon is greater than this by about one-third. When anomalous propagation occurs, the range may be many times the normal and for this to be possible the rays must be bent or refracted so that they remain near the earth's surface. Adequate downward bending of the rays occurs when the conditions near the surface are such that with increase of height either the temperature increases or the humidity decreases at rates greater than certain critical values. For temperature this value is almost 4 degC per 100 feet, the exact figure on any occasion depending on the actual temperature, pressure and humidity. If the humidity is expressed in terms of mixing ratio – the number of grammes of water vapour per kilogramme of dry air – then at a temperature of 15°C the critical value of the humidity lapse is 0·5 grammes per kilogramme per 100 feet while at other temperatures the value is somewhat modified. Thus for super-refraction to occur there must be a marked inversion of temperature or a decrease of absolute humidity with increase of height. In practice, the variations of temperature and humidity need to be considered together and the conditions to be satisfied then reduce to a single criterion, namely that the algebraic sum:

$$\frac{\text{temperature gradient}}{\text{critical temperature gradient}} + \frac{\text{humidity lapse}}{\text{critical humidity lapse}}$$

should exceed unity.

Special investigations show that either or both of the critical gradients are in fact exceeded in the surface layers on a small percentage of occasions in the neighbourhood of the British Isles, while in subtropical and tropical latitudes the conditions are satisfied much more frequently, especially over the sea.

Radio ducts

In most cases of anomalous propagation, the ray has a path of the type illustrated in Fig. 74. After leaving the transmitter at a small angle to the horizontal, the ray is gradually bent downwards in the super-refracting layer until it meets the earth's surface; here it is reflected and the process repeated. It is found that the direction of emission must be limited to within about half a degree of the horizontal, for otherwise the refraction is insufficient to keep the ray near the surface and it escapes into the upper atmosphere. The layer within which the state of super-refraction

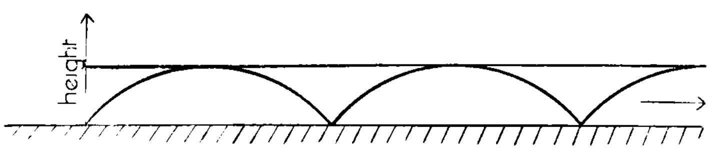


FIG. 74. *Radio propagation through a surface duct*

The ray is alternately refracted in the duct and reflected at the earth's surface (the vertical scale is exaggerated).

exists is called a 'radio duct'. It will be appreciated that the energy of the waves is effectively trapped within the duct, so that the signal strength is maintained over great distances with little attenuation. Naturally for anomalous propagation it is necessary for a duct to exist not only at the site of the transmitter but also over the whole range of propagation.

A radio duct does not necessarily rest on the earth's surface as in the illustration but may occur raised above the surface. Super-refraction is by definition limited to the actual duct, but rays emerging from its lower surface eventually re-enter the duct after reflection at the earth's surface. Below the duct, rays are usually refracted upwards. This does not interfere with propagation provided the duct is near enough to the surface, but if not the ray re-enters the duct at too great an angle to the horizontal and so escapes through the top of the duct. In practice the height of the upper boundary of a surface or elevated duct is often within 1000 feet of the earth's surface, although this figure may well be exceeded in certain regions.

The depth of duct required for abnormal propagation increases with the wavelength, being about 50 feet for a wavelength of 3 centimetres and about 600 feet for a wavelength of 1 metre. Since a surface duct is usually less than 1000 feet deep, the frequency of occurrence of the phenomena is greater at the shorter wavelengths.

Some practical results

Anomalous propagation has been reported from many places but the most striking results occur in tropical and subtropical areas and may be illustrated by observations made in the vicinity of India. There the meteorological conditions at certain times of the year are such as to give super-refraction almost continuously and days on which the range is 'normal' are unusual. Observations from Bombay with a 1·5-metre radar transmitter with a normal range of about 30 miles showed that echoes were received from the Arabian coast 800 miles distant (as well as from ships in the Arabian Sea) on 67 days out of 105 during a period in the early part of the year, while the extreme range exceeded 1500 miles. Similarly in Bengal super-refraction over land was reported with a similar installation on 232 nights in a certain year.

Effect of wavelength

Anomalous propagation is observed only with wavelengths from about 10 metres to less than 1 centimetre (30 to more than 30 000 megacycles per second) and the shorter the wavelength the more effective is the super-refraction. Waves longer than 10 metres (frequency less than 30 megacycles per second) depend on ionospheric reflection for propagation beyond the radio horizon (Section 3). On the other hand, short waves used for radar, even when not super-refracted, are not reflected by the ionosphere.

Meteorological conditions in relation to anomalous propagation

Since the conditions of temperature and humidity under which a radio duct forms are concerned only with a shallow layer near the surface, detailed observations are rarely available to indicate the existence of a duct at the position of the transmitter, still less over the extensive range for which anomalous propagation might be expected. The meteorological processes favouring the formation of ducts are, however, known and the first step is to be able to recognize when such processes are present. Some notes on these will now be given, but whether a useful duct is actually formed on any occasion is, apart from actual trial, a matter requiring quantitative investigation and cannot be gone into here.

Subsidence. When this takes place over the sea, the descending air is warmed while the sea surface temperature remains practically unchanged; hence a surface inversion is set up provided the subsiding air reaches a low level. Also the water-vapour content is high in the air in contact with the sea and lower in the subsided air, so that both temperature and humidity are favourable to duct formation. These conditions are commonly associated with anticyclonic areas and also with certain other types of flow including the north-east monsoon of Asia and the trade winds.

Advection. When relatively warm dry air moves over the sea, evaporation takes place and the resulting humidity lapse is usually sufficient to form a duct; in addition there may be a temperature inversion. This type is found in the vicinity of India and Iran, over the Mediterranean and off the coasts of most continents. When the air passes over coastal water after having been heated over land during the day, both the temperature inversion and the humidity lapse may be adequate for a duct, but the effects diminish as the air drifts farther out to sea.

If cold dry air is passing over a relatively warm sea, the temperature decreases with height and so is unfavourable for a duct, but this disadvantage may be outweighed by the strong humidity lapse. Whether or not a duct forms can be decided only by further investigation.

Over land, the advection of relatively dry air can be effective in producing a duct when the surface is wet, for example, after rain, for a humidity lapse is then established. Naturally these conditions no longer remain favourable after the ground has become dry.

Sea-breeze. An effect of the sea-breeze is to bring moisture inland in the lower layers for a short distance, so setting up a strong humidity lapse; in addition the humidity lapse over the sea surface is increased somewhat. Thus the humidity conditions become favourable for the formation of a duct, which, although limited to an area of a few miles either side of the coastline, may be useful for coastwise transmission.

Nocturnal radiation. This is effective in setting up the necessary temperature inversion over land at night. It is most pronounced in anticyclonic and other conditions associated with clear skies and light winds or calms, and a surface duct forms unless fog develops.

Fog. No surface duct forms in fog, since in this condition the temperature and humidity show little change with height. There may however be an elevated duct near the top of the fog, owing to the rapid falling off of humidity in this region. This is most likely to occur when the top of the fog is sharply defined, as with radiation fog; with advection fog the top is often diffuse and possibly merges into cloud, and no duct would then be formed. A duct elevated in this way can be effective only if the fog is shallow, for it has been seen that a duct which is raised too far above the ground is unable to give super-refraction.

Strong winds and turbulence may be mentioned as factors particularly unfavourable to the formation of ducts, since the tendency with them is towards a fall of temperature with height and a constant humidity mixing ratio. Surface heating is also unfavourable since it produces a temperature lapse, but on occasions this effect is overcome by that of a strong humidity lapse.

Nature of the surface. In addition to the meteorological conditions, the terrain should be considered. The sea presents an essentially flat surface, as the waves are too small to interfere with radio propagation; consequently the determining factors in the formation of a duct over the sea are solely meteorological. Similarly over desert areas the conditions may be almost uniform, but elsewhere over land the undulations of the surface and the variations in type of soil and vegetation make it impossible to assert the existence of a duct over an extensive area except in extreme conditions which should be easily recognized. Over rugged country the problem of duct formation becomes even more difficult, and moreover the method by which anomalous propagation takes place in these circumstances is not fully understood. In such situations the practical results obtained by radio operators are of the utmost value to the forecaster.

78. OPTICAL EFFECTS

Atmospheric optical phenomena provide a field of investigation of great scientific interest, but they have little practical significance in aviation although (Section 51) they occasionally provide some useful information on the constitution of the cloud particles concerned, information which is relevant to the risk of ice accretion. A brief account follows of some of the more frequent optical manifestations.

Rainbow

The common or primary rainbow is a circle, or arc of a circle, of coloured light, the centre of which is always in line with the observer's eye and the sun, while the radius of the circle subtends an angle of about 42° at the eye. The bow results from refraction of the sun's rays both on entering and on leaving a raindrop, together with one internal reflection. In the conventional bow the colours are those of the visible spectrum from red on the outside to violet on the inside, but the coloration is not the same in all rainbows, being dependent on the size of the drops. Brilliant rainbows with red on the outside are formed by drops of diameter greater than 1 millimetre. With drops about 0·3 millimetre in diameter the limiting colour is orange, while inside the violet there are bands in which pink predominates. With smaller drops, supernumerary bows of less brilliance may be seen inside the primary, but with drops of diameter about 0·05 millimetre the rainbow degenerates into a white 'fog bow' with faint traces of colour at the edges. As its name implies, this may be seen in fog opposite the sun, but it may also be seen on a sunlit cloud by an observer in the air. A secondary rainbow concentric with the primary is occasionally seen but with a radius of about 52° and the colours in the reversed order.

Halo

The term halo is used by the meteorologist only for the luminous effects produced by the refraction and reflection of light by prismatic ice crystals and hence in most cases by some form of cirrus cloud. The commonest case is a circle of light of angular radius 22° centred at the sun or moon. When well developed the halo round the sun shows a pure clear red on the inner side but other colours are usually difficult to recognize and the outside of the ring appears white. A halo may be seen in whole or in part on the average on one day in three at any one place in the British Isles. There is a large variety of other halo phenomena of less frequent occurrence, such as the halo of 46° radius, 'mock suns' or patches of light at the same elevation as the sun on the halo of 22°, vertical pillars and curved arcs of various radii. The combined effect is sometimes of spectacular beauty. It is now realized that these occurrences have no particular value for forecasting.

Corona

Coloured luminous rings of various angular radii may often be seen surrounding the sun or moon. They are formed by diffraction of light from water drops and the colours are usually dull. Red appears on the outside and blue on the inside, similar to the succession in the rainbow and opposite to that in the halo. They are most frequently seen when the watery sun or moon shines through a thin layer of altostratus cloud. The radius of the corona is inversely proportional to the size of the cloud droplets and the colours are purest when the droplets are of uniform size.

Glory

Another manifestation often observed from the air is a ring of colours surrounding the shadow of the aircraft on a cloud layer. This is called the 'Brocken spectre' or the 'glory'. It may be seen from an elevated position on the ground if the sun casts the shadow of the observer on a layer of low-lying mist or fog.

Mirage

Rays of light passing through the atmosphere are subject to a certain amount of bending on account of the varying refractive power of air of varying density. For this reason objects below the geometric horizon sometimes become visible and curious distortions of the sun near the horizon are common. Even in ordinary conditions, the sun is already below the geometric horizon before the disc begins to disappear at sunset. In abnormal conditions, refraction accounts for the phenomena of mirage.

Inferior mirage. Mirages are of two kinds. Inferior mirage occurs when air near the ground is much hotter and thus less dense than that immediately above, a common condition by day over a heated desert. Rays coming down from the sky at a gentle inclination may be bent up again to the observer to whom they appear to come from a bright water surface, while the images of objects near the ground are inverted. Mirages of this type may often be seen over smooth road surfaces on calm hot days.

Superior mirage. The opposite effect, known as superior mirage, occurs in conditions of a marked inversion of temperature, the light rays in this case being bent downwards. This is seen more frequently in polar regions. The image, usually inverted but sometimes erect, appears above the object. As the stratification which produces superior mirage is stable, the image is clear and well defined in contrast to the shimmering image of the inferior mirage; moreover, the distances are much greater so that the details can hardly be observed without telescopic aid. The phenomenon of anomalous propagation of radar signals (Section 77) bears an obvious resemblance to that of superior mirage in so far as the temperature effect is concerned.

PART II

METEOROLOGICAL OBSERVATIONS

CHAPTER 12

SURFACE OBSERVATIONS

79. OBSERVATIONS IN GENERAL

Meteorological observations are made for various reasons and the particular parameters which are observed and the way in which the observations are conducted at any particular place or time are determined by the purpose to which they are to be applied. Some observations are intended for climatological and statistical research, some for specific research application, and some for aviation and forecasting services. For aviation, detailed observations are required at regular and frequent intervals from a network of stations including selected airfields. There is a broad division into 'synoptic' and 'supplementary' stations. At the former, all the observations listed in Table 8 are made while at the latter stations only a selection of the observations may be made. Information on the location of land stations and ships which transmit observations is given in Chapter 14.

TABLE 8. *Observations made at synoptic stations*

Element	*Units (British)*	*Instrumental equipment or other aids*
Wind direction	degrees from true north	electrical anemograph, pressure-tube anemograph, or cup anemometer and vane
Wind speed	knots	
Pressure	millibars	mercury barometer or precision aneroid barometer
Pressure tendency and characteristic	tenths of millibar per 3 hours	aneroid barograph
Temperature and dew-point	degrees Celsius	dry-bulb, wet-bulb, maximum, minimum and grass minimum thermometers; thermograph and hygrograph
Visibility	metres or kilometres	known distances of fixed objects; transmissometers, Gold visibility meter (night only)
Weather phenomena, present and past	—	—
Cloud type	—	photographs available for comparison
Cloud amount	oktas of sky covered	—
Cloud height	feet	pilot balloon, cloud-base recorder, cloud searchlight (night only)
State of ground	—	standard patch of bare ground
Precipitation	millimetres	rain-gauge, rate-of-rainfall recorder
Sunshine	hours	sunshine recorder

Pilots and navigators should have a general acquaintance with the distribution of stations within the observing network and the types of observations made so that they may be aware of what information is available and how it is obtained.

In this book only a brief description of the techniques of making observations can be given but this should be sufficient to enable the reader to form an appreciation of the methods by which the raw material of meteorology is obtained. Detailed information is contained in the *Observer's handbook*, while technical descriptions of the instrumental aids are to be found in the *Handbook of meteorological instruments, Part I.*

80. PRESSURE

A full discussion of the significance of atmospheric pressure has been given in Chapter 2. It was explained that the barometric or static pressure at any point is equal to the weight of air contained in the column of unit cross-section extending above that point. In the usual instrument for measuring pressure, the weight of the overlying air is balanced against the weight of a column of mercury. The principle may be understood from a simple experiment: a glass tube about 3 feet in length and sealed at one end is filled with mercury, and the open end is then temporarily closed and immersed in a bath of mercury; on reopening the end beneath the surface of the mercury and securing the tube in a vertical position, the column of mercury will sink until its upper level is about 30 inches or 760 millimetres above the level of that in the bath. The space above the mercury is a vacuum since no air has been allowed to enter the tube, and the column is supported by the air pressure outside; its height, therefore, constitutes a measure of that pressure.

Mercury barometer

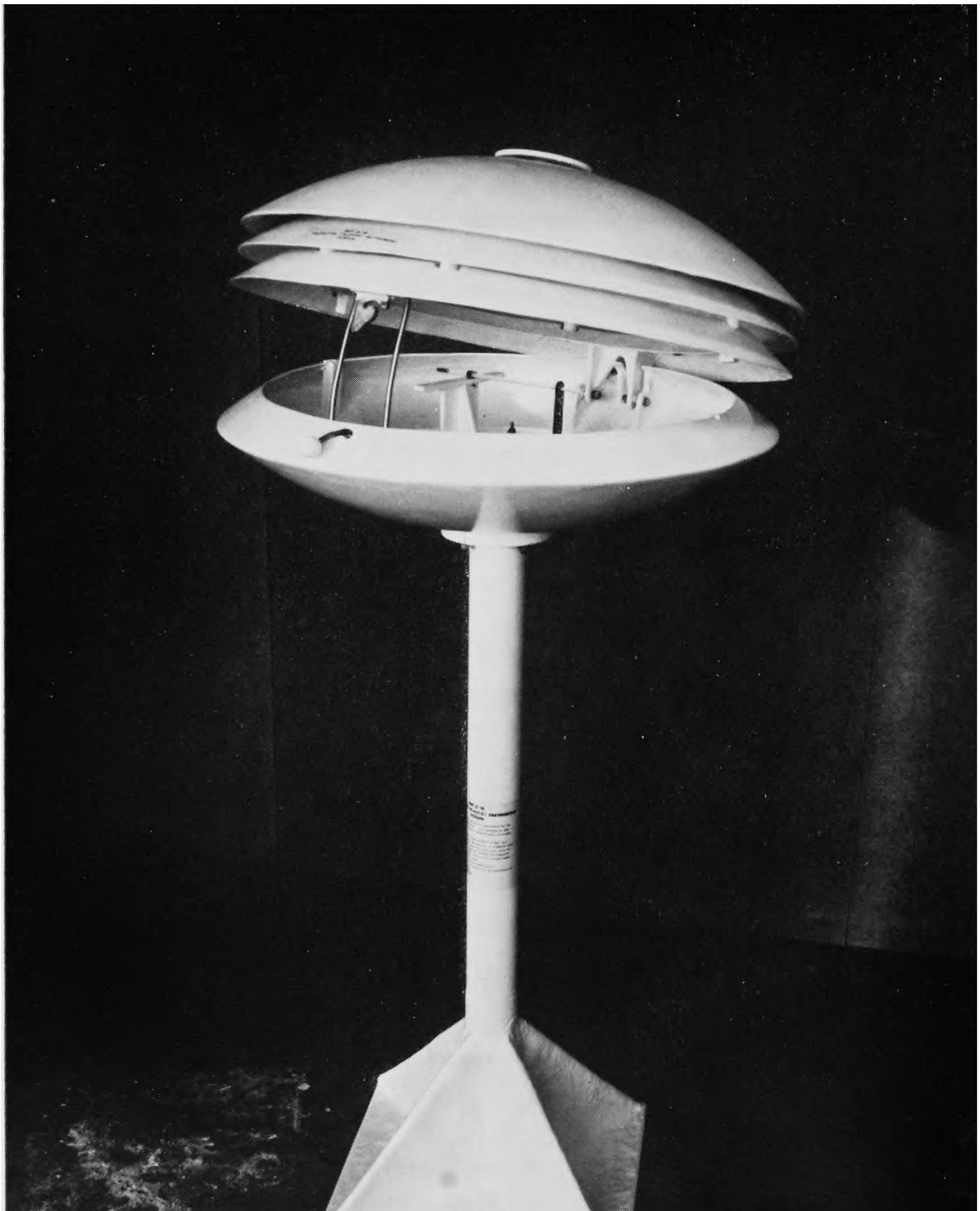
There are several types of mercury barometer in use and the student is strongly recommended to examine any to which he may have access. The form of instrument supplied to meteorological stations in the British Isles is known as the Kew pattern. The column of mercury is enclosed in a glass tube suitably protected and the instrument is swung on gimbals so that it may hang accurately in the vertical. The scale of the instrument, graduated in millibars (some are in inches), is read by means of a vernier attachment. The scale is designed to make automatic allowance for the varying level of mercury in the cistern which is rigidly attached to the tube, and no preliminary adjustment is required before a reading is made. Another type, the Fortin barometer, is slightly different in construction, and a preliminary setting is required to bring the level in the cistern to a fixed point defined by an ivory index or knife edge; this adjustment is made by screwing the plunger to be found at the base of the cistern.

Corrections required. Various corrections must be made to the indicated reading of the scale in order to obtain a value both accurate and comparable with those at other stations.

(i) *Correction for index error.* There are usually slight imperfections of the instrument, e.g. in the graduations. A certificate is provided with each instrument setting out the necessary corrections at different points on the scale.

(ii) *Correction for temperature of the instrument.* The height of the column of mercury at any pressure depends partly on its density which is governed by temperature. Expansion or contraction of the materials comprising the instrument also affect the reading. A thermometer is therefore incorporated

PLATES



Glass-fibre laminate thermometer screen (see Section 81)

PLATE I

Electrical anemograph (see Section 83)

PLATE II

A. Viaut, **Paris** (*France*), 28 *April* 1952, 1305 *local time* (*towards north-east*)

Cumulus humilis with haze

These cumulus clouds are scattered; most of them are fairly dense masses with definite horizontal bases. Their vertical extent is small and they are consequently of the species humilis. In the vicinity of the main clouds there are some isolated fragments (1, 2). Haze veils the distant units.

The station was in old maritime polar air on the south-western margin of a cold upper low centred over north-west Germany, but far from any front and in a zone of weak surface pressure gradients.

$$C_L = 1, C_M = 0, C_H = 0$$

PLATE III

M. Mézin, Paris (France), 3 April 1948, 0901 *local time (towards south-east)*
Cumulus mediocris and cumulus fractus

Cumulus is the only genus present. Some of the clouds are in the form of tufts of cotton-wool (species fractus), others are better developed and exhibit the beginning of bulging growth, already appreciable in places in spite of the early hour (species mediocris). Wind and turbulence cause asymmetry of form and raggedness of outline.

The picture is typical for maritime polar air behind a cold front; winds at lower levels were west-south-west, fairly strong (16 to 24 knots at the surface) and gusty. Thunderstorms with hail were observed in the same air mass.

$C_L = 2, C_M = 0, C_H = 0$

PLATE IV

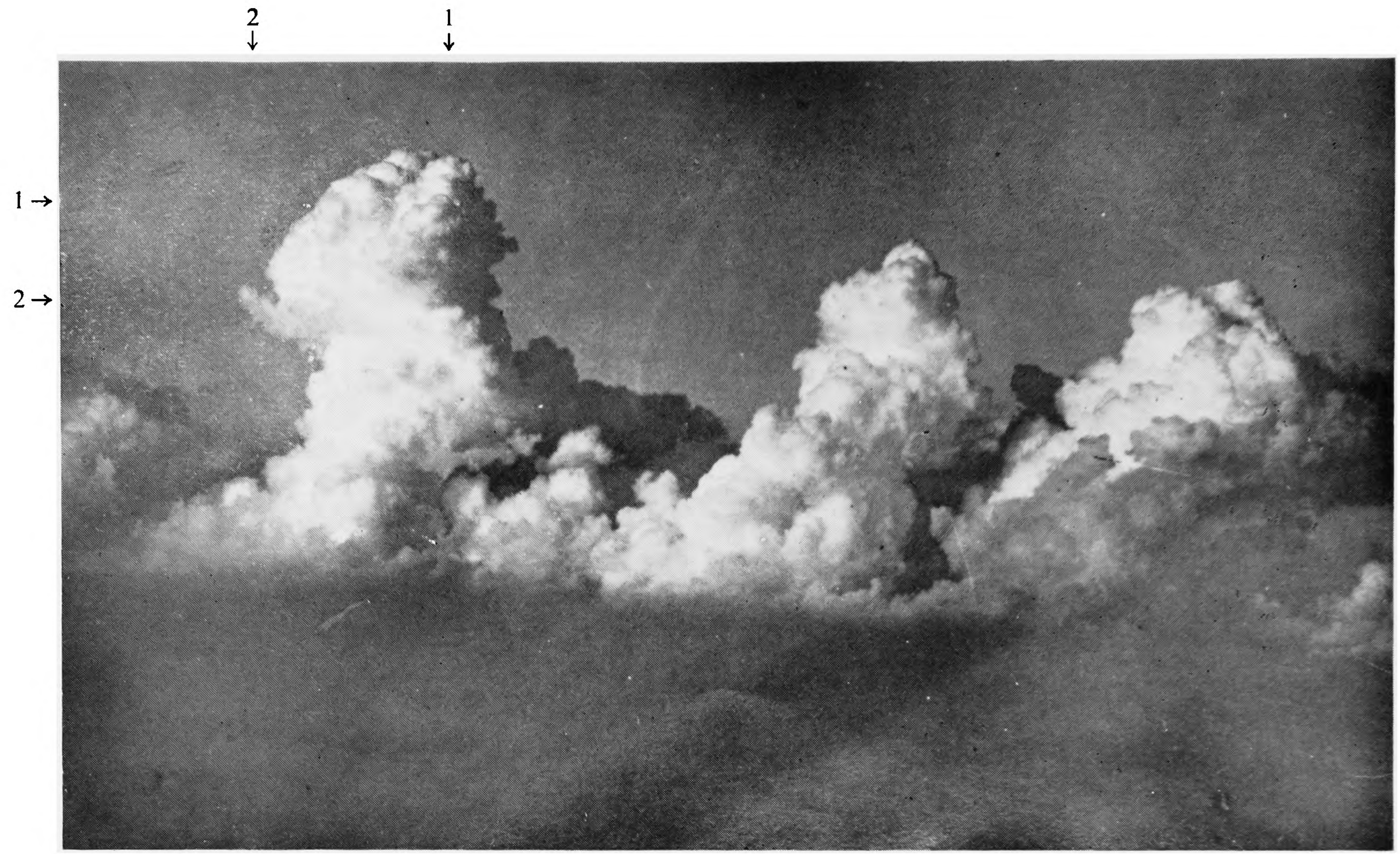
A. Viaut, above the Mediterranean Sea near Egyptian coast (33° N 28° 30′ E) *at* 18 000 *feet*, 16 *January* 1953, 1353 *local time*

Scattered cumulus clouds

The cumulus clouds have sharply outlined rounded tops; they are also dense, which is particularly clear from the dark shadows on the sea surface. There is considerable fragmentation on the sides of the clouds, presumably owing to mixing with the surrounding dry air.

Light westerly winds, associated with a flat low over Asia Minor, prevailed over the eastern Mediterranean.

PLATE V



C. K. M. Douglas

Cumulus congestus

The cumulus congestus ($C_L = 2$) show large vertical protuberances of irregular shape (especially at 1) which indicate the presence of strong ascending currents. The rising towers have mostly sharp outlines but there is some fraying out (as at 2), although without any fibrous appearance. The cloud bases are horizontal. The photograph was taken from an aircraft over the sea and the shadow of the aircraft wing is seen in the left foreground. The cumulus towers themselves also cast deep shadows. The highest reaches 10 000 feet.

PLATE VI

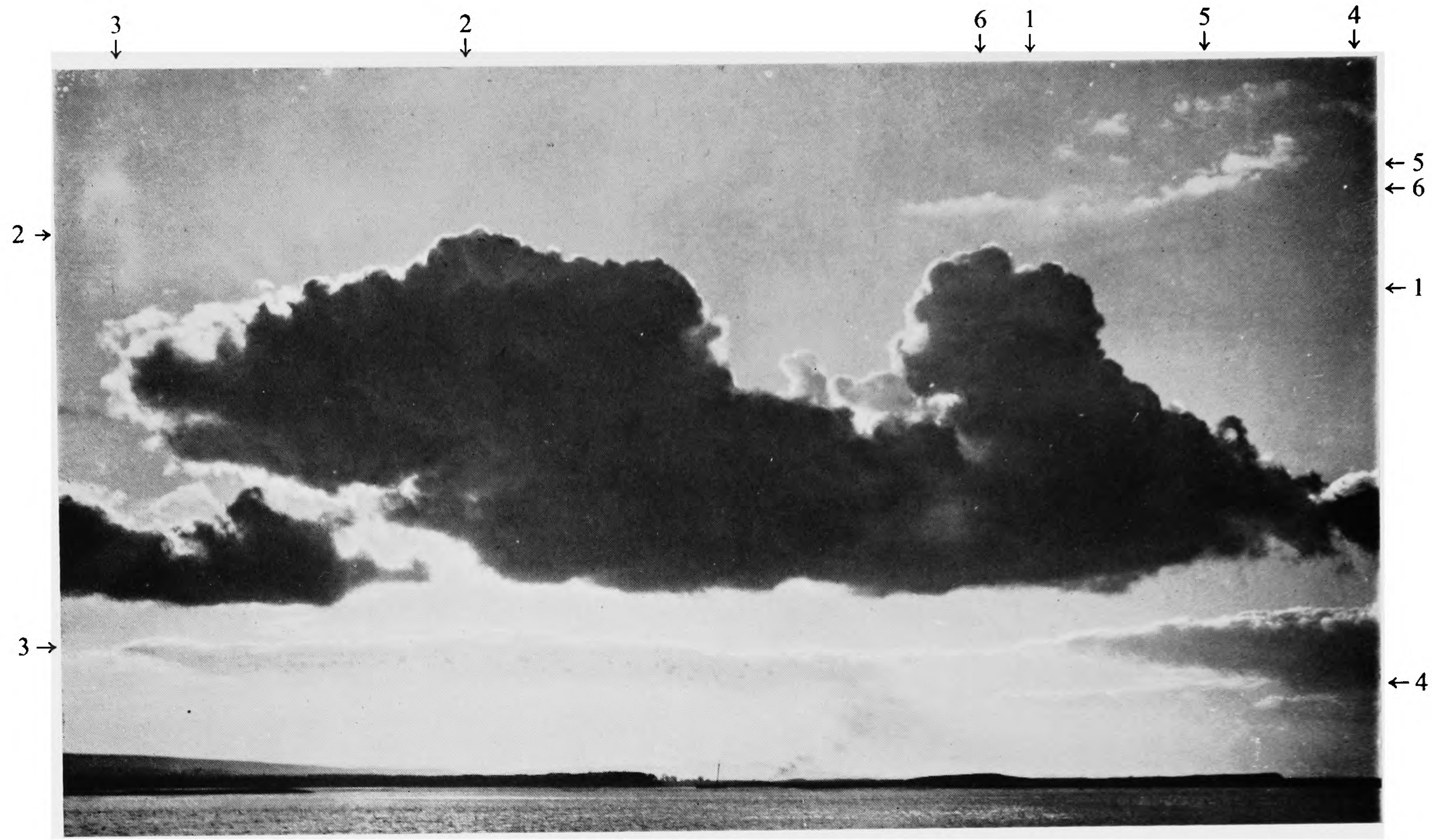
R. L. Martin, Hayes (Middlesex, U.K.), 16 *May* 1957, 1445 GMT (*towards south-east*)
Cumulonimbus calvus

A strongly bulging mass of cumulonimbus breaking through patches of stratocumulus cumulogenitus (1, 2) formed by the spreading out of smaller cumulus clouds. The convective cloud is starting to lose its sharp outline at 3, 4 and there is a suggestion at 3 of the beginning of an anvil formation. Cumulus fractus are present at 5, 6.

The clouds were observed in maritime polar air on the western side of a weak ridge of high pressure.

$$C_L = 3, C_M = 0, C_H = 0$$

PLATE VII



W. J. Day, near Bournemouth (U.K.)
Stratocumulus cumulogenitus
The cumulus congestus clouds (1, 2) are spreading out at the base, producing stratocumulus cumulogenitus. Other parallel bands of stratocumulus are seen nearer the horizon (3–4), their flattened tops suggesting the presence of an inversion. Some fragmentary medium cloud shows a cumuliform structure at 5 (altocumulus floccus), while at 6 it is tending towards altocumulus castellanus.

$C_L = 4, C_M = 8, C_H = 0$

PLATE VIII

G. A. Clarke

Stratocumulus stratiformis opacus undulatus

This stratocumulus consists of a series of well-defined parallel undulations forming a continuous or almost continuous layer (species stratiformis), the greater part of which completely masks the sun (variety opacus).

$C_L = 5, C_M = 0, C_H = 0$

PLATE IX

Royal Air Force, off Cape St Vincent (Portugal), at 2000 *feet*, 26 *July* 1944, 0641 GMT
Stratocumulus stratiformis undulatus (from above)

The upper surface of the level cloud sheet shows a series of regularly arranged, nearly parallel rolls (variety undulatus). The photograph was taken in air drifting southwards ahead of a cold front situated 150 miles to the north-west.

PLATE X

A. Viaut, Meudon (France), 1 *January* 1950, 0902 *local time* (*towards south-east*)

Stratus translucidus and mist

The greyish cloud layer is evaporating and has already broken up into large, somewhat dark portions (1, 2) which are too big, too indistinct and too irregular to suggest stratocumulus. The cloud is already sufficiently transparent to reveal the position of the sun and even its actual outline (3). Beneath this thin stratus the air is misty.

The station was in an anticyclone, in which the sky in general was only slightly cloudy. The surface wind was light north-east.

$$C_L = 6, C_M = x, C_H = x$$

PLATE XI

G. A. Clarke

Cumulus fractus of bad weather (pannus)

The low clouds are mostly rounded but with ragged margins, and they show deep shading; the cloud elements appear to merge together in the lower part of the photograph. A layer of nimbostratus or altostratus is visible through the gaps. The sky is of threatening appearance and the presence of ragged cumulus beneath the layer cloud makes the term 'pannus' appropriate.

$C_L = 7, C_M = 2, C_H = x$

PLATE XII

1 2

2 →
1 →

P. Stahl, *Greenland* (69° 45′ N 50° 16′ W), 23 *September* 1951, 1500 *local time* (*towards south-west*)

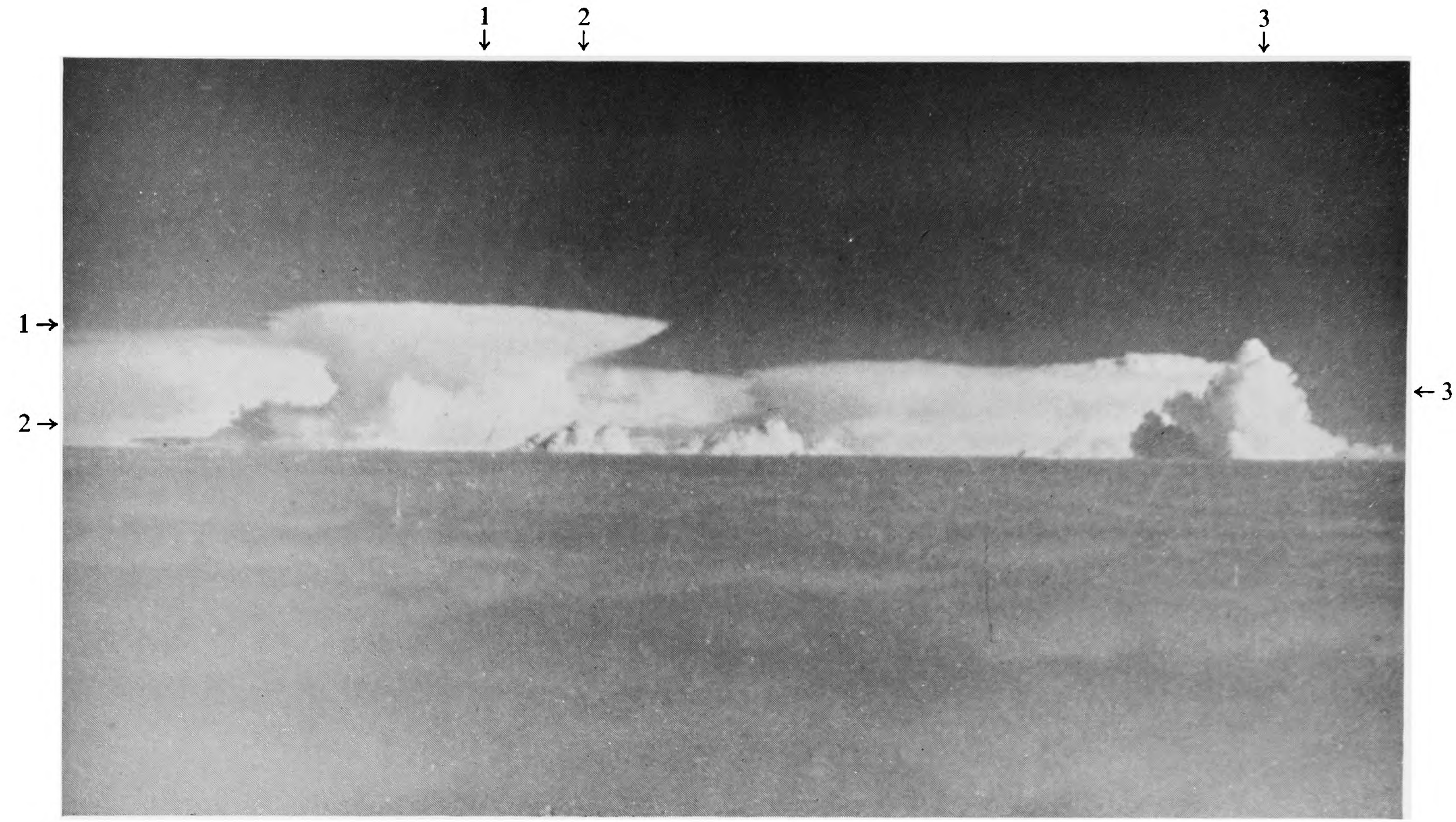
Orographic stratocumulus and altocumulus (wave clouds)

Except for some cumulus fractus at 1 and 2 near the horizon, all cloud patches show a lenticular form with fibrous margins. The darker clouds, being rather heavily shaded, belong to the genus stratocumulus; the lighter clouds are altocumulus. A foliated structure is visible in various places. These typical orographic clouds are often observed on the west coast of Greenland, with wind blowing from the inland ice towards the sea.

On this particular day there was a high over east Greenland, while pressure was relatively low over Baffin Bay and the Davis Strait. At the station, the sea-level wind was light from the south-east, the temperature relatively high (10 °C) and the relative humidity low (30 per cent).

$C_L = 8, C_M = 4, C_H = 0$

PLATE XIII



Royal Air Force, *Bremerhaven* (*Germany*), 27 *August* 1944, 1755 GMT, *from* 12 000 *feet*
Cumulonimbus capillatus incus

The continuous stratiform layer has its top at 11 000 feet. Rising through this layer is a belt of massive cumulonimbus associated with a slow-moving cold front. The (obscured) base of the convection cloud is at 4000 feet and the anvil tops reach to 32 500 feet. The upper parts of the cumulonimbus are clearly cirriform (species capillatus) in the shape of an anvil (incus); the anvil at 1 is almost ideal. Convection cloud in earlier stages of development (cumulus congestus) is present at 2, 3.

PLATE XIV

G. A. Clarke
Altostratus translucidus, cumulus humilis and cumulus fractus

This altostratus is sufficiently translucent to reveal the position of the sun. At a lower level, there are dark clouds lit from behind; most of these are typical cumulus humilis—they have more or less horizontal bases (1, 2) and their tops are flattened or only slightly rounded. Some cumulus fractus are seen at 3, 4. The coding of the low cloud as $C_L = 7$ implies bad weather conditions associated with precipitation occurring at or about the time of observation.

$$C_L = 7, C_M = 1, C_H = x$$

PLATE XV

French Meteorological ***Service***, ***Paris*** (*France*), 11 *August* 1949, 0820 ***local*** *time* (*towards south*)

Nimbostratus

The dark grey cloud layer has a faintly striated appearance (1, 2) which is rendered diffuse by continuously falling rain.

The rain area near the occlusion of a disturbance, crossing France from west to east, affected the district around Paris.

$C_L = 0, C_M = 2, C_H = x$

PLATE XVI

Royal Air Force, North Sea, at 10 500 *feet*, 31 *July* 1945, 1951 GMT
Altocumulus stratiformis translucidus perlucidus

Many of the elements of the altocumulus show shading, but it may be assumed that the position of the sun could be determined through the greater part of the layer (variety translucidus). Moreover the sky can be seen through spaces between some of the elements (variety perlucidus), although in places the elements have become merged. At a lower level is seen the upper surface of a continuous layer of stratocumulus.

Pressure was high to the west of the British Isles. The observation was made in the vicinity of a weak occlusion lying north–south over the North Sea.

PLATE XVII

G. A. Clarke, Aberdeen (U.K.), 29 *October* 1927, 1510 GMT

Altocumulus lenticularis

Some of the lenticular patches are darkly shaded in their central parts (1, 2) and show no detailed structure: others are without shading and consist partly of small rounded elements (3). The lenticular shape of the altocumulus is most probably due to stationary waves, caused by orography. Above the altocumulus there are some filaments of fine cirrus (4) and a thin veil of cirrostratus extending down to the horizon.

The sky was observed in maritime polar air with fresh south-west winds, 300 miles ahead of a warm front type occlusion.

$$C_L = 0, C_M = 4, C_H = 6$$

PLATE XVIII

A. Viaut, *Saint-Palais-sur-Mer* (*France*), 22 *July* 1950, 1108 *local time* (*towards north-north-east*)
Altocumulus stratiformis perlucidus undulatus

The cloud layer is made up for the most part of fairly large rolls (1–2, 3–4), roughly rectilinear and parallel. Clear sky is visible between the rolls (variety perlucidus). The clouds were progressively invading the sky.

The photograph was taken in a maritime polar air mass between a cold front and an occlusion, both of thundery character.

$$C_L = 0, C_M = 5, C_H = 0$$

PLATE XIX

A. Viaut, near Tarbes (France), 24 *July* 1951, 1705 *local time (towards south-south-west)*

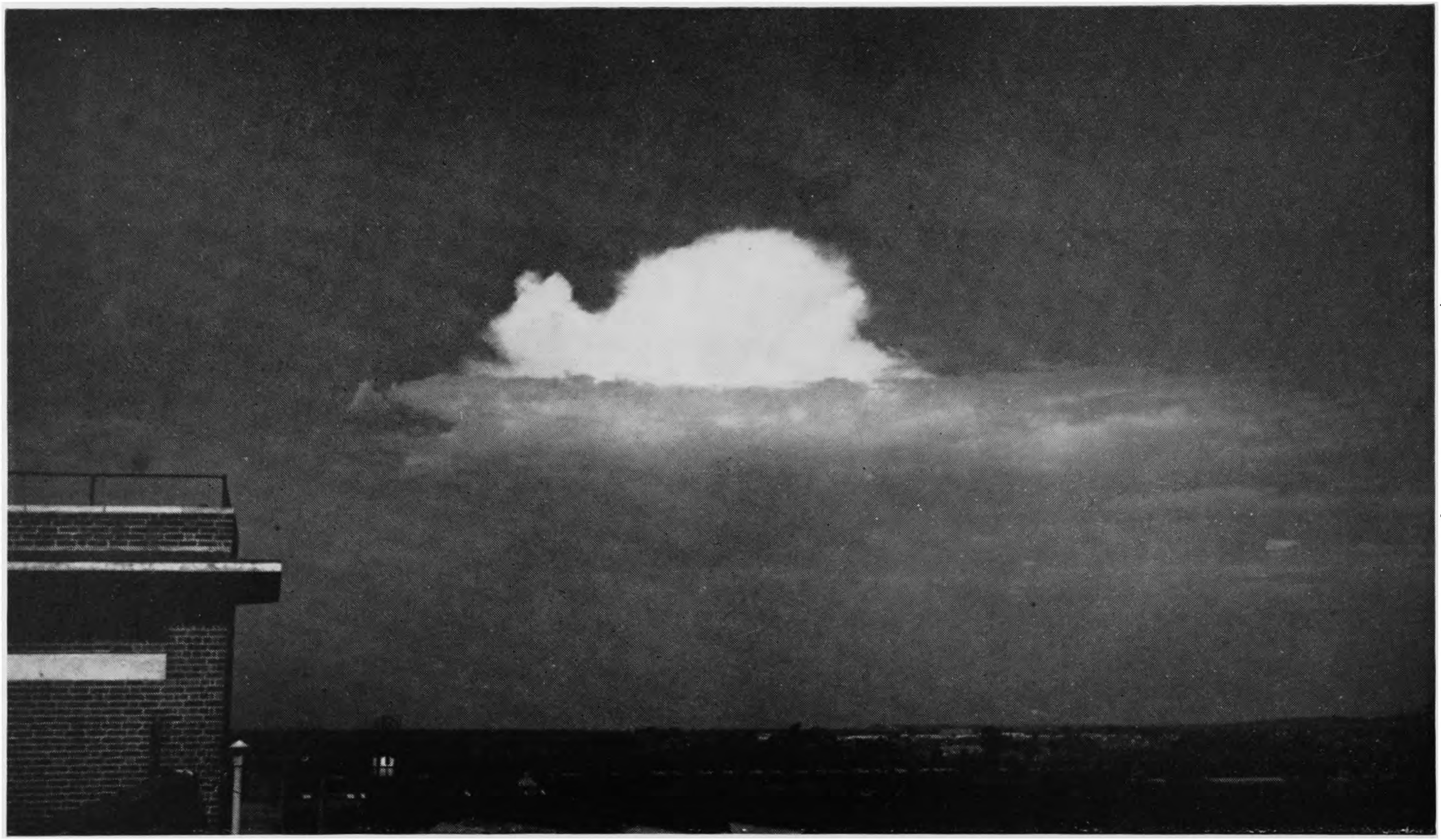
Altocumulus cumulogenitus

The principal patch of altocumulus (1) resulted from the spreading out of the top of a moderately developed cumulus (2). Patches of altocumulus (3, 4) on the margin of the main patch are disintegrating. At 5 a whitish sheet constituting the top of a cumulonimbus capillatus, can be seen.

The station was situated in a northerly flow of polar air on the eastern side of a high over the Bay of Biscay.

$$C_L = 9, C_M = 6, C_H = 0$$

PLATE XX



Royal Aircraft Establishment, **Farnborough** (*Hampshire, U.K.*), 13 *July* 1945, 0840 GMT

Altocumulus castellanus

A patch of altocumulus with base at 8000–10 000 feet has developed strongly in the vertical (1) (species castellanus). Other patches of altocumulus are dimly visible, as at (2).

An area of low pressure covered the Bay of Biscay and Spain, while pressure was high over the North Sea. Slight thunderstorms occurred in southern England during the following night.

$$C_L = 0, C_M = 8, C_H = 0$$

PLATE XXI

French Meteorological Service, **Paris** (**France**), 11 *September* 1950, 1500 *local time* (*towards north*)

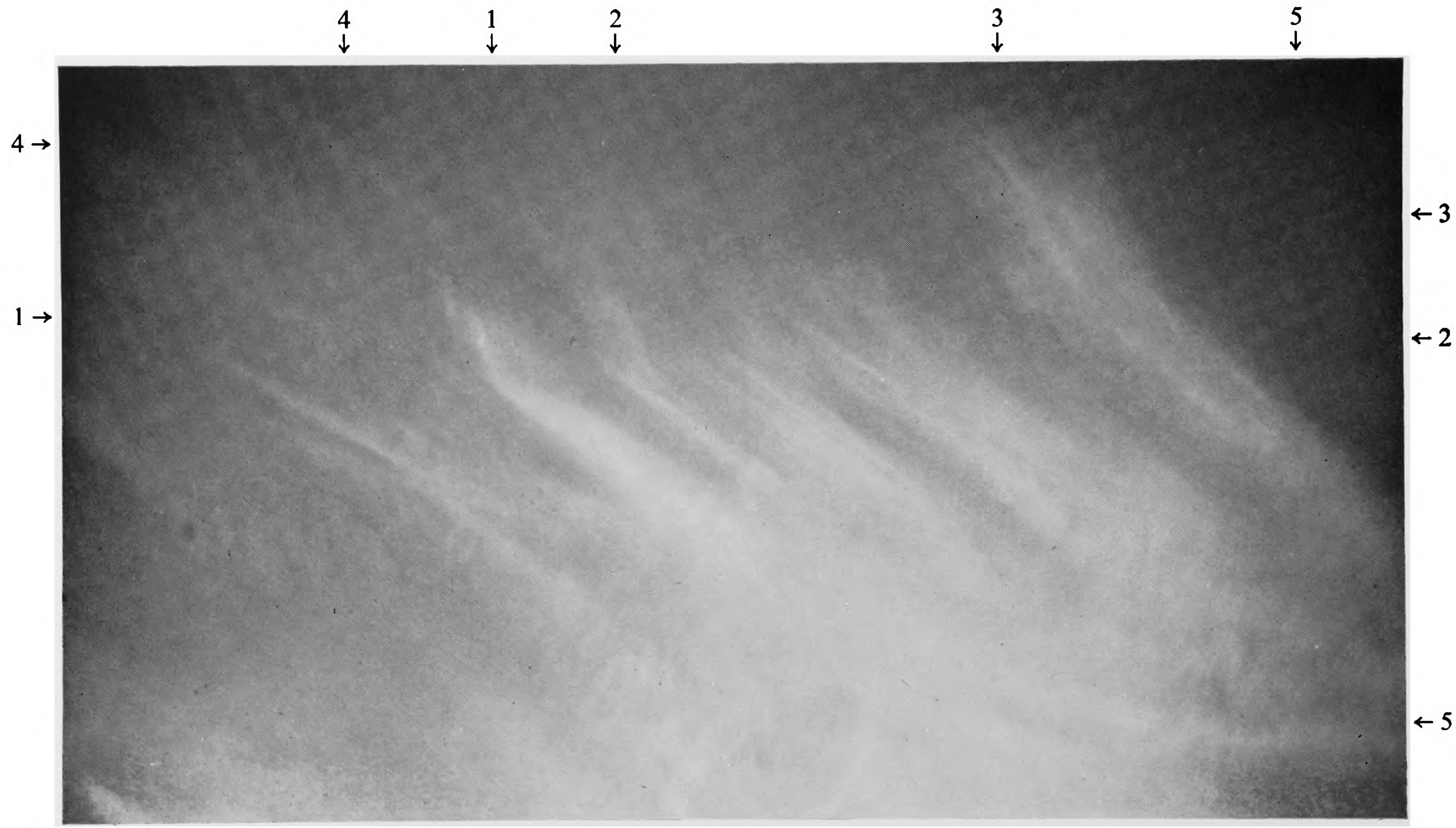
Chaotic sky

The sky has a chaotic appearance due to the presence of clouds of different genera and species at various levels. Cirrostratus, apparently originating from a cumulonimbus anvil, is present at the top of the photograph (1–2), and the white patch at 3 with mamma at 4 is presumably cirrus spissatus cumulonimbogenitus, brilliantly illuminated by the sun. The altocumulus is of varying opacity and occurs in the shape of elongated spindles (5), cumuliform sproutings (6, 7) and fine ripples (8, 9). At 10 there is cumulus mediocris in the shadow of higher clouds.

A depression was moving from west to east over the British Isles. Many thunderstorms developed over France in the unstable maritime tropical air moving from the south-west.

$C_L = 2, C_M = 9, C_H = 8$

PLATE XXII



Royal Aircraft Establishment, Farnborough (Hampshire, U.K.), 9 February 1945
Cirrus uncinus

Most of the parallel filaments in the centre of the photograph terminate in hooks or tufts (1, 2, 3) (variety uncinus). Further trails are dimly visible, as at 4. The clouds are progressively invading the sky (coding $C_H = 4$) and in the foreground they have merged together (5); since a fibrous structure is still present, this cloud layer would be classified as cirrus fibratus.

The clouds observed were several hundred miles in advance of a warm front approaching from the south-west.

$$C_L = 0, C_M = 0, C_H = 4$$

PLATE XXIII

T. Bergeron, Stockholm (*Sweden*) (63° 20′ N 12° 30′ E), 6 *September* 1943, 0815 *local time* (*towards south-east*)
Cirrostratus fibratus and cirrocumulus

A thin, hardly visible cirrostratus layer, having a fibrous appearance as a whole, completely covers the sky; part of it (1, 2) is obscured by patches of cirrocumulus. In the cirrostratus the lower part of the 22° halo and the lower tangent arc are visible at 3 as a brighter illumination. Near the horizon some cumulus (4–5) and stratocumulus (6) are seen.

The station was 370 miles ahead of an occlusion slowly approaching from the west.

$C_L = 8, C_M = 0, C_H = 7$

PLATE XXIV

Royal Air Force, over U.K., at 36 000 *feet*, 28 *February* 1950, 1245 GMT, *near Benson* (*towards west*)

Royal Air Force, over U.K., at 36 000 *feet*, 28 *February* 1950, 1300 GMT, *near Newbury*, 25 *miles south-west of Benson.*

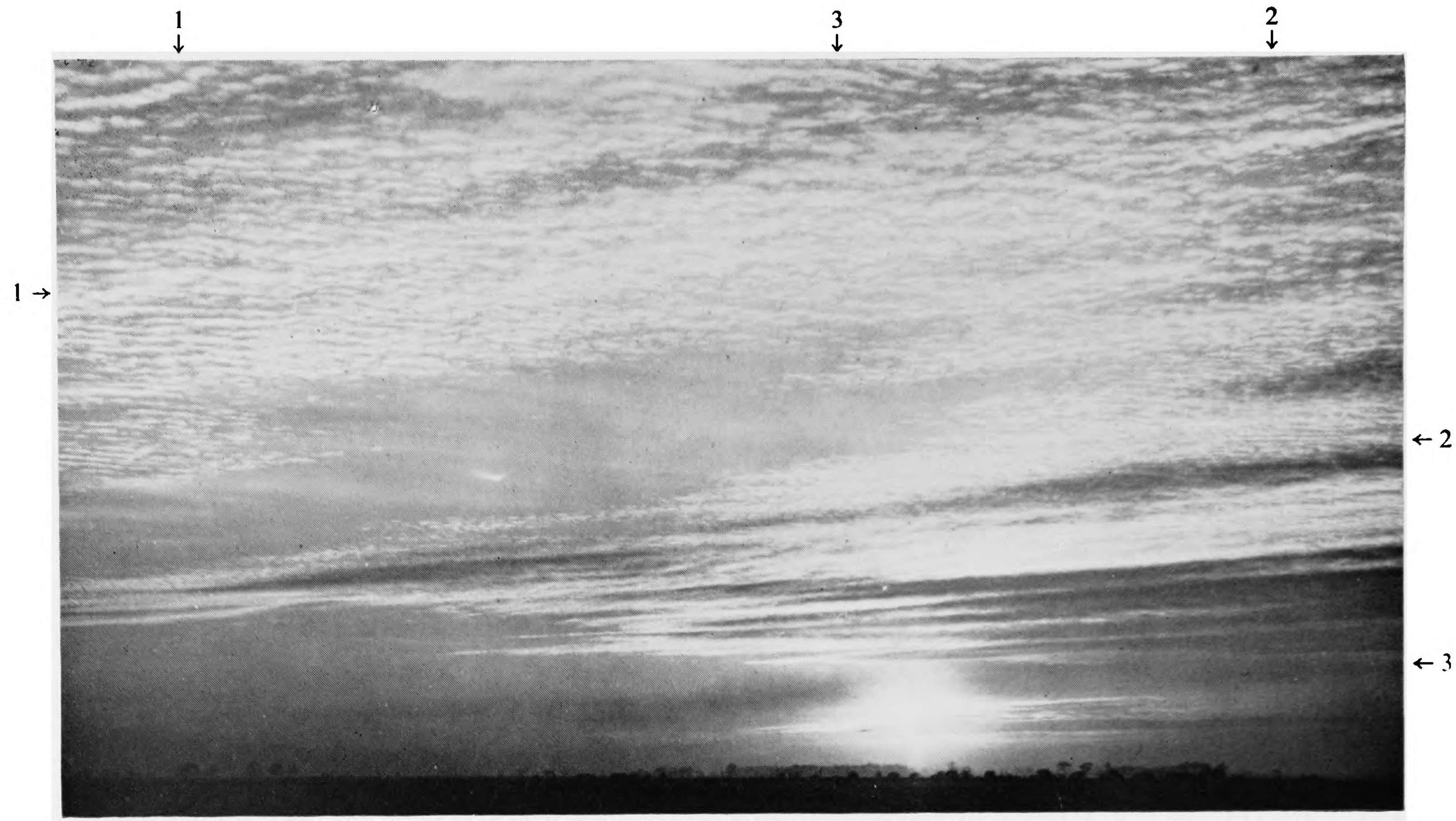
Broken layer of mainly stratiform clouds

The altitude of the upper surface of the clouds in both pictures is approximately 23 000 feet. The left-hand picture shows at 1 the discontinuous and fibrous structure of the cirrus. At a greater distance the cirrus clouds appear closer together and near the horizon long straight bands of cirrostratus are observed. Apparently the same bands, with waved upper surface, cross the right-hand picture from upper left to lower right. Between the main bands rounded elements of cirrocumulus, presumably at a somewhat lower level, can be seen.

In both pictures the cloud is broken and mainly stratiform. A weakening frontal system, extending from Iceland over the Irish Sea towards Spain, was approaching slowly from the west. Pressure was high over Scandinavia.

PLATE XXV (*a*)

PLATE XXV (*b*)



Royal Air Force, Cranwell (Lincolnshire, U.K.), 15 November 1954, 1600 GMT (*towards south-west*)
Cirrocumulus stratiformis undulatus

The cirrocumulus forms an extensive sheet (stratiformis) without shadows; in several places (1, 2) the individual elements are arranged in an undulatory pattern (variety undulatus). The elongated clouds with sharp edges and deep shadows (as at 3) are altocumulus lenticularis.

At 1200 GMT there was a warm front along the east coast of Ireland and an anticyclone over the Midlands. The cirrocumulus increased from one okta at 1300 GMT to three oktas at 1600 GMT. Rain associated with the warm front began at Cranwell at 2340 GMT.

$C_L = 0, C_M = 4, C_H = 9$

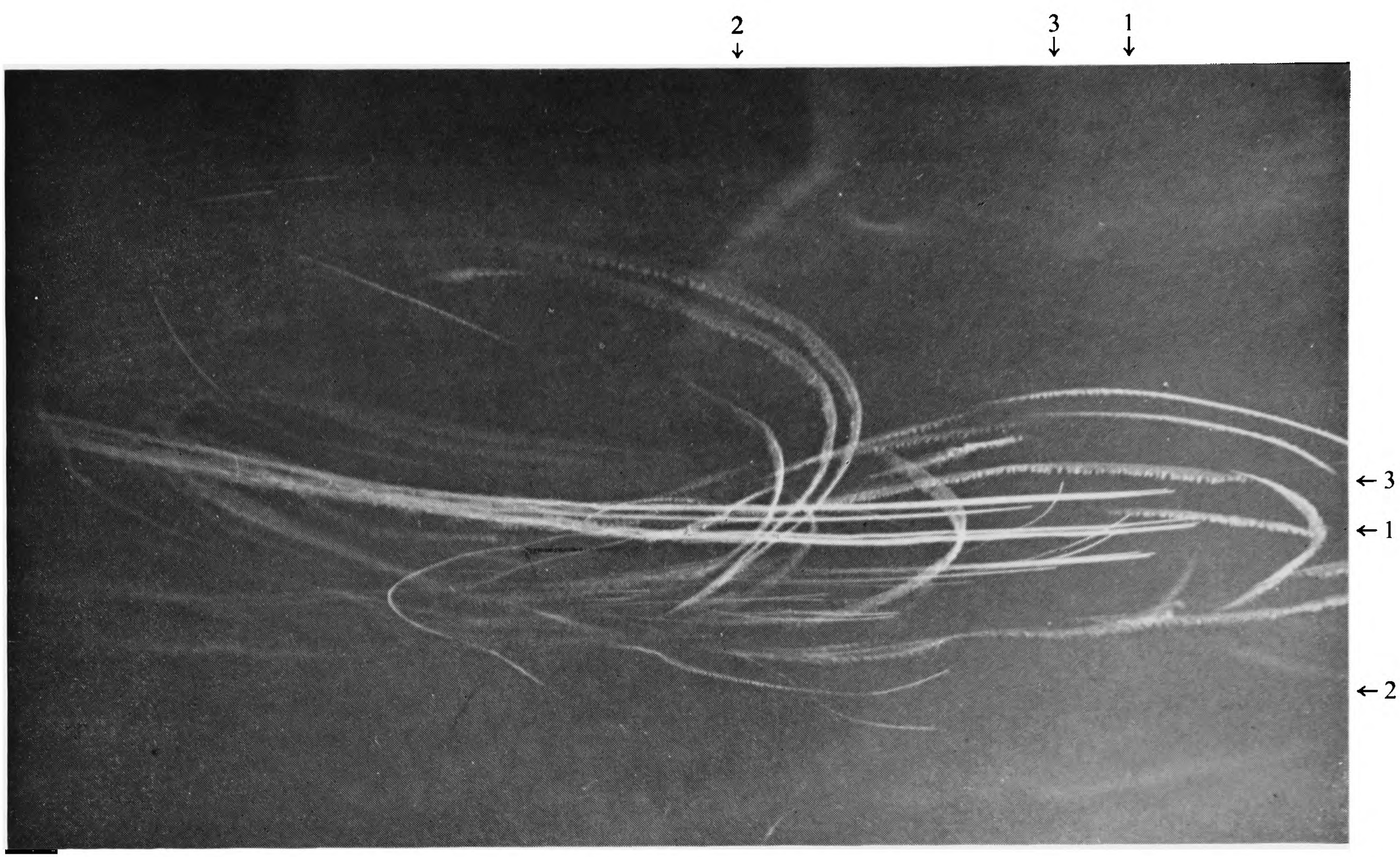
PLATE XXVI

Royal Air Force, 75 *miles south-south-east of Damascus* (32° 24′ N 36° 49′ E), *at* 10 000 *feet*, 17 *April* 1951, 1400 *local time*

Wall of dust

The distinct forward edge of the base of the wall of dust can be seen at 1–2; it is dark and slightly saw-toothed. The wall itself reaches up to 11 000 feet, and all parts of it are formed of grey turbulent masses of dust raised into the air, probably without any condensation of water vapour. The phenomenon was caused by a cold front, moving slowly towards the south-east. The invading polar air skirted the eastern edge of an anticyclone.

PLATE XXVII



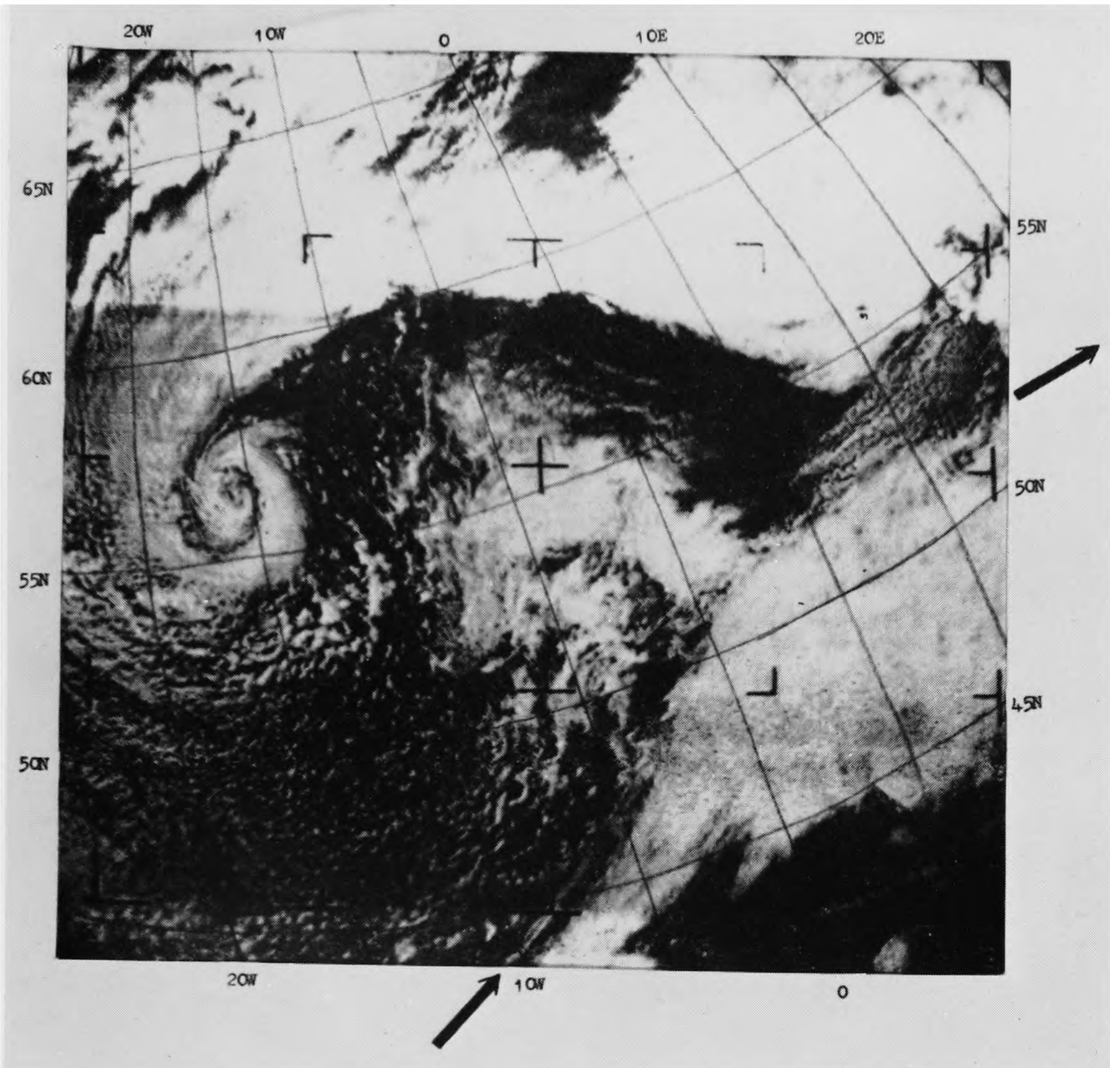
Royal Air Force, Uxbridge (Middlesex, U.K.), 25 *September* 1949, 1559 GMT (*towards south-south-west*)
Condensation trails (contrails)

Persistent contrails are seen at about 27 000 feet. The newly formed contrails (1) are sharply defined white streaks. With passage of time the trails broaden by diffusion and assume a fleecy aspect; in places they have developed a cellular structure (2) or a series of pendant columns (3).

PLATE XXVIII

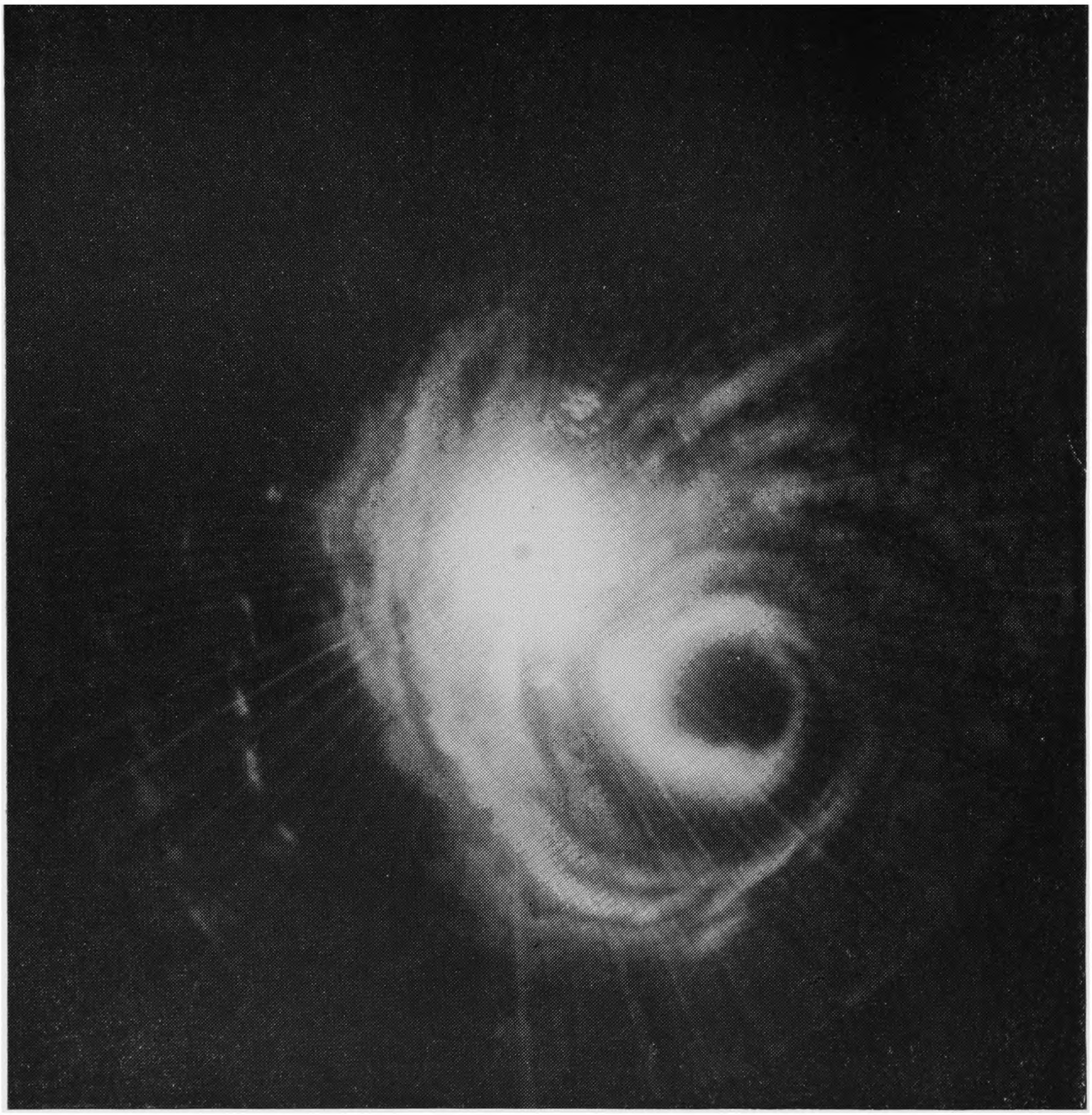
Summit of a giant storm cell
The storm cell (1) is about 5–6 miles in diameter at 60 000 feet. It protrudes some 10 000 feet above the surrounding turbulent cloud area (2) which slopes down in a rough flat cone to the main anvil sheet (3) near 40 000 feet—the height of the tropopause. Thus the giant storm rises about 20 000 feet into the stratosphere (see Section 42). (The photograph was taken at about 65 000 feet over mid-west U.S.A.)

PLATE XXIX



Satellite television picture centred over southern Scotland
This picture was received at Bracknell from the automatic picture transmission from ESSA II, orbit 835, height 1353 kilometres, at 0950 GMT on 5 May 1966 (see Section 96). The arrows indicate the assumed position of the core of a jet stream.

PLATE XXX



Reproduced by courtesy of U.S. Navy

Tropical revolving storm as viewed by radar (see Section 118).

PLATE XXXI

and with the aid of tables the reading is corrected to a temperature of 0°C.

(iii) *Correction for the local value of gravity.* Allowance is made for the variation of gravity with latitude by correcting to a standard value 980·665 centimetres per second per second. A correction for variation of gravity with height is also required but is negligible for heights up to 200 metres above mean sea level.

(iv) *Reduction to standard level.* When the previous corrections have been applied, the reading gives the pressure at the level of the barometer cistern. As explained in Section 7, a further adjustment is necessary to obtain the pressure at any other height, for example at aerodrome level or sea level, and a correction table is available for this purpose. In routine practice, in order to obtain the corrected pressure at mean sea level the four corrections are reduced to a single operation by means of a table which is specially prepared for each instrument in its given situation. Separate tables may be prepared to facilitate the determination of the pressure for each of the altimeter settings, QNH, QFE and QFF (Section 9).

Barometers on board ship. Instead of a correction table, a device known (after its inventor) as the 'Gold slide' is usually attached to the instrument; corrections for altitude, latitude and temperature are made by a simple setting of the slide. The conditions of observing at sea make the mercury barometer unreliable and difficult to use.

Aneroid barometer

An aneroid barometer usually consists of a metallic chamber exhausted of air and hermetically sealed. Variations of atmospheric pressure produce variations in the dimensions of the vacuum chamber and these variations are magnified mechanically, optically or electrically so that the atmospheric pressure may be read on a convenient scale. The majority of aneroid barometers indicate the pressure by means of a pointer which rotates around a graduated dial. The vacuum chamber, usually called the aneroid capsule, has to provide the force needed to move the pointer, and thus a faithful representation of the precise instantaneous atmospheric pressure is inhibited.

This type of instrument is useful in showing pressure changes, and some of the better quality instruments are suitable for all pressure readings. The Meteorological Office has adopted a type of precision aneroid barometer in which the force required to operate the indicating mechanism is provided by the observer, and thus the capsule is allowed to respond freely to pressure changes. The movements of the capsule are measured with a micrometer screw and the pressure is shown in millibars on a digital counter.

Corrections required.

(i) *Correction for index error.* A certificate is provided with each instrument, setting out the necessary corrections at different points on the scale.

(ii) *Corrections for temperature of the instrument.* The instruments are almost completely compensated for the temperature variations experienced in normal use, but if an instrument is used under extreme conditions the necessary corrections may be obtained from the test certificate.

(iii) *Correction for the local value of gravity.* Corrections for gravity variations are not required.

(iv) *Reduction to standard level.* When the previous corrections have been applied the reading gives the pressure at barometer level. As explained in Section 7 a further adjustment to the reading is necessary to obtain the pressure at any other level.

Barograph

In the most usual type of barograph the changes with pressure of the thickness of a stack of aneroid capsules is magnified by a series of levers, the last of which carries a pen. The pen draws a trace on a graduated chart wrapped round the outside of a cylinder rotated by clockwork. The chart with its trace is known as a barogram. The barograph is used mainly to indicate pressure changes, and is inferior to both the mercury barometer and a good aneroid barometer for measurements of pressure itself.

Barometric tendency and characteristic

Full synoptic reports include the amount and type of change in pressure over the three hours preceding the time of observation. The barogram is used in choosing the type of change (the characteristic) and may be used to determine the amount of change (the tendency) as described in the Meteorological Office publication *Handbook of weather messages, Part III.*

81. TEMPERATURE

The different scales used in the measurement of temperature have already been described in Section 10 and an account of the physics of surface temperature has been given in Section 12.

Various observations of temperature

The surface observations required for use in connection with aviation are the following:

(i) The air or 'dry-bulb' temperature.
(ii) The 'wet-bulb' temperature. In conjunction with the dry-bulb temperature, this is used for calculating the 'dew-point' temperature.
(iii) The maximum and minimum temperatures of the air over a given period.
(iv) The minimum night temperature near the ground, or 'grass minimum' temperature.
(v) Sea surface temperature is included in observations made at sea.

Types of thermometer

The usual type of thermometer is the mercury-in-glass pattern in which the temperature is indicated by the position of the end of the liquid column.

Maximum thermometer. The Meteorological Office pattern consists of a thermometer resting nearly horizontally in which the retraction of the mercury into the bulb on cooling is prevented by a constriction in the tube, as in the clinical thermometer. As a result the position of the far end of the mercury indicates the highest temperature reached since the instrument was last set.

Minimum thermometer. The thermometer used for recording the minimum temperature usually consists of alcohol in glass, with a small index inserted in the thread of spirit. The thermometer is supported horizontally and movement of the index occurs only when it is dragged towards the bulb by the meniscus (that is, the end of the column of spirit) as the temperature falls. On re-expansion with rise of temperature, the spirit flows past the index which remains stationary and so indicates the lowest temperature reached.

Grass minimum thermometer. This is a minimum thermometer used for measuring the lowest temperature reached at night very near the ground which, on clear nights, may cool by radiation well below the screen temperature. The thermometer is not screened but is placed in the open supported on two Y-shaped wooden pegs with its bulb just touching the tips of the blades of carefully trimmed grass. The reading is of most value in connection with the occurrence of ground frost.

Resetting maximum and minimum thermometers. Maximum and minimum thermometers must be reset after each reading. The maximum thermometer is reset by swinging it, bulb outwards, at arm's length until the mercury in the stem becomes continuous with the mercury in the bulb. For the minimum thermometer the instrument is tilted to allow the index to slide along the tube until it comes into contact with the end of the thread of spirit.

Wet-bulb thermometer. This is described in the next section.

Exposure of thermometers

In order to obtain representative readings the thermometers (dry-bulb, wet-bulb, maximum and minimum) must be exposed to the free passage of air and at the same time be protected from direct radiation by day from the sun, the ground, or neighbouring objects, from loss of heat by radiation at night, and from precipitation. The standard Meteorological Office screen, designed to serve this purpose, is a wooden cupboard with double-louvered sides to allow of free ventilation, mounted securely so that the thermometer bulbs are approximately 4 feet (1·25 metres) above ground level, and painted white to minimize the absorption of radiant heat. The door must be kept closed except for the short periods when observations are being made and, to avoid the penetration of direct sunlight when open, the door should face north in the northern hemisphere and south in the southern. Glass-fibre laminate louvered screens giving similar protection are now in use on a limited scale and provide a satisfactory alternative (see Plate I).

The screen should be placed in an open position well removed from sheltering buildings and over reasonably level ground. The height above the ground is important as there are sometimes considerable variations in temperature within the first few feet. On ships, precautions need to be taken to avoid unrepresentative readings caused by the passage of air over heated parts of the ship's structure.

Recording or distant-reading thermometers

The temperature-sensitive devices most frequently employed in self-recording instruments are the bimetallic strip and the mercury-filled Bourdon tube. The former is used in the thermograph mounted in the screen and the latter is used to record remotely indoors; and both are in the form of coils which tighten or slacken with variations of temperature. These movements are used to control a pen recording on a chart on a rotating drum.

Wire-wound resistance thermometers, now in use on an increasing scale, can be connected to a manually balanced resistance bridge for direct read-out, or to an automatically balancing chart-recorder.

Sea-surface temperature

The temperature of the sea surface is best obtained by drawing a sample of sea water from over the ship's side, forward of all discharge pipes, in a specially designed canvas bucket. The temperature of this sample is taken with a standard thermometer fitted with a reservoir for retaining a small quantity of sea water round the bulb should it be necessary to remove the thermometer from the bucket to read it. Ship observations of sea temperature are sometimes made by noting the temperature of the water entering the engine intake tubes. Sea temperature is sometimes measured in suitable open water adjacent to a land station, for example at the end of a pier or near a lighthouse.

Specially designed resistance thermometers have recently been mounted against the plates of a ship's hull, at the waterline, and the measurements of sea temperature were found to be in close agreement with the conventional readings.

82. HUMIDITY

The definitions of several measures of humidity have been given in Section 31. Numerous instruments, called hygrometers, have been designed for measuring humidity but most of them are suited only to laboratory work. The type usually adopted for meteorological purposes is the dry-bulb and wet-bulb hygrometer, alternatively known as the psychrometer. The wet-bulb thermometer is identical with the ordinary dry-bulb used for measuring air temperature, except that the bulb is covered with a single layer of thin muslin which is kept moist by means of a wick (a few strands of thick cotton) dipping into a small vessel of distilled water. When evaporation takes place from the muslin the latent heat required by the water vapour is taken mainly from the air surrounding the wet-bulb and from the water in the muslin. The wet-bulb temperature is therefore lower than the dry-bulb temperature by an amount that depends on the rate of evaporation and this in turn depends on the relative humidity of the surrounding air and on the ventilation around the muslin. Ventilation is reasonably standard inside the normal Meteorological Office screen but is fully standardized in a self-contained instrument called the Assmann psychrometer which requires no screen. In this instrument the dry-bulb and wet-bulb thermometers are protected from radiation by polished metal shields and a stream of air is drawn over the bulbs by a fan. There are known relations between dry-bulb and wet-bulb readings and the other humidity measures (such as dew-point, relative humidity, vapour pressure) can be obtained from tables or from a special slide-rule. It is important not to confuse the readings of the wet-bulb thermometer with the dew-point.

Some self-recording hygrometers and hygrographs employ a different principle, namely the variation with humidity of the length of a strand of human hair. The hygrograph records the relative humidity directly but is mainly used to show variations with time rather than to give exact values.

Neither the psychrometer nor the hair hygrograph is ideal for meteorological humidity measurements. The former will not give accurate results when the wet-bulb temperature is below 0°C and the latter must be checked frequently if reliable results are to be obtained.

In recent years automatic dew-point hygrometers which overcome both these difficulties have been developed for field use. These instruments control the temperature of a surface so that the vapour pressure of a deposit of water or ice on the surface is equal to the vapour pressure in the surrounding air. If the deposit on the surface is pure water or ice, the surface must usually be cooled below the air temperature to obtain the vapour pressure balance. However, some instruments employ a surface coated with a saturated solution of a hygroscopic salt, such as lithium chloride, which has a vapour pressure much lower than that of pure water at the same temperature. In these instruments the surface must normally be heated to obtain the vapour pressure balance. Both types of instrument indicate dew-point directly.

83. WIND

Wind is the motion of the air over the surface of the earth and its expression requires a statement of both speed and direction. The direction is that from which the wind is blowing, usually expressed in degrees from true north; thus 315° true means that the wind blows from north-west. For aviation purposes, the speed of the wind is usually given in knots but metres per second and kilometres per hour are sometimes used. To convert from one unit to another, it is useful to remember that one metre per second is approximately equivalent to two knots and that one kilometre per hour corresponds to half a knot.

Instrumental measurement of surface wind

As with other meteorological elements, the measured velocity and direction of the wind depend on the exposure of the instrument. Because of the marked variation of speed with height near the ground, the 'surface wind' is conventionally defined as the wind at a height of 33 feet (10 metres) above ground in an unobstructed situation. The more important types of measuring equipment are described briefly below.

Cup generator anemometer and wind vane. The anemometer consists of a small electrical generator maintained in a weather-proof housing and driven by the rotation in the wind of a three-cup rotor carried on a vertical spindle. The voltage generated increases with wind speed and is used to operate remotely-situated indicating dials graduated in knots.

The wind vane uses electrical transmission to indicate wind direction on dials graduated in degrees and cardinal points.

The cup generator anemometer is used in conjunction with the remote transmitting vane for the observation of surface wind at most aerodromes, dials for wind speed and direction being installed in parallel in the meteorological office and the air traffic control room.

The cup generator anemometer and remote transmitting wind vane are combined as the wind transmitting head of the electrical anemograph, Plate II, in which a continuous record of wind direction and speed, showing gusts and lulls, is made by means of two pens recording side by side on a moving duplex chart marked for wind speed and direction. A section of a chart, or anemogram, for Valley is reproduced in Fig. 19 (see also Chapter 20). In addition to the recorder, several speed and direction dials can be used in parallel with one transmitter.

Pressure-tube anemograph. This instrument, still in use at some stations, also gives a detailed record of the wind structure. It is based on the same principle as that of the airspeed indicator on an aircraft: the pressure produced by the wind blowing into the open end of a tube is interpreted in terms of wind speed. The tube is kept facing into wind by means of the wind vane and the pressure produced is transmitted along the tube to the recording apparatus. By fitting just below the vane a further tube pierced laterally with a number of small holes, a suction effect also is obtained from the wind and transmitted to the recording apparatus. In addition to contributing to the motion of the pen, which traces a record of wind speed on a rotating chart, the suction pipe connects the apparatus to the free air outside the building in which the recorder is housed, so rendering the record independent of any slight pressure differences between the inside and outside of the building. The rotation of the vane is transmitted by couplings to give a record of wind direction on the same chart as that used for speed. This instrument in its standard form has its recorder at the foot of the mast on which the vane is mounted and it does not read remotely. Also it has to be visited daily because the chart must be changed and it needs expert attention to keep it running in a satisfactory manner. For these reasons it has largely been replaced by the electrical anemograph.

Air meters and hand anemometers

These are portable hand-held instruments which give a local reading of wind speed at the place where they are held. The air meter consists of a small fan which rotates on a horizontal spindle connected by gearing to a number of counter dials; it must be held with the plane of rotation of the fan facing the direction from which the wind blows. It indicates 'run of wind', i.e. the amount of wind flow past the observer in a measured time, from which the mean wind speed can be deduced. With a sensitive air meter very low speeds can be measured, but the instrument is delicate and must be handled with care.

The hand anemometer is a small cup anemometer; the rotation of the cups, whose spindle carries a rotating permanent magnet, gives rise to eddy currents in a copper or aluminium disc or drum mounted on a spindle, and causes it to rotate in the same direction as the cups; its rotation is controlled by a hairspring. The angular movement of the disc or drum provides, by means of a pointer moving over a scale, a measure of instantaneous wind speed. Readings below 5 knots are unreliable.

Estimating wind force – Beaufort scale

Before the advent of aviation, those most concerned with wind were seamen. A set of descriptive terms of wind strength, evolved according to the effect on sailing craft and on the disturbance of the sea, came to be used with a fair degree of uniformity. In 1805 Captain (later Admiral) Beaufort devised a numerical scale ranging from 0 to 12 in which each number corresponded to one of the descriptive terms and was further defined by the effects of the wind on a 'well conditioned man-of-war'. As times changed the original specifications became of little value but the same scale was retained and is still known as the Beaufort scale. The modern seaman bases his estimate mainly on the effect of the wind on the sea and, although the precise effect is difficult to express, the observer gradually attains proficiency and a high standard of consistency is maintained. The same scale has been adapted for use on land according to the effect on smoke, trees and buildings, and further precision has been given by assigning to each number a definite range of speed. In conformity with the conventional definition of surface wind, the equivalent speeds given in Table 9 are taken to refer to observations made by an instrument exposed at about 33 feet above level ground in open country. Although most reporting stations are supplied with instruments for measuring wind, observers are encouraged to estimate the wind speed with the aid of the Beaufort scale and to compare the result with an instrumental reading. Skill acquired in this way is valuable when on occasions no anemometer is available.

TABLE 9. *Beaufort scale of wind force*

Force	Description*	Specifications for use on land	Equivalent speed at 33 ft (10 m) above ground			
			Knots		Metres per second	
			Mean	Limits	Mean	Limits
0	Calm	Calm; smoke rises vertically.	0	<1	0·0	0·0–0·2
1	Light air	Direction of wind shown by smoke drift, but not by wind vanes	2	1–3	0·8	0·3–1·5
2	Light breeze	Wind felt on face; leaves rustle; ordinary vane moved by wind.	5	4–6	2·4	1·6–3·3
3	Gentle breeze	Leaves and small twigs in constant motion; wind extends light flag.	9	7–10	4·3	3·4–5·4
4	Moderate breeze	Raises dust and loose paper; small branches are moved.	13	11–16	6·7	5·5–7·9
5	Fresh breeze	Small trees in leaf begin to sway; crested wavelets form on inland waters.	19	17–21	9·3	8·0–10·7
6	Strong breeze	Large branches in motion; whistling heard in telegraph wires; umbrellas used with difficulty.	24	22–27	12·3	10·8–13·8
7	Near gale	Whole trees in motion; inconvenience felt when walking against wind.	30	28–33	15·5	13·9–17·1
8	Gale	Breaks twigs off trees; gengenerally impedes progress.	37	34–40	18·9	17·2–20·7
9	Strong (severe) gale	Slight structural damage occurs (chimney pots and slates removed).	44	41–47	22·6	20·8–24·4
10	Storm	Seldom experienced inland; trees uprooted; considerable structural damage occurs.	52	48–55	26·4	24·5–28·4
11	Violent storm	Very rarely experienced; accompanied by widespread damage.	60	56–63	30·5	28·5–32·6
12	Hurricane	—	—	⩾64	—	⩾32·7

*In plain-language forecasts the word breeze is usually omitted except for forces 2 and 3. Forces 1, 2, 3 are sometimes grouped together and described as 'wind light'.

84. CLOUD

Classification of clouds

The study of clouds is particularly important for the purposes of aviation. In order that the character of the sky on any occasion may be fully appreciated and the weather map understood, familiarity with the names and appearances of the more important types of cloud is essential. The study of the causes of cloud formation and an account of the associated flying conditions has already been dealt with in earlier chapters.

Clouds are continuously evolving and present an unlimited variety of forms. In these circumstances it is necessary to confine the formal descriptions to certain frequent and characteristic forms, omitting many intermediate and transitional forms. The selected characteristic forms comprise 10 genera; each genus may be sub-divided into species and varieties. The definitions of the 10 genera are given below; it is most important to be familiar with these as they form the basis not only of the more detailed classification but also of the symbolic representation of cloud forms on synoptic charts.

Cloud genera

The following formal definitions of the 10 genera are based on the *International cloud atlas, Vol. I* .The recognized two-letter abbreviations are also given.

Cirrus (Ci). Detached clouds in the form of white, delicate filaments or white or mostly white patches or narrow bands. These clouds have a fibrous (hair-like) appearance, or a silky sheen, or both.

Cirrostratus (Cs). Transparent, whitish cloud veil of fibrous (hair-like) or smooth appearance, totally or partly covering the sky and generally producing halo phenomena.

Cirrocumulus (Cc.). Thin, white patch or layer of cloud without shadows, composed of very small elements in the form of grains, ripples, etc., merged or separate, and more or less regularly arranged; most of the small regularly arranged elements have an apparent width of less than 1°.*

Altocumulus (Ac.). White or grey or both white and grey sheet, layer or patch of cloud generally with shadows, and usually composed of laminae, rounded masses, rolls, etc., which may or may not be merged and are sometimes partly fibrous or diffuse; most of the regularly arranged small elements have an apparent width of between 1° and 5°.†

Altostratus (As). Greyish or bluish cloud sheet or layer of striated, fibrous or uniform appearance, totally or partly covering the sky, in parts thin enough to reveal the sun at least vaguely, as through ground glass. Altostratus does not show halo phenomena.

Nimbostratus (Ns). Grey cloud layer, often dark, the appearance of which is rendered diffuse by more or less continuously falling rain or snow which in most cases reaches the ground. It is thick enough throughout to blot out the sun. Low ragged clouds frequently occur below the layer, with which they may or may not merge.

Stratocumulus (Sc). Grey or whitish or both grey and whitish sheet, layer or patch of cloud that almost always has dark parts, non-fibrous (except for virga**),

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*Or the apparent width of the little finger, at arm's length.
†Or the apparent width of three fingers, at arm's length.
**Trailing precipitation which evaporates before reaching the ground.

and composed of tesselations, rounded masses, rolls, etc., which may or may not be merged; most of the regularly arranged small elements have an apparent width of more than 5°.

Stratus (St). Generally grey cloud layer with a rather uniform base which may give drizzle, ice prisms or snow grains; sometimes stratus appears in the form of ragged patches. When the sun is seen through the cloud, its outline is clearly discernible. Stratus does not produce halo phenomena except possibly at very low temperatures.

Cumulus (Cu). Detached clouds, generally dense with sharp outlines, developing vertically in the form of rising mounds, domes, or towers, of which the bulging upper part often resembles a cauliflower; the sunlit parts of these clouds are mostly brilliant white; their base is relatively dark and nearly horizontal. Sometimes cumulus are ragged.

Cumulonimbus (Cb). Heavy and dense cloud with a considerable vertical extent in the form of mountains or huge towers. At least part of its upper portion is usually smooth, fibrous or striated and nearly always flattened; this part often spreads out in the form of an anvil or vast plume. Under the base of this cloud, which is often very dark, there are frequently low ragged clouds (cumulus fractus or stratus fractus), either merged with it or not, and precipitation sometimes in the form of virga.

Altitude of the various cloud genera

The identification of clouds is aided by a knowledge of the approximate altitudes at which each genus normally occurs. By convention, that part of the atmosphere – the troposphere – in which clouds are usually found is divided into three levels: low, medium and high. The approximate altitudes are given in Table 10. The overlapping between the medium and high categories will be noticed. It should too be remembered that the limits given are not intended to be applied rigidly.

TABLE 10. *Approximate altitudes of cloud levels*

Level	Polar regions	Temperate regions	Tropical regions
High	10 000–25 000 ft (3–8 km)	16 000–45 000 ft (5–13 km)	20 000–60 000 ft (6–18 km)
Middle	6500–13 000 ft (2–4 km)	6500–23 000 ft (2–7 km)	6500–25 000 ft (2–8 km)
Low	From the earth's surface to 6500 ft (2 km)	From the earth's surface to 6500 ft (2 km)	From the earth's surface to 6500 ft (2 km)

It can now be indicated in which level (or levels) each cloud genus is most frequently found:

Cirrus, cirrostratus, cirrocumulus. High.

Altocumulus. Medium.

Altostratus. Medium, often extending to high.

Stratocumulus, stratus. Low.

Nimbostratus. Low, often extending to medium and high

Cumulus, cumulonimbus. Low, often extending to medium and high (especially cumulonimbus).

Cloud species

The clouds pertaining to a given genus are subdivided into species. The determination of the species is based on the form of the clouds, on their structure, and whenever possible on the physical processes involved in their formation. Any particular cloud can belong to only one species. These are described briefly below, and illustrations of some will be found in the cloud photographs (Plates III–XXVI); several of these Plates with their accompanying descriptions are included in the *International cloud atlas, Vol. II.*

Fibratus. Clouds in the form of filaments but without tufts or hooks. The term applies mainly to cirrus and cirrostratus.

Uncinus. Cirrus having the form of commas terminated at the top by either a hook or a tuft which is not rounded.

Spissatus. Cirrus of sufficient optical thickness to appear greyish when viewed towards the sun.

Nebulosus. A cloud like a nebulous veil, showing no distinct details. This term applies mainly to cirrostratus and stratus.

Stratiformis. Clouds spread out in a very extensive horizontal layer. This term applies to altocumulus and stratocumulus and occasionally to cirrocumulus.

Lenticularis. Clouds having the form of lenses or almonds, often very elongated, with sharp margins. These clouds are most often of orographic origin. The term applies mainly to cirrocumulus, altocumulus and stratocumulus.

Castellanus. Clouds which present cumuliform protuberances and so have a crenellated aspect. Such clouds are generally connected by a common base and seem to be arranged in lines. This term applies only to cirrus, cirrocumulus, altocumulus and stratocumulus.

Floccus. A species in which each cloud unit is a small tuft with a cumuliform appearance, the lower part of which is more or less ragged and often accompanied by virga. This term applies to cirrus, cirrocumulus, and altocumulus.

Fractus. Clouds of which the units have a ragged appearance. This term applies only to stratus and cumulus.

Humilis. Cumulus, the cloud units of which present only a small vertical development and generally appear somewhat flattened.

Mediocris. Cumulus clouds of moderate vertical development, the tops being accompanied by fairly small protuberances.

Congestus. Cumulus, the cloud units of which are strongly sprouting. The cloud often has a great vertical development and its bulging form frequently resembles a cauliflower.

Calvus. Cumulonimbus in which no cirriform part can be distinguished but in which at least some of the protuberances are beginning to change from a cumuliform to a fibrous structure.

Capillatus. Cumulonimbus with distinct cirriform parts, frequently having the form of an anvil. Cumulonimbus capillatus is usually accompanied by a shower or a thunderstorm, often with squalls and sometimes with hail; it generally produces very well-defined virga.

Cloud varieties

A cloud of a given genus or species may present certain characteristics. The particular aspect of a single one of these characteristics will determine a 'variety' of this cloud. The characteristics most often considered for the determination of varieties are on one hand the transparency of the cloud and on the other the arrangement of the larger cloud elements. The same cloud may possibly be attributed to more than one variety. Nine varieties have been defined but it is unnecessary to reproduce the definitions here. Some examples are given in the descriptions accompanying the cloud photographs.

High-level clouds

The cloud forms which are described below are additional to those of the preceding classification.

Nacreous clouds. The form of these clouds resembles that of cirrus or altocumulus lenticularis. They show very strong irisation, that is, tinted patches. Nacreous clouds are not the only ones to show irisation; it occurs also in altocumulus and in cirriform cloud. Observations of nacreous clouds are infrequent, and are mainly from Scotland and Scandinavia. Photometric measurements indicate an altitude of about 20–30 kilometres, that is, in the ozone layer, but the physical constitution of the clouds is not definitely known.

Noctilucent clouds. These clouds resemble cirrus but they usually have a bluish or silvery colour although sometimes orange to red. They are visible only at night when they are directly illuminated by sunlight and where the sky is dark enough for their weak luminescence to be perceptible; good visibility and absence of intervening ordinary cloud are also necesssry. They are observed when the sun is between 5° and 16° below the horizon, i.e. soon after sunset and just before sunrise, and so could be visible all night in summer in northern regions. The height of the cloud is between 80 and 85 kilometres.

Noctilucent clouds are most frequently observed around 55°N and during June and July, although they have been seen as far north as 72° and as far south as 45° and in late May and up to mid-August. In Scotland most of the observations are of cloud below an elevation of 10° and to the north but on rare occasions it appears to cover the whole sky.

At one time it was assumed that the clouds were composed of dust particles but as a result of rocket soundings made in Sweden in 1962 it is now thought that the clouds consist of ice particles.

Abnormally produced clouds

This class includes, among others, condensation trails (see Plate XXVIII and Section 70); clouds from fires; clouds due to volcanic eruptions; clouds resulting from industry and explosions.

Cloud amount

The amount of cloud is given by the number of oktas (or eighths) of the sky covered by the cloud. The amount of any low cloud present is always reported together with the total amount of cloud. The number of oktas of sky covered by any particular cloud layer is estimated as if no other clouds were present; this can normally be done satisfactorily during the day-time but at night it is more difficult, especially when there is more than one layer.

For the purposes of aviation, information is required on the height and amount of all layers of cloud likely to be of operational importance. The selection of layers to be reported is made in accordance with the following requirements:

(i) The lowest individual layer of any amount.
(ii) The next higher individual layer the amount of which is three oktas or more.
(iii) The next higher individual layer the amount of which is five oktas more.

Pilots should realize that reports of cloud amount are usually made from a fixed position on the ground. Should a pilot fly within or near a cloud layer he perhaps may not encounter gaps visible from below and might report having flown through continuous cloud when actually it was broken. Alternatively a patch of very low cloud which entirely obscures the sky from an observer on the ground might be seen from above to be merely a local formation obscuring only a fraction of the ground below. Cloud amounts as observed from the air do not necessarily tally with those observed from the ground.

Height of cloud base

Some methods of estimating and measuring cloud heights are described in the following paragraphs. Observations of cloud height from aircraft are described in Section 95.

Visual estimation. With experience, an observer becomes reasonably proficient in estimating the height of the base of clouds. This ability is developed partly by comparing the estimates with heights obtained by direct measurement, as described below, and partly with the aid of inferences from the appearance of the cloud, its type and structure and the general weather conditions. The height of cloud in relation to neighbouring hills or possibly objects such as aerial masts can sometimes be noted, although it should be remembered that the height of the cloud overhead may differ appreciably from the height at a distant point. When the cloud consists of an extensive uniform sheet, an accurate eye estimate of the height is usually difficult; in such cases an instrumental determination should be made whenever possible. Aircrew sometimes can check their own estimates against an altimeter reading at the cloud base, although care should be taken to avoid errors due to lag.

Cloud-base recorder. This is now the primary instrument for measuring the height of the cloud base at airfields. The equipment, which operates by day as well as by night, is fully automatic and capable of unattended continuous operation. The complete system consists of three units, a transmitter, a receiver, and a recording unit.

The method by which the equipment determines the height of the cloud base involves measuring the angle of elevation of a light beam, scanning in the vertical plane, at the instant at which a proportion of the light scattered by the cloud is received by a vertically pointing photo-electric cell at a known distance from the light source; this distance is 350 feet. The transmitter emits a nearly parallel beam of two degrees divergence. The beam is swept in a vertical arc and is modulated to ensure that the receiver is sensitive only to light emitted by the transmitter. The receiver unit comprises a photo-electric cell and an angle-of-view restrictor to ensure that light reaches the photocell only from vertically above. A pen in the recording unit, moving in sympathy with the intersection of the transmitter beam and the 'cone of acceptance' of the receiver, writes when a cloud signal is received. A current is passed through the pen tip so burning a trace on the chart. Reflection takes place from within the cloud as well as from the base, as can be seen from Fig. 75. Since the recorder is primarily designed as an aid to aviation, the maximum height measured is 4000 feet and the scale on the chart is so arranged that the lowest 1000 feet is expanded to cover half the chart and the remaining 3000 feet occupies the other half. Part of the record for London/Heathrow Airport on 18 January 1967 is shown in Fig. 75.

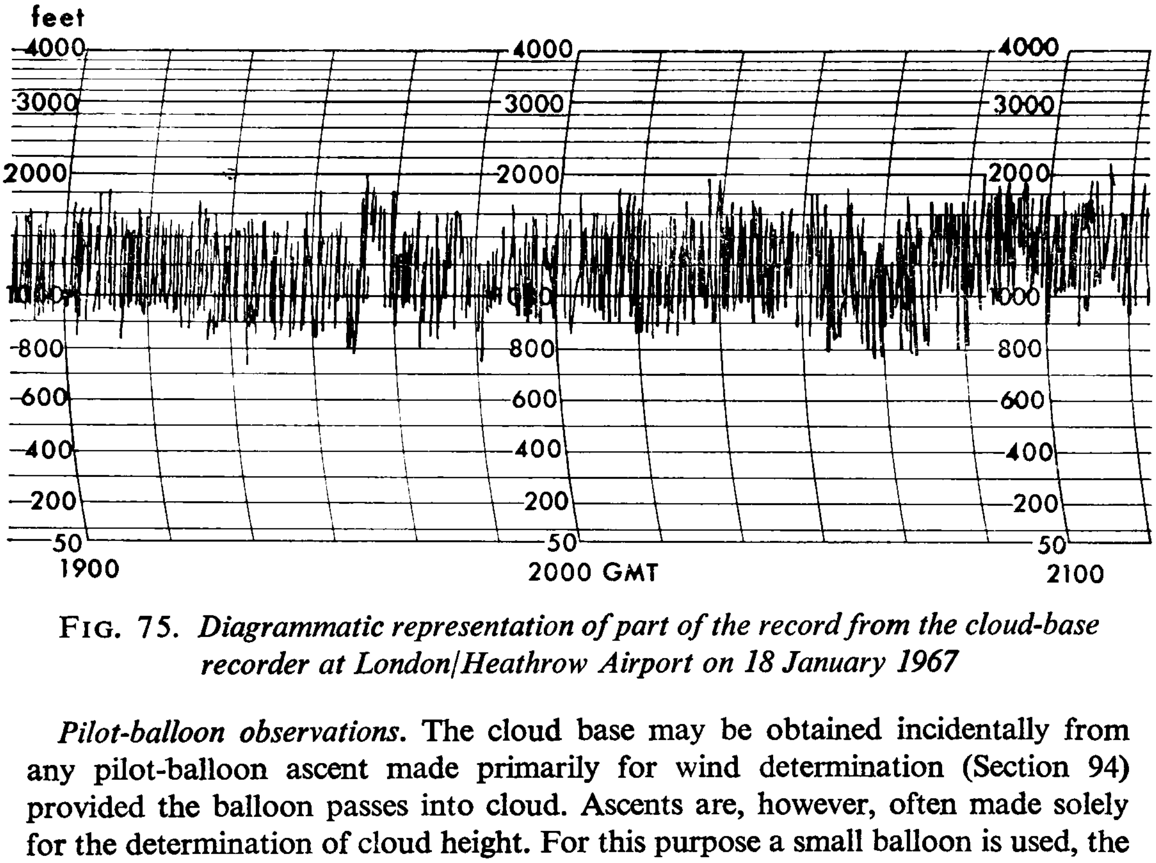


FIG. 75. *Diagrammatic representation of part of the record from the cloud-base recorder at London/Heathrow Airport on 18 January 1967*

Pilot-balloon observations. The cloud base may be obtained incidentally from any pilot-balloon ascent made primarily for wind determination (Section 94) provided the balloon passes into cloud. Ascents are, however, often made solely for the determination of cloud height. For this purpose a small balloon is used, the rate of ascent being 400 or 500 feet per minute. A stop-watch gives the time between the release of the balloon and its entry into the cloud, whence the height of the base is found. The method is particularly useful for low cloud layers. It may also be used at night with the aid of a small lantern suspended from the balloon.

Cloud-searchlight observations. The cloud searchlights installed at most airfields provide a rapid and accurate method of determining cloud heights at night. A beam of light is projected vertically on to the base of the cloud and the angular elevation of the spot of light so formed is observed by the aid of a sighting device known as an alidade. This observation, together with the known distance of the alidade from the searchlight, suffices to determine the cloud height. The searchlight is switched on whenever required and the observation performed within a minute or so.

Accuracy of measurements of cloud height

The height of the base of low cloud may vary considerably both in space and time. The topography of the ground affects the cloud base which usually becomes lower over hills. Even over comparatively level ground, variations in the degree of turbulence and changes in the intensity of any precipitation falling out of it will produce variations in cloud base. At a given place the cloud base often fluctuates rapidly with time about a mean height which itself may also be changing. The Meteorological Office practice is to attempt to measure the lowest point of the fluctuation.

Cloud thickness

The vertical thickness of a layer of cloud is a matter upon which a pilot would often welcome information, but there is no ready means of determining this from the ground. Information may be available from in-flight reports made by pilots on operations, while the forecaster is also helped in his assessment of this element by radiosonde data (Section 94). The height attained by convection cloud can sometimes be estimated visually, while radar viewing of the clouds may also give some indication of vertical development. A further source of information is the cloud-base recorder, described above, which gives some idea of the density of the cloud and may indicate the top of the lowest layer in multi-layered cloud.

85. VISIBILITY

The relationship between the meteorological visibility and one's ability to see any particular object in given circumstances depends on several factors which have been discussed in Chapter 9. Among others, the visibility between air and ground and the measurement of runway visual range, or the visibility along an airfield runway in conditions simulating those experienced by a pilot about to land an aircraft, have been described. The present section is concerned only with observations of the horizontal visibility at ground level; in any weather report or forecast the term 'visibility', when unqualified, has this restricted meaning. In practice it may be interpreted as the greatest horizontal distance at which a person of normal sight can recognize prominent objects under normal conditions of daylight illumination.

Determination of visibility by day

To aid the estimation of visibility, it is necessary to select a number of suitable objects located approximately at certain standard distances from the point of observation, ranging from 20 metres to the visible horizon. The visibility distance is then easily determined by reference to the furthest objects visible and may be expressed as so many metres or kilometres or in other required units. In addition, or alternatively, to the actual distance, descriptive terms are also employed. Full details of the standard visibility ranges and descriptive terms will be found in the *Observer's handbook.* As examples, we may note that 'fog' is used to describe visibility less than 1000 metres, the obscurity being due to the suspension in the air of very small water droplets, while mist describes a visibility of 1000 metres or more when the relative humidity is between 95 and 100 per cent. The term haze, on the other hand, is used to describe the suspension in the air of extremely small dust particles when humidity is less than 95 per cent.

It is frequently impossible to select a set of standard visibility objects for all the required distances; at sea it is never possible. In such circumstances the observer must make an estimate based on such objects as are visible and on the apparent clearness of the atmosphere.

Determination of visibility by night

In daylight the visibility as defined above is determined almost entirely by the transparency of the atmosphere and at night it is desirable that the same property should be made the basis of the estimate of visibility. The instrument in use for this purpose is known as the Gold visibility meter. It consists of a simple visual photometer which measures the apparent brightness of a distant light of known intensity and hence the transparency of the intervening air. The light is observed through a variable filter until it is only just visible, the setting of the slide is noted and the reading converted to the equivalent daylight visibility by means of a diagram. A separate calibration needs to be carried out for each observer using the slide, in order to eliminate errors caused by variations in the sensitivity of the eyes between one observer and another. Before using the slide, the observer should allow about two minutes for his eyes to become adapted to the darkness.

When a visibility meter is not available, the visibility is judged by observation of suitable fixed lights at known distances, but for good results it is necessary to know the candle-power of these lights. Tables are available showing the relation between daylight visibility and the distance at which white lights of known candle-power can just be seen.

In circumstances when neither of these two aids nor a transmissometer (see below) is available, the visibility can be estimated only from the general clearness of the night; when the visibility permits, it is surprising how much of the landscape can be seen when the eye becomes accustomed to the darkness even on a really dark night.

Transmissometer. This instrument is used to record the optical transparency of the air by measuring the decrease in brightness of a horizontal searchlight beam after it has travelled a certain distance, say 200 metres, through the atmosphere. A photo-cell receiver is used to detect the light. The air is sampled several times each minute and the transparency is recorded on a chart. Not only can minute-to-minute variations in optical transparency be seen, but gradual changes are brought to the attention of the observer. In the majority of meteorological conditions there is a close relation between optical transparency and meteorological optical range. This is a quantity which is identical for practical purposes with visibility and is defined as the length of path in the atmosphere required to reduce the luminous flux in a collimated beam of light to 0·05 of its original value. The chart is calibrated in terms of meteorological optical range (MOR).

86. PRECIPITATION

Most of the terms used in describing precipitation are sufficiently well known to require no explanation. Although there are many conventions relating to matters of detail which have been laid down for the sake of uniformity, it is necessary here to draw attention to only a few points which are common causes of confusion.

Drizzle and rain

The distinction between these two elements is based not on the quantity of precipitation but on the size of the droplets. Drizzle consists of droplets which are so small that they make no perceptible impact on water surfaces. Drizzle frequently occurs in association with mist or fog but sometimes produces low visibility in otherwise clear air. Continuous drizzle may produce a run-off from roofs and road surfaces. Rain, on the other hand, consists of drops of appreciable size. It is not possible to define a particular size of drop at which drizzle ends and rain begins. If individual droplets make a distinct splash on striking the ground or a water surface, they should be recorded as rain, even though few in number.

Showers and intermittent rain

Showers, whether of rain, snow, sleet or hail, are always associated with convection clouds. After a shower there is a tendency for breaks to appear in the cloud, even if the sky does not entirely clear. If precipitation falls from time to time from extensive clouds with no marked clearances, it is reported as intermittent. The distinction is important if the report is to be given its full significance, for both the causes of the two types of precipitation and also the accompanying flying conditions are essentially different.

The various forms of frozen precipitation have been defined in Section 38.

87. RADAR DETECTION OF CLOUD AND PRECIPITATION

The location of distant objects by radar depends on the emission of a suitable short-wave radio pulse from a transmitter, and the reception of the return pulse after reflection from the object. The time interval between the emitted and received pulses provides a measure of the distance of this object, while its bearing is related to the direction of the maximum reflected signal. Radars which have been developed specifically for the detection of precipitation are now in use throughout the world.

The intensity of the radar response from a cloud of n drops each of diameter d is proportional to nd^6, so that one drop of diameter 2 millimetres can give an echo equivalent to that from a million drops of diameter 0·2 millimetre. Although any cloud consists of drops of various sizes, the echo from a cloud containing raindrops is far greater than that from other clouds, so much so that it is true to say that precipitation, and not cloud, is detected by the normal rain-detecting radar apparatus, which uses 10- and 3-centimetre wavelength transmissions.

The maximum range of any radar depends on, amongst other things, the power transmitted and the sensitivity of the receiver. In practice the range of a weather radar under normal propagating conditions cannot exceed 240 nautical miles. Radar waves follow almost straight paths, and though refracted seldom curve as much as the earth's surface. Because of this the radar beam at 0° elevation cannot 'see' below 40 000 feet when the range reaches 240 nautical miles. However, during conditions of anomalous propagation the range of the radar beam may be increased (see Section 77).

Two forms of radar display are in use. The more common is the PPI (plan position indicator) which shows the azimuth and the range from the station of the received echoes. A transparent overlay showing the local geography may be used on the screen to illustrate the location of the rain more readily. Normally the range scale of the display can be selected from several alternatives, for example 30, 60, 120 and 240 nautical miles.

The other form of display is the RHI (range height indicator) which gives a useful cross-section of the precipitation in a chosen area and enables the observer to gauge its height and to see the shape of the leading edge. By means of adjustments which can be made to the radar receiver it is possible for an experienced operator to distinguish between different types of precipitation and to form an assessment of its intensity. Echoes are received from hail and from snowflakes, and near the melting level a 'bright band' of intense echoes is often observed; this is believed to be due to large flakes starting to melt and being covered with a film of water, thus presenting a large echoing area.

88. ATMOSPHERICS

It is well known that atmospherics originating from lightning discharges interfere with radio reception. When a lightning flash occurs, radio waves are emitted over a wide frequency band, the most intense waves having frequencies less than 50 kilocycles per second (wavelength greater than 6000 metres). This was realised by Watson-Watt in the years following 1920 and led him to build the first thunderstorm cathode-ray direction finder. In this apparatus the bearing of the discharge is displayed visually by means of the cathode-ray tube, and by taking simultaneous observations on the same lightning flash from widely separated receivers, the source of the flash is accurately located.

The British so-called 'sferic' network consists of four stations in the British Isles and stations at Gibraltar, Malta and Cyprus. As atmospherics often occur in quick succession and each lasts for only a small fraction of a second, a special method of synchronization is required to ensure that the virtually simultaneous recordings made at the seven stations actually refer to the same flash. The bearings are transmitted to a control station where they are transcribed on to a chart and the source of the atmospheric is thus determined. Observations are made between 10 minutes to the hour and the hour for every hour of each day; additional observations are made between 20 and 30 minutes past each hour for a limited period each day. During these periods a random selection of thunderstorms is made by using a device which ensures a quiescent period of five seconds between each selected flash.

The sferic network maintains a watch on the distribution and intensity of thunderstorms occurring within about 1500 miles of the British Isles.

89. WEATHER

In such terms as 'weather report' or 'weather map', the word 'weather' is taken to cover all the meteorological factors concerned, but the word is also used in a restricted sense, when it refers only to the state of the sky and the accompanying precipitation or other phenomena – thunderstorms, fog and so forth. A simple system of notation, devised by Admiral Beaufort and known as the Beaufort letter notation, permits the observed weather to be recorded concisely by means of letters; in general the initial letter of the observed phenomenon is used. Certain elaborations of the original notation have been adapted to meet modern requirements. The intensity of a phenomenon is indicated by capital letters for unusual intensity and by small letters for moderate intensity, while slight intensity is indicated by a small letter with suffix $_0$. Thus:

S	heavy snow
s	moderate snow
s_0	slight snow

A phenomenon which is continuous is noted by repeating the letter, while one of an intermittent character is indicated by a prefix i. Thus:

RR	continuous heavy rain
ir_0	intermittent slight rain

When reporting observations it is usual to give both 'present weather', referring to conditions at the time of observation, and 'past weather', the conditions during the interval since the previous routine observation. The observer must therefore keep a close watch on weather changes between the routine times of observation.

90. SUNSHINE

The duration of sunshine is measured by a recorder in which a glass sphere is used as a lens to burn a trace on a card whenever the sun is shining. During a day of continuous sunshine the burn traces out a continuous line on the card as the position of the sun alters. When sunshine is intermittent the line is broken, the combined length of all the parts giving a measure of the duration of sunshine.

91. STATE OF GROUND

This is reported by choosing the most suitable of ten standard descriptions (see the *Observer's handbook*). Some of the descriptions – dry, moist, wet or frozen – refer to a selected or prepared place of bare ground (not grass-covered) which should be representative of soil in the vicinity. The remaining descriptions refer to the general condition of open ground easily visible from the station and within 100 feet (30 metres) of it in altitude.

92. STATE OF SEA

Ships at sea and certain coastal stations give information regarding the condition of the sea. These reports include the height of the waves, the direction from which they are coming, and the time interval between successive waves.

93. AUTOMATIC WEATHER STATIONS

In recent years a few automatic weather stations have been introduced into the observing network. As time goes on more will certainly come into use. They enable observations of most of the 'surface' parameters to be obtained from locations where it is not convenient, or a practicable proposition, to maintain observing staff. The observations so taken are transmitted on demand or according to a predetermined time schedule and received at a remotely situated manned office. Such stations may also be equipped to record continuously all the observed parameters. In this way climatological data are obtained. The recording continues irrespective of the transmission of real time information for synoptic use. Land-lines are the preferred medium of transmission but where such facilities do not exist radio links may be used.

CHAPTER 13

UPPER AIR OBSERVATIONS

94. BALLOON OBSERVATIONS

Observations of the meteorological conditions in the free air are of obvious importance for flying operations and they also form an essential part of the meteorologist's basic data, the variation of wind, temperature, etc., in the upper levels being just as important for weather analysis and forecasting as the variations from place to place at the surface. Upper air observations can be made in many ways, but attention here will be confined mainly to routine methods and of these one of the most valuable is that which depends on balloons to carry instrumental equipment aloft.

Radiosonde observations

The standard method of obtaining values of temperature and humidity at various pressure levels is by use of a balloon carrying a small radio transmitter known as a radiosonde. The carrier wave of this transmitter is modulated by an audio-frequency oscillator, into the circuit of which are switched in turn three variable inductances controlled by the meteorological elements, pressure, temperature and humidity.

The latest type of radiosonde to be used by the United Kingdom is transistorized and is both compact and sensitive. Pressure is measured by an aneroid capsule, temperature by a resistance thermometer unit of fine tungsten wire, and humidity by a piece of gold-beater's skin. The temperature and humidity elements are exposed but the remainder of the instrument is enclosed in an insulated container. A small electric motor turns the switch which couples the meteorological elements to the transmitter circuit. A small battery supplies power for all these purposes.

The heights attained by radiosonde balloons vary from about 65 000 feet (20 kilometres) for small balloons to 115 000 feet (35 kilometres) or more for large balloons.

Equipment is required at the ground station to measure and record the fluctuations of the audio-frequency of the radiosonde. The British equipment consists of a radio receiver and electronic instruments which measure and record the time elapsed for 100 cycles of any frequency to pass through the equipment. These measurements of periodicity are directly related to frequency. In consequence, when the receiver is tuned to the radiosonde transmission, the record produced by the ground equipment shows variations of pressure, temperature and humidity as the balloon ascends but not of course in the conventional units. Before use, therefore, the radiosonde requires calibrating. This is done by artificially subjecting the instrument to known changes of pressure, temperature and humidity, and noting the corresponding variations in audio-frequency. Graphs are drawn of these changes. Measurements of audio-frequency from the airborne radiosonde can then be referred to the calibration graph and converted to the conventional values of the elements being observed.

During flight the meteorological elements are switched into the circuit with sufficient rapidity to produce what appears to be an almost continuous record, especially in the case of temperature. From this record, points of significant change of temperature lapse and hydrolapse* are selected for computation. Computed values of temperature and humidity are plotted against pressure on a suitable diagram and these graphs are used for selecting values for the synoptic message.

Radar winds

As a free balloon ascends, it is carried along horizontally by the wind at its level so that a series of observations of its position enable the average wind to be determined over the layers traversed between successive observations. The tracking of the balloon is usually carried out by radar. British stations use primary radar which depends on the reflection of radio waves emitted by a ground station and for this purpose a reflector, made of metallized nylon mesh, is suspended from the balloon. The radar method enables the bearing, distance and height of the balloon to be determined at each observation, usually at intervals of one minute, regardless of weather. Some nations use secondary radar which depends upon a radio transmitter carried by the balloon. Bearings are taken from the ground on the transmissions from the balloon.

The present British wind-finding radar sets enable wind velocities to be measured up to the limits of the balloon's performance.

Radar-wind ascents may be carried out independently or in combination with temperature and humidity observations, one balloon carrying both the radar reflector and the radiosonde apparatus.

Pilot-balloon observation of wind

Before the introduction of radar methods, upper winds were commonly determined by observations of a 'pilot' balloon. This small rubber balloon is filled with sufficient hydrogen to cause it to rise approximately at a known rate, usually 500 feet per minute. During the ascent, observations of elevation and azimuth are made at frequent intervals by means of a theodolite (essentially a surveyor's telescope) and the wind direction and speed are evaluated by means of a special slide-rule. This method is still quite widely used, but it has serious disadvantages; thus its use is severely restricted by low visibility or the presence of cloud, while strong winds carry the balloon out of sight before it has had time to rise more than a few thousand feet. However, given good visibility and winds not too strong, a height of 10 000 feet is generally attainable, while with light winds 30 000 feet is not unusual provided clouds do not interfere. Variations in the pre-arranged rate of ascent of the balloon may occur for several reasons, but a check can be obtained by measuring in the eyepiece of the theodolite the apparent length of a 'tail' suspended from the balloon. A more accurate method uses simultaneous observation of the balloon by two theodolites at the ends of a measured base line, but this is involved and unsuited to routine use.

Ascents may be made at night by observing a lantern attached to the balloon. The use of pilot balloons to determine cloud height has already been noted (Section 84).

95. AIRCRAFT OBSERVATIONS

Many different kinds of meteorological instrument have been designed for use in aircraft. Some are required to meet the needs of research or are suited only to aircraft carrying a meteorological observer while others form part of the standard equipment of civil or military aircraft. Apart from visual observations, the brief

*Change of humidity with height.

descriptions which follow are mostly confined to the more familiar types of instrumental observation and aim only at illustrating the principles of measuring the properties of the atmosphere from an aircraft in flight.

Pressure

This is measured by an aneroid barometer. The instrument is provided with a zero adjustment so that before take-off the aneroid may be set to read the pressure at the height of the instrument above the airfield, in agreement with that obtained from a corrected mercury-barometer reading. Owing to hysteresis or lag, the pressure indicated in flight will be greater than the true pressure when the aircraft is climbing and less when descending. The error depends on the rate of change of height and may amount to several millibars. In order to obtain a reliable reading, level flight is needed for at least a minute, but if the change of height has been rapid a longer time should be allowed.

The aneroid barometer is basically the same instrument as the pressure altimeter and is subject to the same instrumental errors; these have been described in Section 8.

Temperature

Several types of aircraft thermometer are in use. All are provided with an anti-radiation housing, protection from solar radiation having been found necessary even with the high ventilation obtained with aircraft. The exposure of the thermometer element must be in a position unaffected by engine heat or aircraft slip-stream, or by solar heating of adjacent parts of the airframe.

Spirit-in-glass thermometers. In this type, dry-bulb and wet-bulb thermometers are usually mounted side by side, the combined instrument being known as an aircraft psychrometer. It is fitted to a strut or bracket near the observer's window and is a standard instrument for meteorological reconnaissance aircraft. Some patterns can be read visually at a distance of a few feet, but another, which has the advantage of very small lag, needs to be within about one foot of the observer.

Distant-reading thermometers. With this form a capillary tube leads from the bulb to the Bourdon tube (see Section 81) and dial which may be mounted in a convenient position in the cockpit. The fluid consists of either mercury or an inert gas. While the former is the more accurate, its range is limited since mercury freezes at —38°C; it also has a large lag. The thermometer bulb, with its longitudinal axis in line of flight, is bolted to the under-side of the aircraft wing or fuselage.

Electrical resistance thermometers. The 'bulb' of these instruments is an electrical resistance which varies with the temperature. There are various patterns of bulb, one being the Meteorological Office 'flat plate' which is exposed flush with the surface of the aircraft. The thermometer element should in any case be mounted on the under-side of the aircraft nose or wing depending on the engine position. Changes in the resistance are measured with the aid of a Wheatstone bridge and galvanometer and the corresponding temperature may be indicated on a dial in the cockpit.

Thermometer lag. When exposed in an environment of different temperature, an aircraft thermometer does not immediately take up that temperature but is subject to a lag which varies with the temperature difference, the type of thermometer and size of bulb, the ventilation (and hence airspeed of aircraft), the air density (and hence altitude), and the rate of climb or descent. The types with the smallest lag are the electrical resistance thermometers and one pattern of the spirit-in-glass thermometer. In order to obtain an accurate reading, the aircraft should be flown at the required level at constant airspeed for about one to two minutes depending on the type of thermometer, airspeed and altitude.

Correction of indicated temperature for airspeed. The phenomenon of kinetic heating at the surface of an aircraft in flight was explained in Section 47 where it was seen that in clear air the temperature is increased by the effects of both adiabatic compression and air friction. The result is that the indicated temperature is too high by an amount given by

$$\Delta T = \alpha \left(\frac{v}{100} \right)^2$$

where v is the true airspeed and α is the speed-correction coefficient. The precise value of α depends on the type of instrument and mounting as well as on the units employed; for temperature in degrees Celsius and true airspeed in knots, its value is in the neighbourhood of unity. Thus the excess temperature due to kinetic heating begins to become appreciable at a true airspeed of about 100 knots; in high-speed aircraft it causes large errors in the indicated temperature.

Effect of cloud and rain on indicated temperature. When passing through a water-drop cloud the thermometer bulb may become wet; it then acts as a wet bulb. When the temperature is greater than 0°C and the airspeed is less than 150 to 300 knots, the precise limit depending among other things on the water content of the cloud, little or no error would result from the wetting of the thermometer since the vapour must be practically saturated. At temperatures less than 0°C, if the cloud drops freeze on striking the bulb, the release of latent heat of fusion tends to maintain the bulb at a temperature very near freezing-point although some types of thermometer are less subject to this effect than are other types.

At greater airspeeds, kinetic heating becomes sufficient to cause partial or complete evaporation of the cloud particles as they approach the dry bulb. The temperature excess at the dry bulb due to kinetic heating will consequently be less than that in clear air at the same airspeed, but no simple rule can be given. In clouds consisting entirely of ice crystals, there is little effect on the speed correction since the ice content itself is small.

Similar considerations apply to the measurement of air temperature when the aircraft is flying through rain. In some cases the humidity may be well below the saturation value, for example in rain beneath a shower cloud; the indicated temperature is then likely to be too low, unless it is counteracted by the effects of kinetic heating at high airspeeds.

When an aircraft emerges from cloud or rain with the thermometer bulb or housing wet, there is an effect analogous to lag. The instrument takes time to dry and until then a temperature below the true temperature will be observed, apart from the effects of kinetic heating. The observer should therefore wait until the temperature becomes steady before he takes a reading.

Considerations such as these indicate the difficulty of obtaining the true ambient temperature when flight is in cloud or rain. These conditions should accordingly be avoided as far as possible when representative observations are required. On the other hand the indicated temperature in cloud or precipitation is a useful guide to the likelihood of ice accretion although it should be remembered that the temperature is not the same at all points of the surface of the aircraft.

Humidity

Similar difficulties apply to the measurement of humidity as to the measurement of air temperature. If cloud is present at the level of observation but the observation is taken in clear air, then the recorded relative humidity will be lower than that representative of the level although the wet-bulb temperature and absolute humidity will not necessarily be unrepresentative. It is, however, difficult to make a direct measurement of the humidity of the air inside a cloud since the cloud particles are likely to affect the hygrometer. It is better to assume that the cloudy air is saturated with respect to water or ice according to the nature of the cloud. If, as the result of a corrected observation in cloud, the air appears to be appreciably supersaturated, it is probable that cloud particles have been evaporated, for example by adiabatic heating, in their approach to the hygrometer. Similar difficulties are experienced in attempting to measure the humidity of air through which rain is falling.

In Section 31 it was seen that the saturation vapour pressure over ice is less than that over supercooled water at the same temperature. It follows that the relative humidity of a given sample of air is necessarily greater with respect to ice than it is with respect to water. This difference becomes progressively greater with decrease in temperature. The practice in the Meteorological Office is to refer the relative humidity at temperatures below 0°C to a plane surface of water, but the wet-bulb temperature is observed when the bulb is covered with ice.

Aircraft psychrometer. In this instrument dry-bulb and wet-bulb thermometers of the spirit-in-glass type are mounted side by side. This method has rather serious drawbacks, some arising from the excessive ventilation of the wet bulb, others from the freezing of water on the wet bulb when the temperature falls below 0°C and the subsequent disappearance of the ice by evaporation.

Gold-beater's skin and hair hygrometers. The use of human hair for measurement of humidity at the earth's surface was mentioned in Section 82. Gold-beater's skin reacts in a similar way to changes in humidity, and both materials have been adapted to the making of observations in aircraft. In either case the indicated relative humidity refers to a water surface whatever the temperature.

Dew-point or frost-point hygrometer. This instrument gives accurate measurements of the dew-point or frost-point at all temperatures down to —80°C. Air is drawn into the aircraft from a forward-facing scoop at a point unaffected by engine exhaust and is passed over a cooled metal surface. The temperature of this surface is controlled by spraying a suitable cooling agent to its under-side by means of a hand pump and the temperature is noted at which a deposit of dew or hoar-frost begins to form; subsequently the surface is allowed to warm and the temperature noted at which the deposit begins to evaporate. The dew-point or frost-point is then the mean of the two temperatures.

Lag of hygrometers. The lag of the wet-bulb thermometer is generally less than that of the dry bulb, so that the steady level flight necessary to obtain the dry-bulb temperature with satisfactory accuracy will suffice also for the wet-bulb temperature. An exception occurs when the water covering the wet bulb is in process of freezing, for then the release of latent heat tends to maintain the wet-bulb temperature at 0°C. During this period it is possible for the wet bulb to read higher than the dry bulb. If the dry-bulb temperature is only slightly below 0°C and the relative humidity is high, the time required to complete the freezing of the wet bulb, and so to obtain a reliable reading, may be five minutes or more.

Correction of hygrometer readings for airspeed. The main effect of airspeed on hygrometer readings arises from the increase of temperature due to kinetic heating; since this heating does not change the amount of water vapour associated with a given mass of clear air, the ultimate result is that the indicated values of relative humidity are usually too low. As with the dry-bulb temperature the correction in-increases with the square of the airspeed so that observations should in general be made at the minimum airspeed consistent with steady level flight and in clear air where possible. When observations are made in cloud, saturation should be reported as already mentioned.

Wind

There are several methods of determining the wind speed and direction from an aircraft in flight and all depend fundamentally on the triangle of velocities. Thus if an aircraft is flying on a heading and at an airspeed which in still air would take it from A to B in time t (Fig. 76), then B is its 'air position' at that time and AB

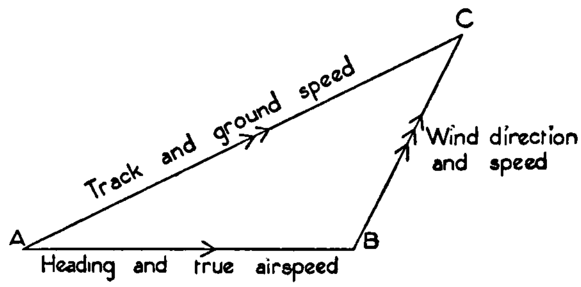


FIG. 76. *Vector triangle for determination of wind from an aircraft*

represents the heading and airspeed. If, however, the aircraft arrives not at B but at C, then BC represents the mean wind speed and direction during the time t while AC is the track made good. The success of methods of wind measurement clearly depend, among other factors, on the accuracy with which the aircraft's position can be determined, whether visually or in some other way. Details of methods of wind measurement will be found in books on navigation and will not be considered here.

Clouds as seen from the air

The classification of clouds seen above the aircraft present no special difficulty to an observer who is already familiar with their observation from the ground and, similarly, detached clouds can be readily classified whatever the viewpoint. Continuous cloud sheets however present characteristic upper surfaces which can be viewed only from above, and special attention should be paid to these. Several photographs of clouds viewed from above are included in Plates III to XXIX, while some further remarks on this aspect are given below.

Cirrus. Very fine cirrus is often invisible from above, and the ground or lower clouds are seldom obscured by it. If of great vertical thickness it appears opaque, with a milky aspect.

Cirrostratus. This cloud has a milky aspect when viewed from above. The ground or lower cloud is usually visible if the cloud is thin, but hardly perceptible when thick.

If the flight is in cirrus or cirrostratus, the ice crystals may at times be seen glittering in the sunlight. On the other hand, it is possible to fly through these clouds without being aware of them.

Cirrocumulus. Viewed from above, it is seen as small or discontinuous sheets, the upper surface of which generally appears fleecy.

Altocumulus. The elements of this cloud may be separate or merged. The upper surface presents a smooth undulating layer, or is slightly fleecy; in some cases it is continuous, in others there are cracks or distinct gaps which permit glimpses of the ground or of lower clouds. *Altocumulus castellanus* from above resembles well-developed cumulus emerging from a smooth undulating layer. The tops may also be seen to emerge from a haze layer. *Altocumulus floccus* appears like small cumulus clouds surrounded by, or emerging from, a milky white area.

Altostratus. The nearly horizontal upper surface usually appears smooth, flat or undulating with a fairly long wavelength. Tops of cumuliform clouds may be seen projecting above the cloud layer.

Nimbostratus. Often the upper part of this cloud merges into medium clouds. In other cases the upper surface has a somewhat flat but diffuse aspect, unless it becomes unstable and takes on a cumuliform appearance.

Stratocumulus. As with altocumulus, the cloud units may be separate or merged in a continuous sheet. In the former case, the upper surface looks like a 'sea of cloud' with a fleecy appearance; more or less marked open spaces or cracks appear, and the surface is frequently pierced by heads of cumulus or cumulonimbus. When the cloud is continuous, the upper surface often has undulations in the form of long parallel bands.

Stratus. The upper surface is generally undulated with a rather short wavelength, but sometimes the undulations are more pronounced and follow the irregularities of the ground. Hills often protrude above the cloud top.

Fog. The upper surface of fog, which it is convenient to refer to here, is like a 'sea of cloud', flat and smooth, but sometimes shows slight undulations. Surface obstacles as well as hills may protrude above the top.

Aircraft observations of cloud height

Height of cloud can be relatively accurately determined by observations made in aircraft at cloud level, provided the altimeter is used properly to obtain the true height (see Section 8). Valuable information may be obtained from in-flight observations made by pilots of aircraft, particularly in regard to the height of clouds hidden from a ground observer by other cloud. Certain difficulties should be noted: some high-level clouds are so tenuous that an aircraft observer cannot tell whether he is in cloud or not; sometimes the cloud base is ill defined; but an aircraft observer can occasionally see a thin haze layer or cloud layer against the background of the ground even though a ground observer cannot distinguish the layer against the sky.

Radar detection of cloud and precipitation

The use of ground-based radar for this purpose was described in Section 87. An apparatus suitable for use in aircraft has also been devised, the main object being to assist the pilot in avoiding regions of abnormal turbulence such as occur in large cumulus or cumulonimbus, especially if such clouds cannot be seen visually as when flying in darkness or in layer cloud. It will be remembered that radar echoes are obtained only from clouds containing drops of raindrop size and that these are the clouds in which turbulence is most likely to be severe. Use of the apparatus as an aid to location of icing conditions is described in Section 51.

Icing and turbulence

The formation of ice on aircraft is the subject of Chapter 8, and turbulence has been discussed in Chapters 5 and 10. Observations of the type and intensity of airframe icing and of the intensity and frequency of turbulence may be made according to certain codes (see *Handbook of weather messages, Part II*).

96. SATELLITE OBSERVATIONS

The Meteorological Office has been receiving pictures from weather satellites since 1964, following the introduction of the automatic picture transmission (APT) facility on TIROS VIII. Routine operational use did not become possible, however, until March 1966 shortly after the launch of ESSA II, which was in a near-polar orbit. Present practice ensures that pictures are available from the current ESSA satellite and additionally from the NIMBUS satellite when orbital times do not overlap. Some 15 to 20 pictures are received daily, each covering a square of side approximately 1500 nautical miles. There is an overlap of approximately 30 per cent between successive pictures in the same orbit and pictures from adjacent orbits also overlap, the amount varying with latitudinal distance from the equator.

After reception, latitude and longitude lines are determined for each picture; this requires a set of pre-computed grids and a knowledge of the orbital data and the exact time at which the picture was taken. The pre-computed grids are projected on a screen to provide a magnification appropriate to the altitude of the satellite. The accuracy of 'gridding' is to about $\frac{1}{2}°$ under good conditions.

Pictures when gridded are ready for interpretation by the meteorologist. The fundamental principle of interpretation is found in the constant confrontation of features in the photograph with data from other sources. In doing this the limitations of resolution must be borne in mind. At the centre of the picture, resolution is about $1\frac{1}{2}$ nautical miles, but it decreases towards the edges of the picture to about 4 nautical miles. Cloud pictures must be interpreted in the light of all available information from other sources. Only in this way can experience be obtained and subsequently be safely applied in areas where surface observations are completely lacking. Experienced interpretation of satellite pictures can provide more information than just about the main cloud systems which are evident from a cursory glance. The six most important characteristics which help to identify clouds are brightness, pattern, structure, texture, shape and size. An examination of these features makes it possible to identify types of cloud (with some limitations however), to detect cloud layers of different heights, and to make inferences about wind structure and atmospheric stability.

Large-scale cloud patterns and cloud distributions are an aid in identifying positions of jet streams, fronts, ridges and troughs. The stage of development of extratropical depressions and tropical cyclones can be inferred from cloud pictures. The effects of mountains and islands on cloud distribution can be seen and when conditions are favourable the existence of lee waves may be inferred. The pictures can be used also for routine mapping of ice and snow, and for determining the extent of fog and stratus.

An example of a satellite picture is given in Plate XXX and the contribution satellite meteorology can make to forecasting is indicated in Section 142.

CHAPTER 14

COLLECTION AND CHARTING OF OBSERVATIONS

97. INTERNATIONAL CO-OPERATION AND OBSERVATIONAL NETWORKS

The rapid collection of a large number of synchronous observations from stations scattered over a wide area is an essential requirement for any meteorological service. The speed of modern aircraft and the long distances travelled necessitate the production of weather maps covering large areas and hemispherical charts are a usual feature in central forecasting offices. The exchange of this volume of information between countries requires close international co-operation and this is achieved through the World Meteorological Organization (WMO). This organization was incorporated as a specialized agency of the United Nations Organization in 1951, and took the place of the International Meteorological Organization, which had been formed in 1878.

In the specialized field of aeronautical meteorology, the standard procedures necessary for the safe, economic and regular operation of international air services are formulated by the Meteorological Division of the International Civil Aviation Organization (ICAO). This body works in close collaboration with WMO in order to ensure common procedures and to avoid unnecessary duplication of services. In general, ICAO accepts the recommendations of WMO concerning general meteorological questions. These organizations have standardized the meteorological procedures used in making and disseminating observations so that weather reports and forecasts can be exchanged internationally in a coded form, without the complication of language differences. Amendments to codes and procedures are agreed and brought into operation to meet changing conditions.

A network of meteorological observing stations has been established which provides a collection of weather reports required internationally for the preparation of weather maps. The density of the network is influenced by difficulties of terrain and climate, by large areas of ocean and by the differing degrees of organization.

Land stations

Main reporting stations make full observations of the weather elements at three-hourly intervals throughout the day and night, and a high proportion make hourly reports. Additional stations make abbreviated observations for limited periods of the day, or in some cases at hourly intervals throughout the 24 hours.

Each meteorological observing station throughout the world is allocated a station index number of five digits. The first two figures, called the block number, represent the country or zone and the last three the serial number of the station in the national list of the country concerned. A world-wide list of station index numbers is given in the WMO *Publication No.* 9, Volume A. Also the *Handbook of weather messages, Part I* contains a list of British stations together with an abridged list of certain foreign stations.

Ship stations

Reports from most ocean areas of the world are obtained through the voluntary co-operation of ships' officers. Regular observations are made on selected ships and the reports are transmitted to the appropriate shore radio station designated by international agreement to accept such messages.

Ocean weather ships. A network of ocean stations, established by international agreement through the medium of ICAO, operates continuously. The ships are maintained by interested maritime states. They make regular surface and upper air observations, provide radio navigational aids for transoceanic flights, and form part of the search and rescue organization.

Weather reports from aircraft

Meteorological information from aircraft in flight or from aircrews after flight is used to supplement the synoptic weather data. In areas where the network of observations is sparse, meteorological reports from transport aircraft are particularly valuable. The reports are transmitted to meteorological collecting centres from which they are rebroadcast to other meteorological offices.

98. CODING OF SURFACE OBSERVATIONS

The internationally standardized form of weather message which has been adopted is one of figure codes usually arranged in groups of five figures. A detailed description of the codes employed can be found in the *Handbook of weather messages, Part II,* in *Instructions for the preparation of weather maps,* or in WMO publications. The following brief explanation of the code form which is used for surface weather reports from land stations illustrates the general principles on which all meteorological figure codes are based.

The symbolic form of the SYNOP code consists basically of six groups to which supplementary groups (shown in parentheses) may be added.

SYNOP (II)iii Nddff VVwwW PPPTT $N_hC_LhC_MC_H$ $T_dT_d$app (8 $N_sCh_sh_s$)

1st group	II	block number
	iii	station number
2nd group	N	fraction of sky covered by cloud, in oktas (eights)
	dd	wind direction on scale 01–36 (36 = N, 00 = calm)
	ff	wind speed in knots
3rd group	VV	horizontal visibility
	ww	present weather
	W	past weather
4th group	PPP	last three figures of pressure at mean sea level, in tenths of a millibar
	TT	air temperature, in whole degrees Celsius
5th group	N_h	fraction of sky covered by all C_L cloud(s) present, or if no C_L is present, all C_M cloud(s)
	C_L	clouds of type Sc, St, Cu, Cb
	h	height above ground of base of lowest cloud
	C_M	clouds of type Ac, As, Ns
	C_H	clouds of type Ci, Cs, Cc
6th group	T_dT_d	dew-point, in whole degrees Celsius

	a	characteristic of barometric tendency during the preceding three hours
	pp	barometric tendency during the preceding three hours, in tenths of a millibar
7th group		supplementary group (not reproduced)
8th group	8	group indicator figure
	N_s	amount in oktas (eighths) of the individual cloud layer indicated by C
	C	type of cloud
	h_sh_s	height above ground of the base of the cloud layer indicated by C, in hundreds of feet

The following is an example of a message and its decode:

SYNOP 03772 60110 66022 09410 32602 03001 83830 86075

II = 03	British Isles		T_dT_d = 03	03°C	
iii = 772	London/Heathrow Airport		a = 0	increasing then decreasing; pressure same as or higher than 3 hours ago	
N = 6	6 oktas		pp = 01	0·1 millibars	
dd = 01	010°		N_s = 3	3 oktas	
ff = 10	10 knots		C = 8	cumulus	
VV = 66	16 kilometres		h_sh_s = 30	3000 feet	
ww = 02	sky unchanged		N_s = 6	6 oktas	
W = 2	cloud covering more than half of sky during period		C = 0	cirrus	
PPP = 094	1009·4 millibars		h_sh_s = 75	25000 feet	
TT = 10	10°C				
N_h = 3	3 oktas				
C_L = 2	cumulus				
h = 6	3000–5000 feet				
C_M = 0	no medium cloud				
C_H = 2	dense cirrus				

Synoptic reports from ships are coded in the SHIP form which is similar to the SYNOP form but with additional groups giving the position of the ship, time of the report, movement of the ship and sea conditions.

An abbreviated form of the SYNOP, called SYRED, consists of the wind, weather and cloud groups; supplementary groups, plain-language groups and Q signals may be added.

Aviation routine weather reports are coded in METAR form (see Section 151).

99. TRANSMISSION OF OBSERVATIONS

Observations from a country or region are collected regularly, in many cases every hour, by a collecting centre for the area; the 'collectives' are exchanged between countries which then retransmit the exchanged information to stations in their own region.

The existing system for routine international exchange of meteorological information is based on a vast network of land-line, cable, radio and facsimile transmission channels.

Observations in the British Isles are collected by regional centres and sent by teleprinter to the Meteorological Communication Centre at Bracknell; they are then retransmitted as a collective message to home stations and to national centres overseas within about 40 minutes of the time of origin.

Data from Greenland, Iceland, North Atlantic merchant ships and ocean weather stations are collected and distributed by Bracknell. Exchange of information between

Bracknell and other European sub-centres takes place by land-line teleprinter. Information is exchanged between Europe and America by radio and cable. Facsimile transmissions by land-line and radio allow British and overseas stations to receive diagrams prepared by the Central Forecasting Office, Bracknell, which itself receives good pictures of charts produced in Washington, Moscow and Tokyo. The organization of communications is continually changing and details of current procedures can be found in the latest edition of the *Handbook of weather messages, Part I* and in the meteorological section of the *U.K. Air Pilot.*

100. PREPARATION OF SURFACE CHARTS

In the previous paragraphs information has been given regarding the coding and transmission of weather reports. These reports are received in large numbers at forecasting offices and before they can be comprehensively viewed by the forecaster they require to be plotted on a suitable geographical chart. On the working charts used for this purpose the position of each reporting station is marked by a small circle with its three-figure station number alongside, while the block numbers are shown for each area. The coded message is represented by entries in and around the appropriate station circle, some in figures, some in symbols, but in a standard form which for the most part is agreed internationally in order that charts may be interpreted with equal facility by persons of any nationality. On the few points on which international agreement has not yet been reached, the British practice is adhered to in this book.

An outline of the method of plotting is given below in order to help the reader in the interpretation of charts such as those which appear in Chapter 20, but for full details reference should be made to the *Handbook of weather messages, Part III* or to *Instructions for the preparation of weather maps.* It is not expected that all pilots and navigators will seek to become expert at plotting a weather chart but they should be sufficiently versed in the methods used to permit of ready interpretation of the completed charts. Proficiency in this respect is very desirable at briefing and is required of candidates presenting themselves for meteorological examinations; it can perhaps best be attained by actual practice in the plotting of charts from coded weather reports. Unless the candidate is attached to a recognized school of instruction, it is advisable for him to obtain access to a meteorological forecasting office in order to see the work in progress. Specimen synoptic messages may usually be obtained for practice purposes and the completed charts may subsequently be compared with those prepared officially.

Plotting land station reports

The technique of plotting is based on the station model shown in Fig. 77 where the relative positions of the various elements are shown by the corresponding letter symbols, the meanings of which have been given in Section 98. An example of a plotted report from a land station is given in Fig. 77 (*b*) and the interpretation is further explained by the following remarks:

Colour. Red or black is used in conformity with the station model except that the barometric tendency (pp) and characteristic (a) are entered in red if the tendency is negative. At many subsidiary offices the charts used are facsimile copies transmitted from the main parent office. On these charts the plotting is in one colour only. Further technical advances envisage automatic plotting direct from the received tapes of coded weather messages and these charts too will be in one colour only.

Figure values derived from the coded message are used for the following elements: T_dT_d, VV, PPP, TT, N or N_s, h or h_sh_s, and pp (the figures for N_s and h_sh_s, when supplied, are plotted instead of those for N_h and h).

Symbols are used for the remaining elements.

Wind. The wind direction is shown by the shaft of an arrow flying with the wind. The speed is indicated by feathers and solid pennants: a full-length feather represents 10 knots (5 metres per second), a half feather 5 knots, and a solid pennant 50 knots. A calm is represented by a circle surrounding the station circle.

(*a*) (*b*)

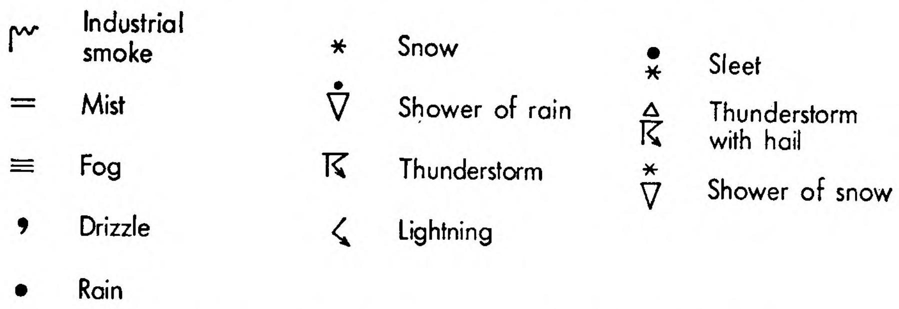
FIG. 77. *Method of plotting a report from a land station*

(*a*) Plotting model

(*b*) Plotted report from:
SYNOP. 72120 48516 93211 37326
10518 83706 85559

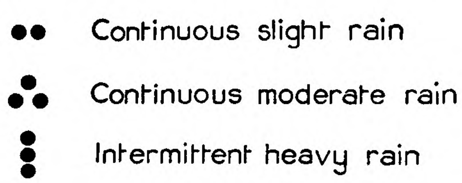
Total cloud amount (N) symbols are plotted within the station circle and range from a clear circle when there is no cloud to a blacked in circle when the sky is completely covered.

Weather (present weather ww, past weather W). The main symbols are as follows:



Note: sleet is the British term for a mixture of rain, or drizzle, and snow, as given in the ww code.

Two or more of these symbols may be combined. A symbol repeated horizontally indicates continuity of the phenomenon, while one repeated vertically indicates greater intensity. For example:



Cloud. The cloud symbols will not be given here in any detail but it is helpful to note the pictorial connection with the form of cloud represented. Thus a horizontal line always represents a flat or stratiform characteristic, while a curve represents a cumuliform or undulatory characteristic. There is no difficulty in recognizing to which height interval any given symbol refers, since the low cloud is invariably plotted below the station circle, the medium cloud above in black, and the high cloud above in red.

Plotting ship reports

Ship reports are plotted by a method very similar to that used for land stations. The latitude and longitude of the ship must first be identified from the message and a 'station' circle drawn on the chart in the appropriate position. Other data are entered as for land stations, with the addition of:

(i) the sea temperature which is entered in black immediately below the dew-point

(ii) the course and speed of the ship, which are plotted below the entry of past weather by a black arrow pointing in the direction of movement together with the code figure for the ship's speed

(iii) wave information, plotted by means of a wavy red line and arrow-head in the direction of movement of the waves, with code figures for the period and height of the waves.

Completing the surface chart

The plotting of charts is a mechanical process which may be learnt without any consideration of meteorological science, but to pass from the plotted chart to the completed weather map it is necessary to have a good understanding of both physical and synoptic problems. Although the full consideration of the technique of chart analysis is necessarily deferred to a later chapter, many students will benefit by acquiring some experience in charting at an early stage in their training. It is recommended that after some familiarity with plotting has been obtained, the student should attempt the drawing of isobars on his first charts, regarding them to start with simply as lines of equal pressure. If he has studied Chapters 2 and 5 he will be familiar with their general properties including their relationship to surface wind speed and direction, and care must be taken to draw the lines in conformity with those properties. For full instructions the student should refer to Section 129, most of which he should be able to assimilate at this stage with little difficulty.

Preparation of upper air charts

Consideration of these is deferred to Chapter 19.

PART III

SYNOPTIC METEOROLOGY

CHAPTER 15

AIR MASSES AND FRONTS

101. INTRODUCTION

In Part I the physical principles underlying meteorological phenomena have been discussed and the individual factors considered independently; it now remains to adopt the point of view of the practical forecaster faced with a succession of synoptic charts and other observational data from which he has to infer probable changes in meteorological conditions over an area or route, involving an analysis of the various factors and an assessment of their combined effect in any particular situation.

Once it became possible to draw weather maps it was soon evident that in most parts of the world large-scale variations in weather conditions are associated with changes in the type of pressure field. The first step towards forecasting came with the recognition of the importance of the formation and movement of depressions and anticyclones. The next step forward was the discovery by Norwegian meteorologists of the nature of frontal zones and the development of a theory of the formation of depressions associated with a polar front. This led to a classification of air masses with fronts as boundaries between them and the frontal cloud masses being formed by the upsliding of warmer lighter air over colder denser air.

Further increase in knowledge of the conditions in the upper air and observations of finer detail in frontal zones by research aircraft have led to the realization that fronts are more complex than the idealized picture suggests. It is now considered possible that depressions may form dynamically, causing convergence of the isotherms into a frontal zone, rather than that the depressions form because of the existence of the frontal zone.

Whatever the cause of the frontal depression may turn out to be, the frontal zones are very real. They may differ in detail from the idealized picture and considerably from one to another, but they have been, and still are, an exceedingly useful feature in the analysis of weather situations portrayed by synoptic charts.

102. AIR MASSES

The properties of the air depend primarily on temperature and humidity; the study of synoptic charts shows that these factors are often broadly the same over wide areas measuring many hundreds or even thousands of miles across. When air of substantially the same characteristics covers a large area it is called an 'air mass'. It is to be noted that, since there is little horizontal difference in temperature through the mass, there is little variation of wind with height above the friction layer (see Section 26), so that the air moves as an almost solid current.

In order that a mass of air shall assume uniform characteristics it needs to remain tolerably stagnant for a period of many days in a region where the earth's surface is itself reasonably uniform. The subtropical belts of high pressure, the polar anticyclones associated with the arctic fields of snow and ice, and the large highs which form over the continents in winter are areas of this kind. Whilst air remains over these regions it becomes more or less homogeneous and when under the influence of wind it eventually moves away to another area it largely retains its source characteristics although these become slowly modified.

Classification and characteristics of air masses

The basis of the classification of air masses is the source region last occupied by the air mass. Since there is perhaps no part of the world which might not on occasion act as a source region, the different labels given to the variety of air masses affecting any particular place are largely a matter of the geography of the surrounding areas and especially of the distribution of land and sea. Here we shall in the main confine ourselves to a general classification determined primarily by the latitude of the source regions; thus in order of increasing latitude we shall speak of equatorial air, tropical air, polar air and arctic (or antarctic) air. The factor most affected by the latitude of the source is the temperature. As the differences in humidity between two air masses of the same basic type depend essentially on whether they have originated over land or over sea, a secondary classification is made into continental and maritime air masses.

The properties of the air as it arrives at any place depend on the characteristics acquired while at the source and on the changes which have subsequently taken place during transit. These changes are due mainly to surface heating or cooling and to evaporation or condensation; their effects have already been studied in Part I. When a polar current penetrates to lower latitudes it undergoes surface heating; this tends to cause convection currents, a steep lapse rate of temperature and perhaps cumulus or cumulonimbus clouds with instability showers. Surface heating moreover raises the temperature further above the dew-point and, unless moisture is freely provided by evaporation, the air near the surface becomes relatively dry. Conversely, when a tropical or equatorial current moves to higher latitudes it undergoes surface cooling. Surface cooling inhibits convection currents, the cooling is confined near the surface and the lapse rate becomes stable often with an inversion above the surface layers. Cooling also tends towards condensation in the form of dew, mist, fog, low cloud, or drizzle, although if the humidity is too low, condensation may not occur and the sky will remain clear.

On the basis of these well understood principles it is a simple matter to infer in broad terms the probable characteristics of any air mass on arrival in another part of the world, account being taken of the differences in temperature between land and sea according to the season of the year.

Air masses of temperate latitudes

In temperate latitudes of the northern hemisphere, the polar and tropical air masses with their subdivisions into maritime and continental varieties are the usual types in all seasons, while arctic air is also experienced occasionally in winter. The source regions and typical properties of these air masses and the modifications they undergo in their passage from the source to the temperate latitudes will now be described. A similar basic terminology applies also in the southern hemisphere; the air masses there will not be described in detail but incidental references to them are made in discussing the climatology of these regions in Part V.

Tropical maritime air. The source regions of this air mass are the subtropical anticyclones in the Atlantic and Pacific Oceans. The air accordingly starts with a high temperature, high relative humidity and high dew-point, and it largely retains these characteristics as it moves towards neighbouring regions. In the North

Atlantic as the tropical air moves north-eastwards towards Europe it is continuously cooled from below and if the wind is light the lower layers become increasingly stable, perhaps with an inversion; with stronger wind, the lapse rate is steepened within the frictional layer and the base of the inversion is lifted. The high relative humidity acquired in the source region is maintained or intensified by the subsequent surface cooling; saturation is often reached and low stratus or stratocumulus, fog, drizzle and orographic cloud form readily. Over the sea the air temperature is characteristically higher than that of the sea itself, and over land in winter the air is mild. In summer, as it passes over land, surface heating by day usually disperses any stratiform cloud or fog, but the pronounced stability usually prevents the development of large convective currents even if there is orographic or frontal lifting, and visibility is usually moderate or poor.

Different conditions are found on the southern side of the North Atlantic anticyclone where the air is moving westwards towards the United States. This movement brings the air over a still warmer sea giving a marked tendency towards instability so that convection initiated by further surface heating over land or by orographic or frontal lifting often gives rise to cumulonimbus with showers and thunderstorms in summer.

Tropical continental air. The source region in winter is north Africa. In summer the source occupies a vast area extending across north Africa and southern Europe to eastern Asia; another source region in summer is the arid region of North America to the west of the Mississippi. Tropical continental air is warm at its source, with low relative and absolute humidity. In moving to higher latitudes the lower layers become cooled, and over the sea the air temperature is higher than the sea temperature. Humidity tends to remain low particularly when the track is over dry land, but when it passes over water, evaporation increases the humidity, both relative and absolute; if the process continues long enough, the air mass becomes changed into a maritime type. Generally, however, the dryness of the air and the effect of surface cooling in stabilizing the lapse rate, combine to prevent cloud formation even if the air is subsequently subjected to intense surface heating. This air mass affects the British Isles only in summer; after crossing Europe, high temperatures are maintained by insolation and hot cloudless weather results. Since there is nothing to remove the fine dust which enters the air mass in its source region or in its later passage over land, the visibility is usually hazy.

Polar maritime air. The source regions of this air mass are the northern parts of the North Atlantic and North Pacific Oceans. At the source, the air is characterized by low temperature, low dew-point, high relative humidity and stable lapse rate, at least in the lower layers. In moving towards warmer regions, the air mass is heated from below but its surface temperature remains less than that of the sea surface. Thermal instability spreads upwards from the lower layers and the lapse rate becomes steep up to high levels. This permits development of the associated cloud formations and often leads to large cumulus or cumulonimbus. Instability showers are associated with the well-developed clouds and at times they are accompanied by thunderstorms, strong gusts or squalls. On occasions the lapse rate aloft is stabilized by a general subsidence of the air mass and cloud development is then restricted to fair-weather cumulus or possibly stratocumulus. Except in showers, visibility is usually good since, at its source, the air is usually clear and any impurities subsequently carried into the air are dispersed by the convection currents or removed by the showers.

If this type of air mass moves from sea to land in summer, the advective surface heating is accentuated over land by insolation and convection is at first intensified. However, after crossing the coast the absorption of water vapour by evaporation at the surface is greatly reduced and the increasing dryness of the air makes it more difficult for convection cloud to form. In winter, polar maritime air is relatively mild as it moves over land cooled by radiation. The air mass then undergoes surface cooling and becomes stabilized in the lower layers, thus taking on some of the characteristics of tropical maritime air; showers decrease in intensity, clouds tend to spread out into layer types, while fog can form readily under a clear sky when the pressure gradient is slack. A somewhat similar situation occurs when the air mass, after moving to lower latitudes, again moves northwards as 'returning polar maritime air'. The surface layers then become stabilized and resemble tropical maritime air, but the upper layers are unstable, so that once convection is started, as for example over land in summer, it is apt to develop freely.

Polar continental and arctic air. The source regions of these air masses are the northern parts of the Eurasian and North American continents and are more extensive in winter than in summer. The arctic regions, situated between these two areas, constitute the source region for arctic air. Since at the source the polar continental and arctic air masses are in contact with a surface mostly covered with snow or ice, the air and dew-point temperatures are low; the lapse rate is stable in the lower layers and perhaps also aloft because of subsidence. Except over the continents in winter, these air masses arrive in temperate latitudes only after travel over a warmer surface with absorption of heat and possibly moisture, so that the lapse rate becomes gradually steeper from the surface upwards.

In winter over the land the convective developments are absent on account of weak insolation and low humidity; clear skies are then the rule. If the air subsequently moves over the sea, evaporation and heating produced by the relatively warm sea surface result in cumuliform cloud with perhaps wintry showers. In fact, the air mass in this process takes on a transitional character and becomes transformed more or less completely into polar maritime air. Thus polar continental air from North America in winter is transformed into polar maritime air as it travels eastwards across the Atlantic Ocean. Another example, in which the process of transformation is not carried so far, is provided by polar continental air which reaches the British Isles in winter after crossing the North Sea; showers commonly occur near the east coast but on moving further inland the source of heat is removed and cloud and showers die out.

In summer the polar continental air, while moving to a warmer part of the land, is at first dry and cloudless. It is therefore subject to surface heating by insolation; some of this heat is transferred upwards by convection, and thus in time the air becomes converted into a warm mass, sometimes called 'transitional' polar continental air. If this mass then moves over the sea, which is cool compared with the heated land, it is stabilized and moistened in the surface layers so that fog or low stratiform cloud may form. This is experienced for example on the east coast of England and Scotland provided the track across the North Sea is long enough to produce a high humidity in the originally dry air. Genuine polar continental air does not reach this country in summer, since the air necessarily undergoes transformation into the polar maritime variety if it travels over the sea, or into warm continental air if it travels overland.

Such in brief are the typical features of the principal air masses in temperate latitudes. In practice no two air masses of the same type are exactly alike, nor is any one air mass strictly homogeneous. The character of an air mass undergoes more or less continuous transition by such processes as loss or gain of heat, changes in humidity and in the vertical motion to which it may be subjected during its history. The effect of subsidence in stabilizing the lapse rate has already been mentioned; this process also dries the air and dissipates cloud; if the air undergoes ascent then it becomes cooled, the relative humidity increases, the lapse rate may steepen and clouds form easily. The association of slow ascent or descent with the pressure distribution and the barometric tendency have been noted in Section 27 and it will now be appreciated how the character of an air mass becomes modified when it moves into a different pressure system.

It should be emphasized that there is no meaning in stating the characteristic properties of an air mass unless it is also stated to what part of the world they refer. The above characteristics refer to temperate regions; it is clear, for example, that what is classed as polar air may develop some properties of warm air on penetrating to higher latitudes, and that stable tropical air may become unstable on moving nearer the equator.

Equatorial air masses

The equatorial belt included between the two trade wind zones forms an extensive source region of equatorial air which for the most part is warm, moist and often unstable, especially after lifting. Subdivisions are conveniently made according to the modifying effects of the different wind systems – the trades and monsoons – each of which has its own characteristics according to the locality and season. For further information, see Section 138 and Chapters 22 and 23.

103. MAIN FRONTAL ZONES

It is implicit in the conception of air masses that in moving from one locality to another the characteristics of the air change slowly, for example, polar air on penetrating the tropics and remaining there must undergo a gradual modification before it may be relabelled as tropical air. Since it is therefore not always possible to draw any definite boundary line between air masses, a zone of transition, more or less wide, must often be recognized. Where different air masses are brought close together the transition may be rapid enough to be represented by a continuous line on a weather map. The change from one air mass to another is never absolutely abrupt but occurs over a transitional zone which may be many miles across, although on the scale of the chart a line is a suitable marking. Such a line is termed a 'front' since it marks the limit of advance (or retreat) of an air mass in its movement over the earth. The term 'frontal zone' may be used when it is desired to emphasize the gradual nature of the change from one air mass to another. It will be seen later that the interaction of two air masses at a front or frontal zone is often responsible for much cloud and precipitation.

We can now indicate the mean position of the fronts, or rather the frontal zones, which mark the transitions between the principal air masses. The mean positions of these zones in the months of January and July respectively, representing the seasonal extremes, are shown on the charts of mean surface winds in these months (Chapter 22). Some further details of these frontal zones will now be given.

Polar front

This marks the boundary between adjacent polar and tropical air masses. Over the North Atlantic its mean position in January extends north-eastwards from

Florida towards the south-west of the British Isles, but in summer it lies further north from the neighbourhood of Newfoundland to northern Scotland. Often it extends eastwards into Europe. The front usually has a wave-like form and it commonly oscillates during a period of a few days over many hundreds of miles, following the movement of the air-mass boundaries when disturbed by travelling pressure systems. It would be useful at this point for the reader to refer to the synoptic charts of Chapter 20 in order to emphasize the great variability in the position of this front, and in the penetration of the air masses on either side of it. A similar front is found in the North Pacific Ocean.

Arctic front

This is the boundary between arctic and polar air masses. It lies further north than the polar front but is often displaced southwards into temperate latitudes in winter.

Mediterranean front

This usually extends roughly from west to east over the Mediterranean in winter and forms the boundary between polar continental air from Europe and tropical continental air from north Africa. There is no corresponding front in this region in summer.

Intertropical front

This lies within the tropics and marks the rather broad zone of separation between air masses conveyed by the trade winds from source regions on opposite sides of the equator, an arrangement which is however subject to modification by monsoonal winds, particularly those of Asia. The front is usually less well defined than the fronts previously mentioned and, partly for this reason, the name 'inter-tropical convergence zone' (ITCZ) is sometimes preferred. Its position normally varies little from to day but it undergoes a regular seasonal shift during the course of the year. (For further details, see Section 159).

104. SOME GENERAL PROPERTIES OF FRONTS

A detailed study of the transitional regions between adjacent air masses was first carried out by Norwegian meteorologists during the First World War when the present terminology was introduced. The differences between the two masses are exhibited by both temperature and humidity as well as other characteristics but the most significant is, as a rule, temperature. Where the warmer mass is displacing the colder, the front is called a 'warm front'; where the colder air is gaining ground, the front is named a 'cold front'. At other times there is little change in position, that is the front is stationary or quasi-stationary. The front, represented by a line on the chart, marks only the dividing line on the earth's surface; the two air masses, however, extend upwards and the division is really a surface in space; this is known as a 'frontal surface'.

Normally the two air masses come from different source regions, but sometimes from the same source by different paths so that some differences in characteristics have developed by the time they are again brought together. It will now be explained why it is possible for two air masses to flow side by side without the colder, heavier mass necessarily flowing under the warmer and lighter mass.

Equilibrium at a front

When a light and a heavy fluid, as for example oil and water, are placed together in a vessel, equilibrium is reached only when the lighter fluid floats horizontally upon the heavier. On the other hand when two air masses of different temperature and so of different density are in proximity, the surface of separation is usually found to be sloping with the cold mass beneath as in Fig. 78. If it is assumed for

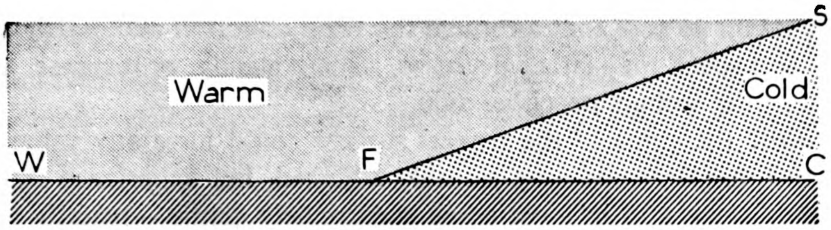


FIG. 78. *Vertical section through a frontal surface*

the present that the surface pressure in the warm air is uniform, then since the cold air becomes gradually deeper from F towards C, the surface pressure, being equivalent to the weight of air above, gradually increases at the same time. The surface isobars will therefore run as in Fig. 79 where FF marks the position of the front. Pressure is uniform on the warm side, where there is no wind. On the cold side the pressure-gradient force is directed towards the front and would be exactly balanced by the Coriolis force associated with the geostrophic wind blowing parallel to the front. The sloping frontal surface then remains in equilibrium with no tendency for the cold air to flow further under the warm.

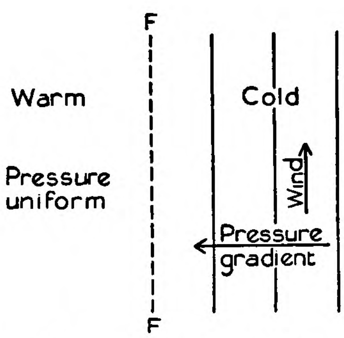


FIG. 79. *Motion of cold air parallel to a front balancing the pressure gradient*

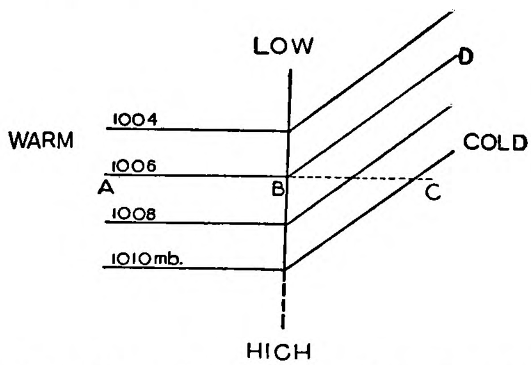


FIG. 80. *Refraction of isobars at a front*

Now the equilibrium of the front would still be maintained if the same uniform velocity were added to both the warm and cold masses. For example, if in the situation illustrated in Fig. 79 with the cold air moving from south a uniform westerly wind were superimposed, then the warm air would be moving from west and the cold air from, say, south-west. The isobars would then appear as in Fig. 80. If we now proceed along an isobar in the direction AB, then after crossing the front at B the depth of cold air increases and so the surface pressure must rise along BC. Consequently the isobar AB cannot continue in the same direction after crossing the front but must change counter-clockwise to some such direction as BD. Thus from whichever side the front is approached, the isobars on the further side appear to be refracted towards the centre of low pressure – in other words, the isobars turn cyclonically on crossing the front. It is seen too that the isobars are sharply kinked where they cross the front.

It follows that a front always lies along a trough, unless it happens to lie parallel to the isobars. The fronts shown on the weather charts in Chapter 20 should be examined as examples of this rule. The rule applies also to the southern hemisphere although the wind direction there must be such as to keep the low pressure on the right.

Other diagrams like Fig. 80 may be drawn but with the isobars differently inclined to the front, or with the positions of the cold and warm air masses reversed In this way it can be seen that although the wind direction necessarily changes at a front, the wind speed may either decrease, remain steady, or increase, according to the spacing of the isobars on either side.

Theoretically the slope of the frontal surface could have any value if the wind velocities and temperatures were suitably adjusted, but experience shows that the slope is rarely steep and in most cases is extremely gentle – usually about 1:100, ranging perhaps from 1:50 to 1:200 or even flatter. An average frontal surface therefore rises 1000 feet in about 100 000 feet or 20 miles, and always slopes upwards from the warm to the cold side.

Velocity of fronts

In Fig. 80 and similar cases where there is motion at right angles to the front as well as relative motion of the air masses parallel thereto, the front is carried along with the winds. In order to estimate the speed of translation it is necessary to determine the component of wind at right angles to the front. The simplest method of measuring the geostrophic component at right angles to any line is that explained in Section 24, namely by laying the geostrophic scale along the line and reading off the speed corresponding to the interval between successive isobars. This is illustrated in Fig. 81 where the speed is seen to be about 12 knots. The method

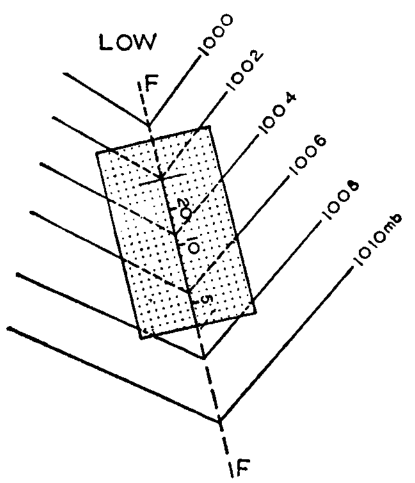


FIG. 81. *Estimating the speed of a front with the aid of the geostrophic wind scale*

gives the correct speed of translation of a front only in so far as the winds are geostrophic, but this is generally a useful approximation. Since the isobars are necessarily continuous on crossing the front, the wind component found in this way is the same in both air masses whence it follows that with geostrophic winds there is no convergence at the front.

Convergence at fronts

The discussion of equilibrium at a front leads to the conclusion that if the winds are geostrophic then there is no tendency for the warm air to rise over the cold and therefore no cooling by ascent and no cloud formation. This fact is fundamental to a proper understanding of fronts; the mere presence of a front does not imply clouds and rain: if these occur it must be because the front is unbalanced, that is because the winds are not geostrophic. The essential property of a front is a difference in temperature between the two sides; disturbed weather is due to disturbed conditions at the front and we must now consider how these may arise.

A discussion of vertical motion in relation to convergence and divergence is included in Section 27 where it was seen that among the factors concerned with upward displacements are departure of the wind from geostrophic, frictional effects and falling barometric pressure. These processes commonly affect the whole area of a depression especially when it is deepening and they are often localized or accentuated in the vicinity of a frontal surface which then becomes the scene of extensive cloud and precipitation.

While the convergence of surface winds at a front causes a net inflow of air near the ground, the magnitude of the effect is insufficient to give more than very slight rain and so cannot be the main cause of frontal precipitation.

With regard to the convergence and ascent which occur in association with falling pressure tendencies, it has been seen that these may result in widespread cloud and precipitation. In a depression it is not difficult to see why the convergence and rainfall tend to be localized near the fronts. Take a frontal trough such as that represented in Fig. 82 and suppose the depression deepens so that the gradients increase; it is

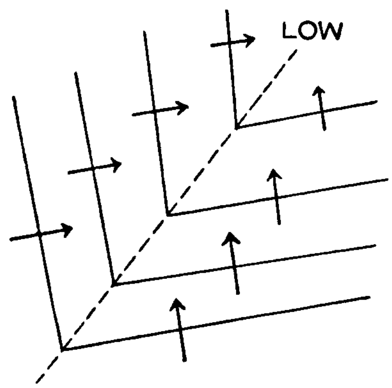


FIG. 82. *Convergence towards a front in a deepening depression*

clear that the effect of falling pressure is to cause a flow of air into the depression roughly at right angles to the isobars as indicated by the arrows, and that the convergence is localized near the front. A word of caution is required here, for as a trough moves pressure must fall ahead of it and rise behind; this gives a system of pressure tendencies due merely to the motion but with no general change of pressure, that is with no development and no convergence. In interpreting the tendencies it is necessary to distinguish real development and deepening from the effect of simple displacement of the isobars without change of gradient – a matter not altogether easy in practice.

It is important to stress that in general a front does not necessarily produce rainfall; for this there must always be some factor setting up convergence and the illustration shows how this occurs in the deepening depression. In regions remote from depressions, convergence is often slight or absent, so that fronts in these regions may produce little or no rainfall.

Changes in the sharpness of fronts

The effect of convergence towards a front is generally to narrow the transitional zone between the adjacent air masses, in other words to sharpen the front. This process is illustrated in Fig. 83 in which it is supposed that there is a gradual fall of temperature across the transitional zone from 12°C in the warm air to 8°C in the cold. The effect of the ageostrophic wind components directed towards the front on both sides is to bring the 8°C and 12°C isotherms closer together, the intervening air being squeezed upwards; the transitional zone thus becomes narrower and the front sharper. The process whereby the frontal zone is narrowed and the change of air mass takes place more quickly is known as 'frontogenesis'; it is the method by which the sharpness of the front is increased or maintained against the natural

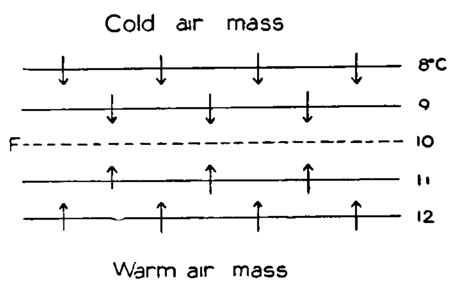


FIG. 83. *Frontogenesis—isotherms brought nearer together by convergence*

tendency towards a diffuse mixing zone. The opposite process which assists the smoothing out of a discontinuity and perhaps leads to the complete disappearance of a front is known as 'frontolysis'. This usually occurs in association with high or rising surface pressures, with descending air and divergence. The effect of divergence would be illustrated by a diagram similar to Fig. 83 but with the directions of all the wind arrows reversed so that the isotherms would become more widely separated. Convergence and divergence are not the only factors which may lead respectively to frontogenesis and frontolysis but they are important, and the above considerations show why in a depression the fronts are usually clearly defined; in other circumstances without convergence they are often difficult to identify and may perhaps disappear altogether.

CHAPTER 16

FRONTAL DEPRESSIONS

105. FORMATION OF A FRONTAL DEPRESSION

The preceding chapter introduced the concept of air masses and indicated how frontal zones develop when differing air masses approach one another. Such frontal zones may become preferred regions for the birth or development of depressions.

Consider, for instance, a stationary of quasi-stationary polar front of the type shown in Fig. 84 and in order to correspond with normal conditions in north temperate regions, let us take the warmer air to the south and the colder to the north as in Fig. 84 (*a*).

It was explained in Section 104 that such a front may be in equilibrium if the winds are geostrophic. Now the surface of discontinuity between two air masses is, like the surface of the sea, subject to the formation of waves. In light winds, ocean waves are of the stable kind, limited in size and moving without change of form; when the winds are strong, the waves are unstable and increase in height until the tops are eventually sheared off by the wind. Both stable and unstable waves occur in a somewhat similar manner at the gently sloping frontal surface between two air masses, except that the wavelengths are very much greater. Because of the slope of the surface of separation, the wave motion gives rise to a horizontal oscillation in the line where the frontal surface meets the ground so that the warm air at the ground forms a bulge into the cold air. When the wave motion is unstable, the amplitude increases and a depression is formed. In the formative stage the depression enlarges and deepens but eventually it begins to fill up and decay, as will be explained later. Investigation shows that frontal waves are unstable only if the wavelength lies between about 800 and 2000 miles and if there is a sufficiently rapid change (shear) of wind between the two air masses. In other cases the front merely oscillates without the development of a depression.

With the unstable waves a fall of pressure occurs where the warm air intrudes into the cold and a cyclonic circulation is then created. The front, being displaced with the wind, passes through the stages (*b*) and (*c*) to the very much distorted shape of Fig. 84(*d*). In the original state the front is almost stationary, but as the depression develops the fronts move with the circulating winds – the section in advance of the centre where warm air is displacing cold air is therefore marked as a warm front, the section behind as a cold front. The region between the warm and cold fronts on the warmer side is occupied by the warmer air and is referred to as the 'warm sector'; the remaining and larger portion is the 'cold sector'.

In addition to the process of deepening or development there is a general translation of the system, for the winds have everywhere a velocity roughly equal to the geostrophic value and the fronts are carried along with them. The depression itself, travelling with the fronts, usually has a velocity roughly equal in speed and direction to that of the geostrophic wind in the warm air.

Continuously falling pressure implies convergence into the depression so that air is forced to rise. It was noted in Section 104 that the upward motion tends to be localized near the fronts. In attempting to explain the process by which the air ascends at the fronts, it is frequently said that the warm air slides upwards over the

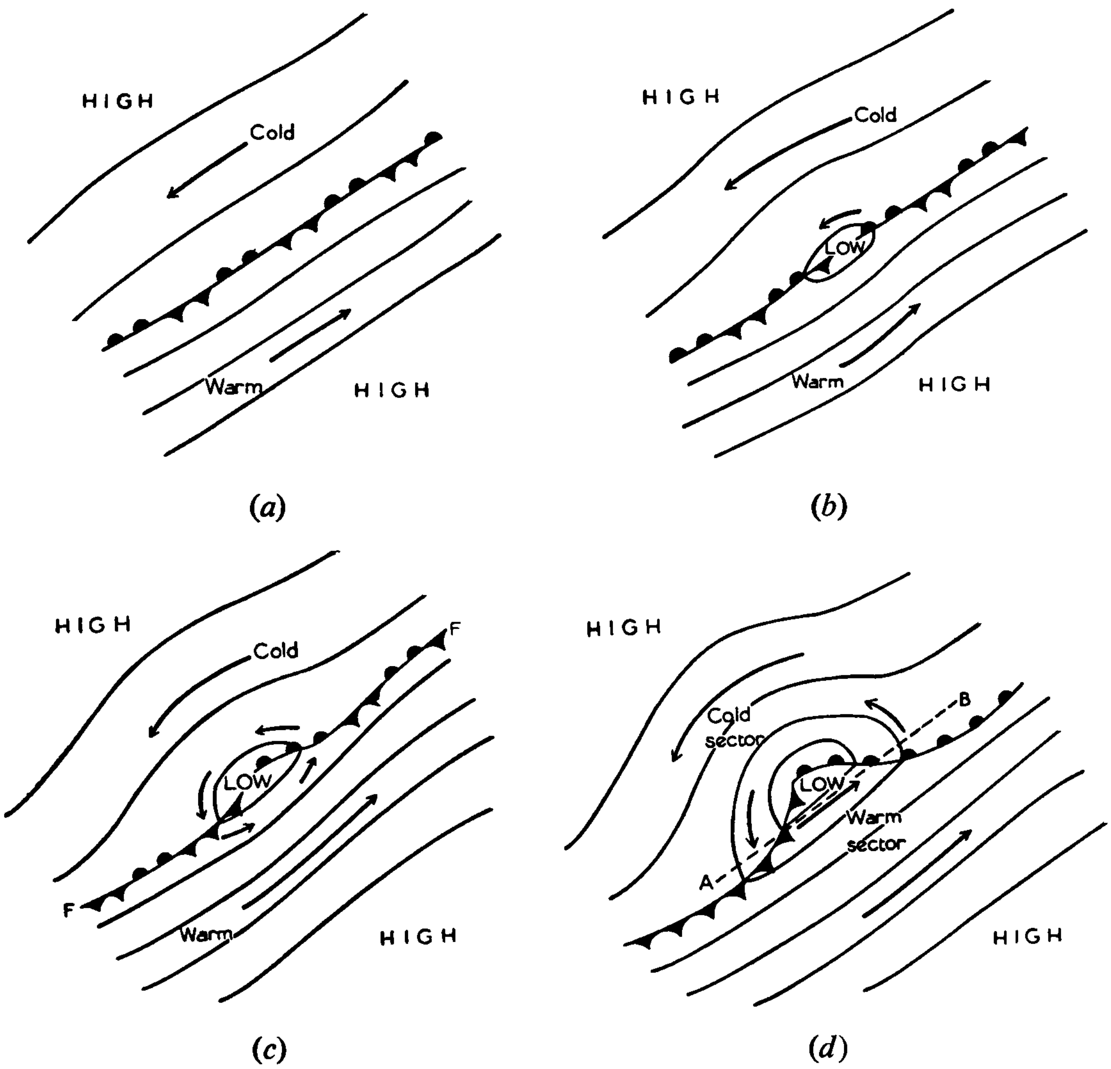


FIG. 84. *Formation of a warm-sector depression on a stationary front*

cold air at the warm front and that the cold air undercuts the warm at the cold front, as though the cold air at either front were a solid wedge. This simple description may be made more explicit by stating that in a direction at right angles to the warm front the warm air moves faster than the cold, while at right angles to the cold front the cold air moves faster than the warm. At both fronts this relative motion is associated with convergence and upflow. Sections 106-108 follow this treatment, but it is as well to emphasize that it is only a description of events and does not provide an explanation of the convergence at the fronts, which is a normal condition within a developing frontal depression but not a necessary condition of all fronts. The structure of a front is consequently related to the local pressure distribution and its changes, and the following descriptions apply in general only to conditions within a depression.

106. WARM FRONT

Fig. 85 illustrates a vertical cross-section through the fronts of a depression along the line AB on Fig. 84(*d*) and intersecting the warm front at W and the cold front at C. The tropopause is higher over the warm air than over the cold air. If a jet stream is associated with the fronts it will be about the position marked J in the diagram.

Even if the wind speeds are not of jet-stream values the strongest winds are likely to be found in a similar position. Ascent near the warm-front surface produces adiabatic cooling and clouds are formed above the frontal surface. The highest clouds are cirrus, often at 30000 feet or above, and as the height of the frontal surface decreases the cloud type passes through cirrostratus to altostratus and finally to nimbostratus. Rain or snow begins to fall from the higher altostratus, perhaps from 20000 feet, but is slight at first and evaporates before reaching the ground; as the clouds lower and thicken, the rain becomes heavier and ultimately reaches the ground some 150–200 miles in advance of the front. From then onwards rain is likely to be continuous until the front arrives, after which the rain clouds quickly clear.

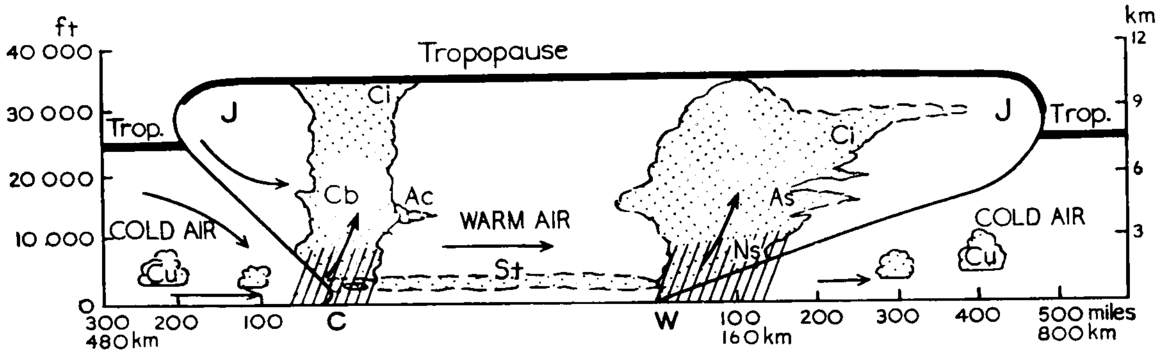


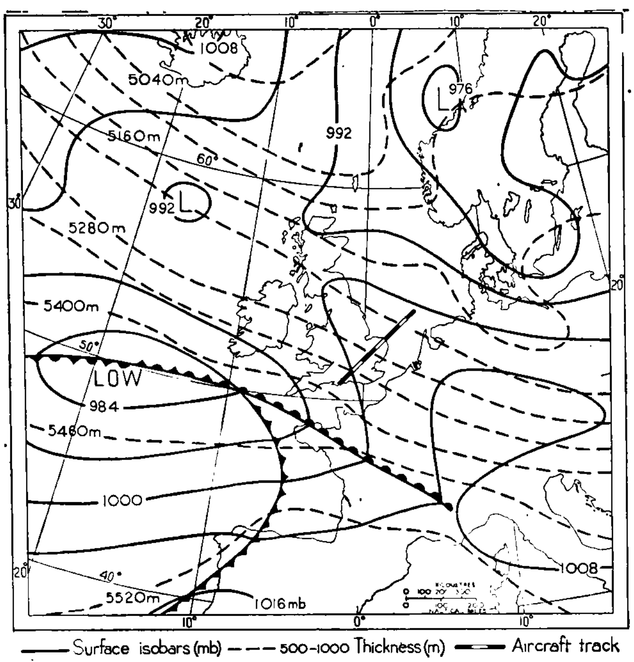
FIG. 85. *Vertical cross-section through the warm sector of a depression*

J Likely position of jet stream

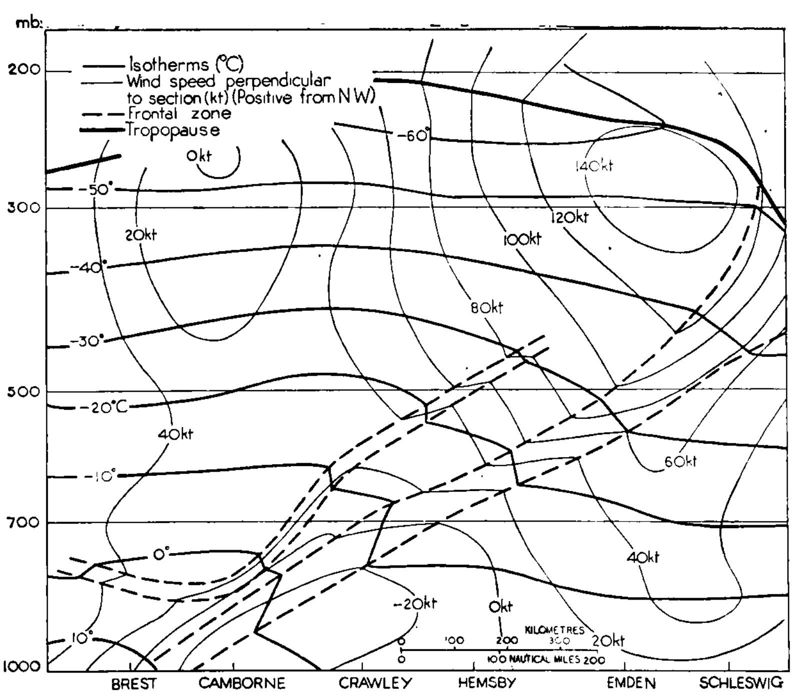
Apart from the true frontal clouds there is generally considerable low stratus fractus or cumulus fractus (scud) formed by turbulence within the cold air below the altostratus and nimbostratus, for the cold air becomes very moist by evaporation of falling rain. The depth and extent of the cloud varies considerably from one front to another; it may extend in a solid mass from the ground to the tropopause, or on other occasions it may be layered with clear lanes between the layers. Research flights through fronts have demonstrated clearly the variability of cloud structure, and the not uncommon existence of tongues of very dry air in or near the frontal zones. Usually the slope of the forward edge of the cloud mass is steeper than that of the frontal surface. The transition from cold to warm air usually occurs over a distance of several hundred miles, with a belt of more rapid temperature change in a narrow frontal zone which may be as little as 25 miles across or as much as 100 miles.

In addition to the cessation of continuous rain, the passage of the surface front brings a veer of wind (as is common to all fronts) together with an increase in temperature and dew-point corresponding with the change in air mass. Normally the pressure falls steadily ahead of the front; after the passage the fall ceases, or continues at a slower rate. The lighter winds within the friction layer cause the frontal surface to advance more slowly near the ground than aloft; this in turn produces a very shallow frontal slope near the ground which combined with vertical mixing means that the surface changes at the frontal passage are seldom sharp. This description applies to a somewhat idealized warm front from which the actual fronts usually differ in one or more particulars. For example, with low humidity in the warm air the frontal cloud may be patchy or may develop in a number of separate layers and the pre-frontal rain may be slight or absent. Exceptionally the warm air may become unstable as it ascends the frontal surface and then warm-front thunderstorms become possible. In fact the real fronts show important departures from the ideal on most occasions but the ideal model nevertheless forms an essential background to any discussion of fronts.

Fig. 86 shows details of a warm front investigated on 13 January 1955 by the Meteorological Research Flight. Features, common to most warm fronts, were: change of temperature through the frontal surface; core of strong winds in the warm air just below the tropopause; tropopause higher over the warm air than over the cold air; dry zone shown by the depression of the frost-point; slope of the forward edge of the cloud mass steeper than the slope of the temperature change.

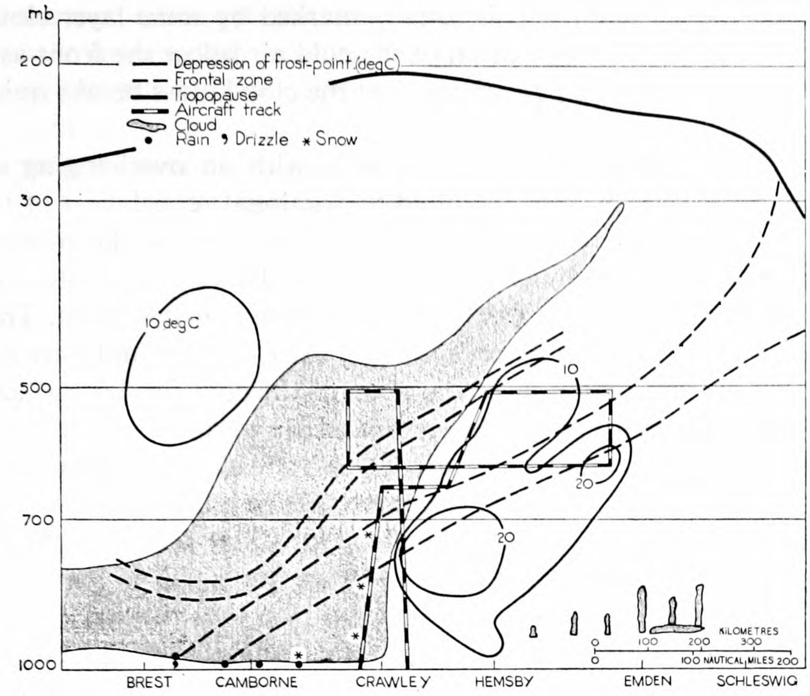


(*a*) Synoptic chart for 1400 GMT



(*b*) Vertical cross-section from Brest to Schleswig showing isotherms and isotachs

FIG. 86. *Warm front of 13 January 1955*



(c) Vertical cross-section showing humidity and cloud

FIG. 86. *Continued*

107. COLD FRONT

It might be natural to expect the cold front to present the same phenomena as the warm front with merely a reversal in the order. As the cold air undercuts the warm air the ascending motion might be expected to produce layer clouds above the frontal surface with rain beginning after the passage of the front and extending in a wide belt behind. On some occasions cold fronts do behave in this way but more usually the relative motion is of the kind shown in Fig. 87. The warm air ascends only in the lower levels; at high levels it descends along the frontal surface while at middle levels there is a horizontal component away from the front. The cold air behind the front is generally subsiding. If the warm air is convectively unstable, the ascending motion ahead of the cold front may lead to outbreaks of instability (Section 35) with the formation of towering cumulus or cumulonimbus clouds and heavy rain of short duration, perhaps with hail or thunder. In these circumstances the rain belt may extend on both sides of the surface front and be heralded by altocumulus clouds carried forward by the increased wind at that level. The frontal surface, sloping

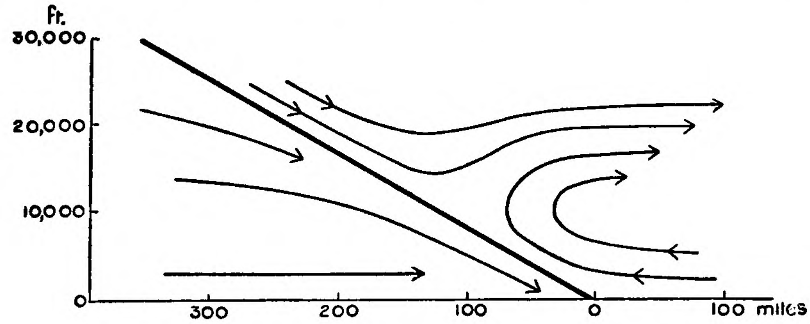


FIG. 87. *Streamlines of motion relative to a cold front*

upwards behind the surface front, is usually marked by some layer cloud of alto-stratus type but the descending motion of the cold air below the front as well as of the warm air above the upper part ensure that the cloud mass breaks quickly to the rear of the front.

In Fig. 85 the cold frontal surface is drawn with an overhanging nose some 2000 feet above the surface. This is caused by frictional retardation of the surface air and is a transient feature, which is alternately formed by the over-running of the air above and destroyed by convective mixing. This process often results in a characteristic roll of low cloud extending along the line of the front. The advance of the surface front thus takes place in an unsteady manner and may give rise to severe gusts. Above the friction layer the slope of the cold frontal surface averages about 1:50 and so is usually steeper than that of the warm front.

The phenomena of the cold front are liable to wide variations. Some fronts pass with little disturbance, some are accompanied by a line squall, sometimes the passage is not immediately evident in the cloud structure as seen from the ground because of the presence of low rain-cloud. Often the squall is the most striking feature. The passage of the front is accompanied by a veer of wind sometimes preceded by pre-frontal backing. On occasions the direction may veer temporarily by as much as 180° before settling back to the new direction in the cold air. For example, the general pre-frontal wind may be south-westerly but with the approach of the front the direction may back to south veering in the squall to north and later settling back to north-west. Temperature falls more or less suddenly and the drop in dew-point is generally well marked. Pressure normally falls somewhat in advance of the front but begins to rise rapidly after the front has passed. When the instability conditions are well developed the front may bring a sudden jump in pressure of a millibar or more, clearly observable on the barogram.

The wide differences between individual cold fronts have been observed by research flights. The cloud masses usually occur in the warm air with the slope of the cloud surface steeper than that of the frontal surface. Dry areas in or near the frontal zone, similar to those shown in the cross-section of the warm front (Fig. 86(*c*)), have been observed at cold fronts.

Although there are large differences in cloud and weather associated with different cold fronts, most of them can be placed in one of two main types termed cold anafronts and cold katafronts. The cold anafront is one in which the warm air is ascending and moving forward less rapidly than the frontal surface. Heavy rain usually occurs at the frontal passage with steady light rain for some time behind the front. It is usually accompanied by a large fall of temperature and a sharp veer and sudden decrease of wind. A cold katafront is characterized by descending warm air which is moving forward faster than the frontal surface; rainfall is usually very slight at the front, the temperature drop is slight and gradual and the wind may veer only very gradually with little change in speed.

108. OCCLUSION

As the air of the warm sector is gradually moving upwards above the warm frontal surface and to a less extent over the cold front, the amount of warm air lying on the surface and therefore the area of the warm sector gradually decrease. This process appears on the weather map as the overtaking of the warm front by the cold front and when eventually there is no more warm air left at ground level, the two fronts coalesce. The surrounding or enclosing of the depression by cold air is called the 'occluding' process; when complete the depression is said to be 'occluded' and the single composite front is known as an 'occlusion'. If the cold air in advance of the warm front and that behind the cold front had identical properties there would be no surface front at the occlusion, but having been widely separated the two cold masses are likely to have been subjected to noticeably different conditions since the depression began to form; there is, therefore, in general a contrast between them although which is the colder is a matter of circumstances. In either case the occlusion lies along a trough extending outwards from the depression usually in a curved path, for the fronts are carried round more rapidly by the stronger winds near the centre than by the weaker winds in the outer parts.

Following from the warm-sector depression of Fig. 84 the further stages are normally as in Fig. 88 although the precise position and shape of the line of occlusion is subject to much variation. In Fig. 88(*a*) the occlusion is about to appear at the centre; in (*b*) the occlusion extends from the centre to O while the portions OW

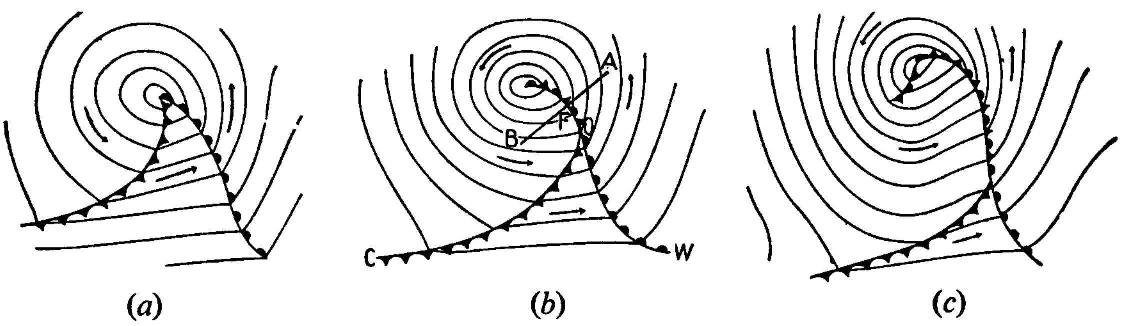


FIG. 88. *Occluding of a depression*

and OC remain as warm and cold fronts respectively; in (*c*) the occlusion extends further, the portion near the centre has become twisted round and there is a trailing end caused by the displacement of the centre along the occlusion. This trailing end which continues to travel round the centre is known as the 'back-bent' occlusion. The occlusion is said to be of the cold-front type or the warm-front type according as the air behind the front is colder or warmer than that in advance of the front. The vertical section through the front along the line AFB is represented in Fig. 89 for each type. The two types are distinguished primarily by the position of the rain

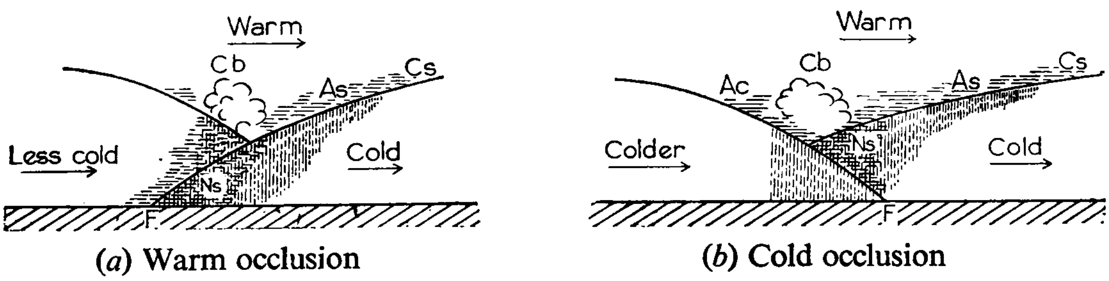


FIG. 89. *Vertical sections through occlusions*

area which in the warm type is in advance of the front but in the cold type both in advance and in the rear. The rainfall is caused partly by the continued ascent of the warm air above the surface and partly by the interaction of the two cold masses near the surface. The occlusion therefore usually continues to give low cloud and rain for some time after its formation, although the tendency is towards a gradual cessation except when the two cold masses have very different characteristics – then the rain may be persistent. Over western Europe the warm-front type is normal in the winter, for the cold air in the east is of continental origin and therefore colder than the maritime air from the west. In summer, conditions are reversed and the polar maritime air is generally the cooler.

It is difficult to give what may be called characteristic features of an occlusion apart from a more or less clearly marked trough of low pressure. The mere information that an occlusion lies along a route is little indication of the weather conditions to be encountered. Sooner or later an occlusion degenerates and disappears from the weather map and in its later stages may present no noticeable features other than perhaps a little cloud.

109. SUMMARY OF FRONTAL CHARACTERISTICS

The normal features associated with the fronts of a depression are summarized in Table 11. It is, however, important to remember that in practice depressions and fronts rarely if ever agree with the ideal model in all respects and that of the characteristics described above and now tabulated, some are sure to be more or less modified in any particular case.

The characteristics of the occlusion are very variable. They may be similar to those of either the warm or cold front (according to type) but are often ill defined.

110. GENERAL DISTRIBUTION OF WEATHER IN A FRONTAL DEPRESSION

In regions away from the weather intimately associated with the fronts, the warm and cold air masses display their own characteristic features. A distinction is therefore made between the frontal weather and the air-mass weather of a depression and some account of the latter will now be given.

Warm-sector depression

Warm sector. The air within the warm sector is typically tropical maritime or continental and the conditions are as described for those air masses in Section 102 with appropriate variations according to locality, season and time of day. Over the Atlantic and north-west Europe the warm air is usually tropical maritime and widespread low stratus or stratocumulus with perhaps fog or drizzle are common, but clear skies can occur when the humidity is comparatively low. When the air mass is tropical continental, as for example with a Mediterranean depression, clear skies would be the rule while even the warm-front cloud and precipitation may be unable to form because of the extreme dryness of the air. Isobars within the warm sector are usually straight and parallel but not necessarily equally spaced. Pressure falls slowly or remains steady depending on whether or not the depression is deepening.

Cold sector. This includes the greater part of the area covered by the depression and conditions within it are those typical of a polar air mass, either maritime or continental, as given in Section 102. There are, however, modifications because of the frontal developments. In the cold air in advance of the warm front when the sky becomes covered with high and medium cloud, surface heating by the sun is reduced and the diurnal cumulus type of cloud is either very restricted in the vertical because of the shallow depth of cold air or does not form at all. There may be also some pre-frontal subsidence of air, making the lapse rate more

TABLE 11. *Normal frontal characteristics*

Element	*in advance*	*at the passage*	*in the rear*
		WARM FRONT	
pressure	steady fall	fall arrested	little change or slow fall
wind	backing and increasing	veer and decrease	steady direction
temperature	steady, or slow rise	rise	little change
dew-point	rise in precipitation	rise	steady
relative humidity	rise in precipitation	may rise further if not already saturated	little change; may be saturated
cloud	Ci, Cs, As, Ns, in succession; St fra, Cu fra below As and Ns	low Ns and St fra	St or Sc may persist; perhaps some Ci
weather	continuous rain (or snow)	precipitation almost or completely stops	dry, or intermittent slight precipitation
visibility	good except in rain (or snow)	poor, often mist or fog	usually moderate or poor; mist or fog may persist
		COLD FRONT	
pressure	fall	sudden rise	rise continues more slowly
wind	backing and increasing, becoming squally	sudden veer, perhaps squall	backing a little after squall, then fairly steady or veering further in later squalls
temperature	steady, but fall in pre-frontal rain	sudden fall	little change; variable in showers
dew-point	little change	sudden fall	little change
relative humidity	may rise in pre-frontal precipitation	remains high in precipitation	rapid fall as rain (or snow) ceases; variable in showers
cloud	St or Sc, Ac, As then Cb	Cb with St fra, Cu fra or very low Ns	lifting rapidly, followed for a short period by As, Ac and later further Cu or Cb
weather	usually some rain perhaps thunder	heavy rain (or snow) perhaps thunder and hail	heavy rain (or snow) for usually short period, but sometimes more persistent; then fine but followed by further showers
visibility	moderate or poor, perhaps fog	temporary deterioration followed by rapid improvement	very good

stable than in a typical cold air mass. This clearance of low cumuliform clouds on the arrival of the high clouds of a new depression is often very noticeable. In the cold air behind the cold front, subsidence often clears the sky completely soon after the frontal rain has ceased; there may then be an interval of clear weather lasting a few hours before the convection clouds and showers of the cold air begin to appear.

By combining the results of the preceding paragraphs a diagram (Fig. 90) may be constructed showing the distribution of clouds and weather over the area of an idealized warm-sector depression.

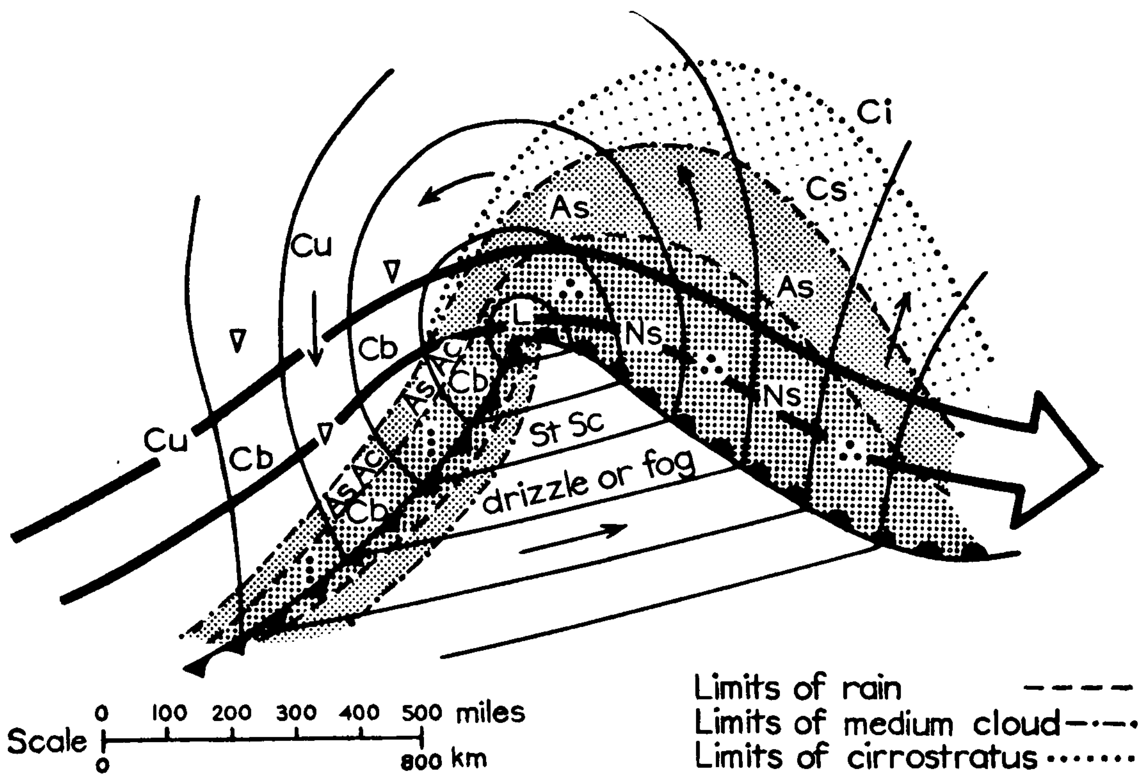


FIG. 90. *Distribution of cloud and weather in an idealized warm-sector depression*

The broad arrow marks the likely position of the jet stream (if present).

Occluded Depression

The depression usually continues to deepen until an advanced stage in the occluding process is reached. By then the supply of warm air has been cut off and the depth of the depression subsequently changes little. At the same time the centre, previously carried along with the warm air, becomes almost stationary. The last stage in the life of a normal frontal depression is thus a stationary or slowly moving system of circulating polar air. This decaying depression fills up very slowly and is often very persistent unless the circulation is destroyed by the approach of a new vigorous depression, in which case an old and deep occluded centre may fill up completely within 24 hours. During the occluding and decaying stages, the weather usually tends to improve slowly. The frontal cloud and rain become gradually less extensive but instability cloud and showers may become more general, their intensity and extent varying with the characteristics of the cold air and the thermal and dynamical processes to which it is subjected. Hence the weather in old depressions ranges from conditions of little cloud to widespread convective cloud with showers and thunderstorms. During the decay of the depression the winds decline slowly and clearing skies over the land at night may permit radiation fog to develop.

111. FAMILIES OF FRONTAL DEPRESSIONS

The development of a frontal depression occurs as a rule on a long slowly moving front between polar and tropical air in temperate latitudes. When one depression develops, moves along the front and finally becomes occluded, the cold front trails back from the point of occlusion and remains continuous with the more or less undisturbed front behind. Conditions may then be favourable for a new development passing through the same stages and in this way there forms a series or family of depressions. As each occludes in turn the cold air spreads round it and penetrates to lower latitudes so that each successive depression tends to follow a more southerly track until at last the cold air sweeps through and goes to feed the trade winds of lower latitudes. By this time the polar front has been displaced far to the south of its normal position and a large anticyclone builds up in the polar air, so breaking the continuity of the front and terminating the family. Meanwhile a new family starts to form on the north-west side of the anticyclone and as the latter drifts away south-eastwards the whole process may be repeated. There is no regularity about the number of individual depressions in a family but there are often three, four or five. The weather changes associated with the passage of the depressions, alternating with brief fine periods in the high pressure ridges dividing them, are typical of 'unsettled' conditions in the British Isles.

112. UPPER WINDS OVER FRONTAL DEPRESSIONS

The winds at an upper level can be estimated by adding to the geostrophic wind at a lower level a 'thermal component' determined from the horizontal distribution of mean temperature in the intervening layer (Section 26). As a frontal deppression is essentially a phenomenon of temperature contrasts, it is to be expected that marked changes of wind with height will occur within it. Over a frontal depression in the northern hemisphere, the mean isotherms or thickness lines between the surface and 500 millibars are usually distributed somewhat as shown in Fig. 91. Since a cold region acts as a region of low pressure in the upper atmosphere the thermal winds blow in the directions indicated in the figure, and when these are added vectorially to the geostrophic wind near the surface a good indication of the upper winds is obtained. In advance of the depression, the upper winds at a high level tend to become north-westerly above the warm-front surface. When, as is often the case with Atlantic depressions in winter, the air to the north-east is very cold the thermal north-westerlies may be strong. A rapid movement of the cirrus clouds from this direction is often a valuable indication of the approach of a warm-sector depression from the west. Similarly to the rear of the depression, the thermal effect accounts for south-westerly winds at high levels above the cold-front surface. Thus when the thermal effects are pronounced, strong upper winds are found blowing approximately parallel to the surface fronts but displaced towards the cold air; these are the jet streams of temperate regions which have been described in Section 67.

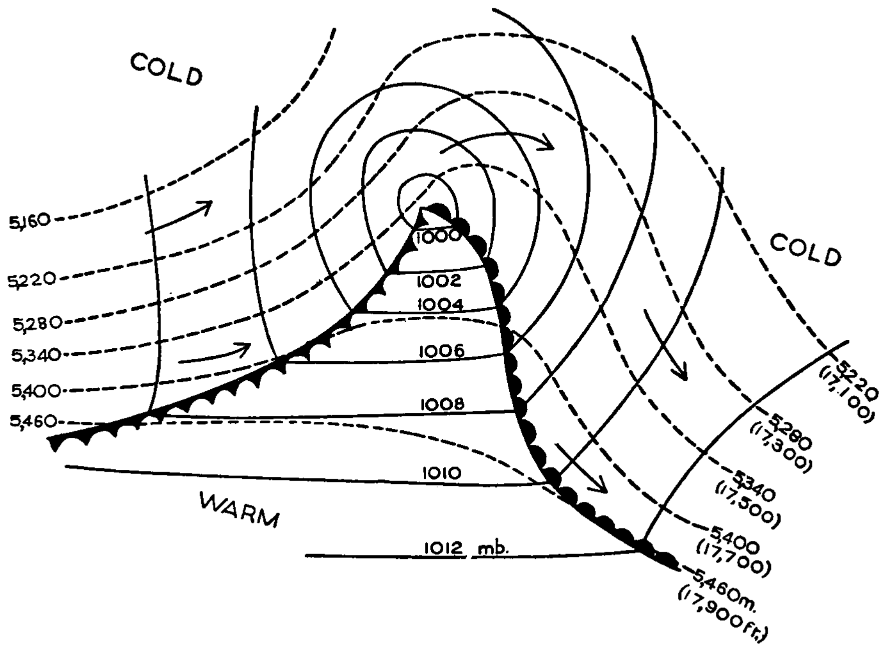


FIG. 91. *Thermal winds in a warm-sector depression*

The broken lines show the distribution of mean temperature by means of the thickness of the 1000–500-millibar layer; the arrows show the directions of the corresponding thermal winds.

With regard to the pressure distribution aloft, it will be seen that at a level high enough for the surface distribution to be overcome there is a ridge of high pressure just ahead of the ground position of the warm front and a trough of low pressure just to the rear of the cold front. Alternatively the contours of a given pressure surface aloft will be seen to be elevated in advance of the warm front and depressed in the rear of the cold front as illustrated in Fig. 92. Within the warm sector little change of wind with height would be expected since the thermal gradients are slight, as illustrated in Fig. 91. North of the centre, the superposition of the thermal westerlies on the surface circulation often eliminates the weak low-level easterlies so that the distribution of the isobars or contours aloft then shows no closed centre but merely a wave-like deformation of the prevailing westerlies.

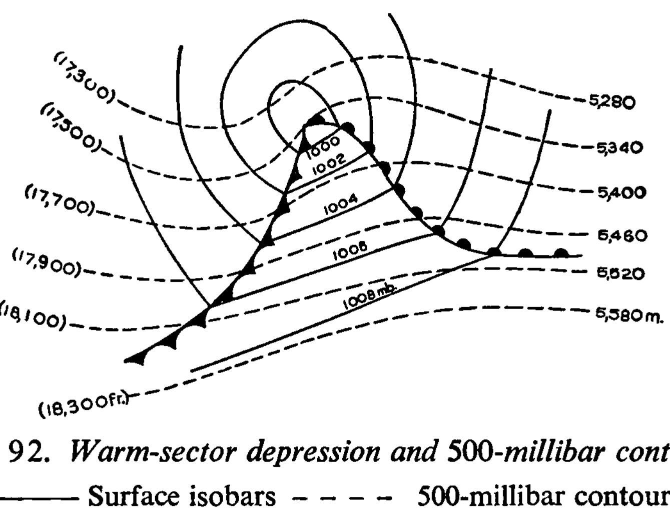


FIG. 92. *Warm-sector depression and 500-millibar contours*

——— Surface isobars – – – – 500-millibar contours

In an old occluded depression entirely surrounded by polar air, the horizontal temperature gradients are relatively weak and the variation of wind with height is less pronounced than it is in the warm-sector depression. The roughly circular form of the isobars (or contours) near the surface is often maintained up to high levels and forms what is known as a 'cold pool', since an area of low contour heights implies a low mean temperature of the air column.

Thus when a frontal depression is regarded as a circulation in three dimensions, it is observed that the formation, development and final occluding transform the circulation from an asymmetrical system, with marked surface troughs and large changes of wind speed and direction with height, to a roughly circular symmetrical depression with the air simply rotating round the centre and the wind varying little with height. While the conception of a depression as a column of rotating fluid extending upwards through the atmosphere may apply to an occluded centre, it has no application to the active frontal depressions responsible for most of the bad weather.

The above discussion takes no account of the frictional effect at the ground, which gives a veer and increase of wind with height through the friction levels in all normal circumstances (backing and increasing in the southern hemisphere).

113. FLIGHT THROUGH FRONTAL DEPRESSIONS

It is a useful exercise when studying synoptic charts to lay off specimen routes and to write out accounts of the weather likely to be experienced along them. The following notes apply to the idealized depression in accordance with the frontal theory and are not likely to represent any actual case absolutely faithfully, for there are always minor differences and allowances must also be made for diurnal and seasonal variations. Thus each chart must be examined in detail before a reliable forecast can be made.

Flight at low levels

To fix ideas, suppose that, in the northern hemisphere, a warm-sector depression is approaching from the west and a flight from east to west is made at about 5000 feet from well in advance of the depression, through the warm sector and out into the cold air beyond. To begin with, in the ridge, winds are probably light and variable with perhaps some broken cumulus. The first sign of the depression is the cirrus of the warm front perhaps 500 or 600 miles in advance of the surface front; as this thickens to cirrostratus the low clouds clear except for local orographic effects, the wind becomes southerly and increases but away from the surface turbulence the air is not bumpy. A period of good flying weather is to be expected at first but the cloud above gradually thickens to altostratus and later nimbostratus while rain starts to fall and soon becomes continuous (sleet or snow if the temperature is near or below 0°C). The cloud base lowers continuously and may ultimately reach almost to the surface – low flying is then hazardous as winds are probably strong and turbulent, while high ground is obscured by cloud. A period of instrument flying should be expected and the height of flight must be chosen to give adequate clearance of high ground. In this connection allowance should be made for any over-reading of the altimeter on account of falling surface pressure. A previous study of the weather map should have given a useful indication, or a report may be obtained while in the air; but failing such help a considerable margin of safety is necessary, for the drop of pressure from the ridge to the trough may be as much as 30 or 40 millibars causing the altimeter to over-read by more than 1000 feet if the subscale setting has not been adjusted.

The passage into the warm sector will usually be made evident by the cessation of rainfall and although the warm sector may be cloudy, clear air will amost always be found at a moderate height, usually above 4000 feet, except perhaps over high ground. Temperature observations are also helpful in confirming that the warm air has been entered.

As the wind may have veered during the passage through the front, it is important to fix position. Flight through the warm sector usually presents little difficulty except perhaps in regard to navigation when low cloud obscures the ground. The pilot should avoid descending through the clouds to fix position unless the altimeter correction is accurately known, for the cloud base may be very low with drizzle or fog on the surface. Normally on a long flight the warm sector will be traversed in the smooth clear air above the low clouds.

The approach to the cold front will be recognized by the pre-frontal high or medium clouds, usually altocumulus, thickening and lowering to a heavy bank on the horizon. An attempt to fly below the clouds may be successful if the front is not severe but unless information is available to the contrary it should be assumed that the front will have line-squall characteristics with violent bumpiness, heavy thundery rain, patches of very low cloud and severe icing if the temperature is below 0°C. Cloud flying in these conditions is clearly not to be recommended. If the clouds are thick and unbroken, the pilot should consider the advisability of flying above them, which would probably mean climbing to at least 15 000 or 20 000 feet. If conditions are severe and a climb is impracticable, it might be advisable to consider reversing course or perhaps making a landing.

The width of the belt of violently disturbed weather does not usually exceed some 20 miles and after this has been negotiated a complete clearance is likely to follow. A veer of wind usually accompanies the passage of the front and allowance should be made for this in navigating. Conditions in the cold air will remain bumpy and secondary cold fronts or squally showers may be encountered but, generally speaking, flying conditions will improve.

Flight at medium levels

The conditions encountered when flying through a depression vary considerably with the altitude. We consider now a flight in the vicinity of 500 millibars (about 18 000 feet) on a track similar to the one already described. The pre-frontal cirrus thickening to cirrostratus will again be the first visual indication of the approach to the warm front, while below the aircraft the cumuliform cloud will become less frequent. At flight level the winds will be westerly at first, but ahead of the surface warm front they will veer perhaps as far round as north-west. At the same time a slow increase in air temperature and the presence of stratiform cloud will indicate that the upper frontal zone is being traversed. The total horizontal distance through the frontal cloud may amount to about 100 or even 200 miles, and, since the temperature will be below 0°C, airframe icing may be encountered.

Once the upper frontal cloud has been left behind, conditions will usually be clear except that the ground is likely to be obscured by low stratocumulus or stratus cloud or fog. As the flight continues westwards, the wind will slowly back and decrease until the aircraft reaches a point over the surface position of the warm front when the upper wind will again be approximately westerly; from this point on there will be little change until the aircraft passes over the surface position of the cold front, when the wind will back further and increase under the thermal influence.

In the clear air, the cloud system of the cold front will be visible from a great distance. It may be desirable to increase height in order to fly above the pre-frontal altocumulus or altostratus, since flight in these would cause the cumulonimbus immediately ahead to be obscured. For the same reason, any anvil cirrus should in general be avoided. Provided the cumulonimbus clouds are kept in view, they can often be avoided at this altitude even though individual peaks may be much higher; but if the front is very active, a climb is likely to be required before a gap can be found. If after due consideration it is decided to fly through the clouds, then appropriate precautions should be taken as described in Section 43 to safeguard against the effects of severe turbulence, lightning and other hazards. During the approach to the cold frontal surface, the wind at flight level gradually backs to about south-west and increases. Within the frontal zone, the wind again begins to veer and decrease and this process continues in the cold air until wind direction finally becomes westerly once more. The passage through the frontal zone is a gradual process marked by a slow decline in temperature and usually by dispersal of the cloud system.

Flight at high levels

With increase of height above the medium levels, the disturbance caused by the clouds and wind of the depression becomes gradually less marked; the warm-front cloud would then be flown through at cirrus levels while the only disturbance from the cold front would be isolated peaks of cumulonimbus, which could be easily avoided, and perhaps some anvil cirrus. At these levels the general wind is westerly and the modification produced by the depression would be expected to take the form of a broad north-westerly current east of the surface position of the warm front and a broad south-westerly current to the west of the surface cold front. The conditions become very different, however, if a jet stream exists where the warm or cold frontal surfaces approach the tropopause. This case, together with other aspects of high-altitude flight, has been discussed in Chapter 10.

CHAPTER 17

OTHER DEPRESSIONS

114. CAUSES OF DEPRESSIONS

The discussions of the two preceding chapters have emphasized the relation between air temperature or density and the pressure field in the neighbourhood of fronts and frontal depressions, but even in these cases the development cannot be attributed solely to changes of temperature or density. In order that a depression shall form and surface pressure fall from, say, an average of 1012 millibars to perhaps 970 millibars over a wide area, some 4 per cent of the air must be removed, for pressure is given by the weight of the air above. The development of the depression leads, however, to inflow towards the centre which tends to an accumulation of air and to a rise of pressure. It is evident, therefore, that air must be removed somewhere in the upper atmosphere and that convergence near the surface is to be regarded as a result of changing pressure and not as the cause of the depression. Similarly, with an anticyclone, the outflow of air near the surface tends to destroy the high pressure. Thus in many cases the cause of the formation of depressions and anticyclones must be sought in the upper atmosphere. This is a fundamental problem which is still far from being completely solved, but with adequate upper air observations it is often possible to identify areas in the middle troposphere where theory indicates that the form of the thickness pattern is associated with cyclonic or with anticyclonic development at lower levels. Such considerations however are more appropriately discussed under weather forecasting and an account will be found in Chapter 19.

It is to be noted that frontal depressions are predominant only in temperate latitudes. The impression that all depressions are associated with warm sectors or occlusions should be avoided, even in temperate regions. Other types of depression fall mainly into two classes:

(i) Thermal depressions, associated with surface heating or vertical instability.
(ii) Orographic depressions which form in the lee of mountain ranges.

Separate consideration however will be given to:

(*a*) depressions which act as secondaries to already existing depressions,
(*b*) tropical revolving storms,
(*c*) tornadoes and waterspouts, and
(*d*) troughs of low pressure.

115. THERMAL DEPRESSIONS

Owing mainly to the distribution of land and sea the surface layers of the atmosphere are subject to unequal heating and there is a tendency for the warmer regions to become areas of low pressure. This comes about somewhat as follows. As the air is heated it expands and the overlying isobaric surfaces are lifted. At any given upper level, the pressure becomes higher than over the surrounding parts with the result that the air starts to move outwards. This in turn reduces the surface pressure and an inflow takes place at low levels which under the influence of the geostrophic force is converted into a cyclonic circulation. Since the higher pressure and outflow aloft imply an anticyclonic circulation at high levels, the thermal depression or 'heat low' weakens with height; the cyclonic winds decrease and are often reversed in the upper atmosphere. However, much depends on the vertical stability and on conditions in the upper atmosphere, so that situations in which warm surface air is associated with high surface pressure can also occur; indeed, the hottest weather in western Europe is almost always anticyclonic. In such cases the lapse rate is so stable that surface heating results at most in shallow lows of little significance. On the other hand, if the lapse rate is unstable or becomes so as a result of continued heating so that instability showers break out, then there is an increased likelihood of a fall in surface pressure and development of a depression. In these cases additional energy is provided by the liberation of latent heat of condensation of water vapour and if this takes place on a large scale the depression may become a major feature of the weather map. There are several types of thermal low to which attention will now be drawn.

Monsoon low

The most obvious example of a thermal low is that which tends to develop over a large continent in summer; the south Asiatic monsoon low which controls the general circulation over that area is the outstanding one. The weather in a monsoon low does not follow any regular pattern, being very dependent on topography and on the characteristics of the air masses which are drawn into the area of the depression. Some account of the associated weather over the areas concerned will be found in Chapter 23.

Equatorial low-pressure belt

This may be regarded from the present point of view as a permanent heat low encircling the earth in tropical latitudes. The lapse rate is generally steep and there is heavy showery precipitation. For further details, see Section 159.

Polar-air depressions

In temperate latitudes depressions showing no obvious fronts sometimes develop within a large mass of polar air – there seems little doubt that their formation is associated with the development of vertical instability. They are roughly circular in shape and, except for their manner of formation, they are not dissimilar from old occluded frontal depressions. Polar-air depressions sometimes develop secondary cold fronts as a result of frontogenesis but there is no warm sector. Precipitation is often limited to local showers but when there is a greater degree of instability and horizontal convergence, thunderstorms and areas of continuous rain (or snow) may develop. Sometimes when the polar front lies near the equatorial side of a polar-air depression a wave is induced on it and as the wave develops some of the warm air is carried into the circulation of the depression which may thereby become greatly invigorated.

Lows over inland waters in winter

The temperature of inland waters in winter is relatively high compared with that of the surrounding land mass. Almost invariably when a current of polar air spreads over large inland seas such as the Mediterranean, the Black Sea and the Great Lakes of North America, the air becomes unstable and local depressions may develop. The arrival of the cold air is marked by a cold front but the depressions often persist long after the whole region has been flooded with polar air and continue to give local showers and squalls. Winds circulate in the ordinary cyclonic manner but there are often large deviations caused by orographic and katabatic effects on the coasts; local gales and calms are typical of the coastal areas in these conditions.

Shallow lows over land in summer

When the pressure gradient is slight, surface heating over land in summer may lead to the formation of shallow depressions. Sometimes these are of little significance but when associated with vertical instability or with pre-existing fronts they may result in deteriorations such as squalls, widespread rain or outbreaks of thunderstorms. Central and western Europe are especially liable to these disturbances which are often associated with widespread thunderstorms.

Tropical revolving storms

These are discussed in Section 118.

116. OROGRAPHIC DEPRESSIONS

When a current of air meets a mountain barrier at a sufficiently large angle, energy is usually required if the air is to rise over the obstruction and there is a tendency, often very marked, for much of the air to sweep round the ends of the barrier, so avoiding the ascent. Something in the nature of a lee eddy is formed but the problem cannot be treated in quite the same way as that of the eddy formed by a building; for, when the obstruction is a large mountain range, the dimensions of the disturbance are such that the earth's rotation exercises an influence and the winds obey Buys Ballot's law. The result is a shallow depression to the lee of the mountains and a ridge of high pressure on the windward side with the sea-level isobars and winds distributed somewhat as in Fig. 93.

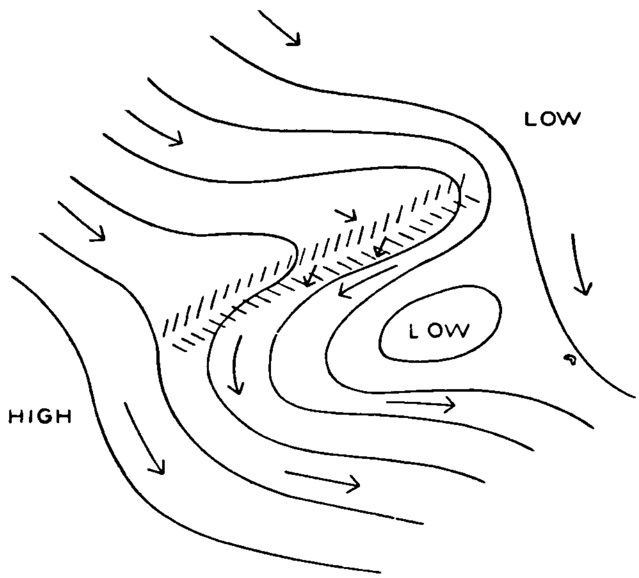


FIG. 93. *An orographic low*

The effect of a range of mountains on isobars and winds

Although the term may not meet with the approval of the theoretical meteorologist the depression is usefully looked upon as a 'partial vacuum'. The low may be maintained as long as the wind current persists but, apart from the local strong winds which blow round the ends of the range, the only disturbance in weather may be cloud and rain of orographic origin. Within the depression itself weather may be fine and warm as a result of the föhn effect as air passes over the barrier. An important case arises when a cold front reaches a mountain range after being more or less parallel to it and when there is warm stagnant air on the lee side. The lower part of the front is obstructed by the barrier with the result that the slope of the front becomes steeper and a lee depression begins to form. The upper part of the front may then pass over the crest of the range and overrun the warm air below. This results in severe instability with heavy showers and thunderstorms until the warm air is finally displaced by cold air sweeping over the barrier. Alternatively if the cold air can sweep round the ends of the barrier, a bulge is formed on the front with the air in the lee of the range acting as a warm sector and an intense depression may then develop.

A large proportion of the disturbances which occur over northern Italy are formed in association with a cold front held up by the Alps – the associated strong wind down the Rhône valley is the well-known 'mistral'. Similar developments are recognized in various other parts of the world, for example the depressions which form in the lee of the Scandinavian mountains when winds generally are from the west.

The barrier effect of mountain ranges is important in controlling and modifying the movements of all winds in their vicinity; moreover the meteorology of high-level or mountainous districts presents many special problems which cannot be discussed in this book.

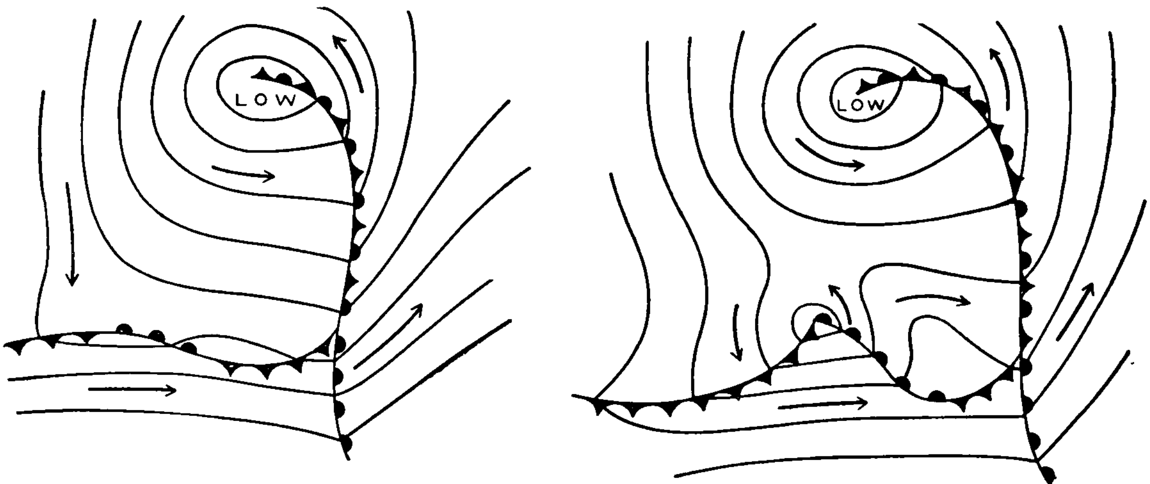
117. SECONDARY DEPRESSIONS

When a relatively small depression is enclosed within the circulation of a larger or primary depression, it is called a 'secondary'. Any process which leads to the formation of a depression may be responsible for the formation of a secondary. Generally speaking a secondary moves round the primary in a cyclonic sense, being carried along by the larger circulation. It is impossible to give in brief any adequate account of the variety of conditions which may lead to the formation of secondaries but some of the more common types are described below.

Frontal secondaries

A common place of origin of a secondary depression is the cold front of a primary where it trails back from the end of the occlusion. If the new disturbance forms well outside the primary circulation it forms the next member of the family (see Section 111), but when it moves rapidly along the front it may appear as a secondary to the previous member. A development of this type is illustrated in Fig. 94.

In (*a*) the new centre is shown as a small wave disturbance on the cold front but in (*b*) it has moved along the front and deepened to appear as a secondary. It not infrequently happens that the new formation becomes deep while the old occluded centre fills up and the secondary then becomes the dominant and controlling depression of the two. An old primary which becomes involved in the circulation of a new vigorous development behaves as any other secondary and often rotates round the new centre in a cyclonic direction. When the two centres



(a) (b)

FIG. 94. *Development of a frontal secondary depression*

are of similar depth they tend each to rotate round the other, the combined effect being a rotation round the col between them as indicated in Fig. 95.

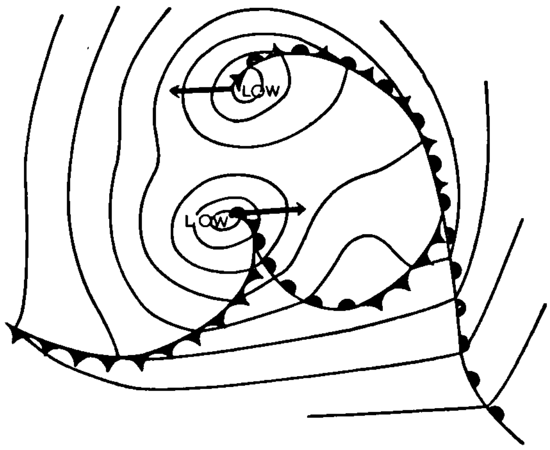


FIG. 95. *Movement of two neighbouring depressions*

Sometimes small disturbances form on the cold front well within the primary circulation, as in Fig. 96(*a*). Such a development may be recognizable only by a local widening of the isobars near the front. It moves along the front, but usually without much development and soon becomes absorbed again in the primary circulation. Although apparently of minor significance, such a disturbance often delays or even reverses the movement of a slow-moving cold front. The accurate timing of the passage of the front across (for example) an aerodrome is then a matter of considerable difficulty.

A third position favourable for the formation of a secondary is the tip of the warm sector of a depression which is already partly occluded. The remaining warm air acts as a warm sector for the secondary which develops while the primary centre continues to fill up. The development is illustrated in Fig. 96(*b*). Experience shows that this development is most probable when the primary depression and occluded front are held up by a mountain barrier, for example in southern Greenland or Norway; in either case the secondary develops and moves away eastwards to the south of the land mass. Secondary frontal depressions may also form on a secondary cold front behind the primary.

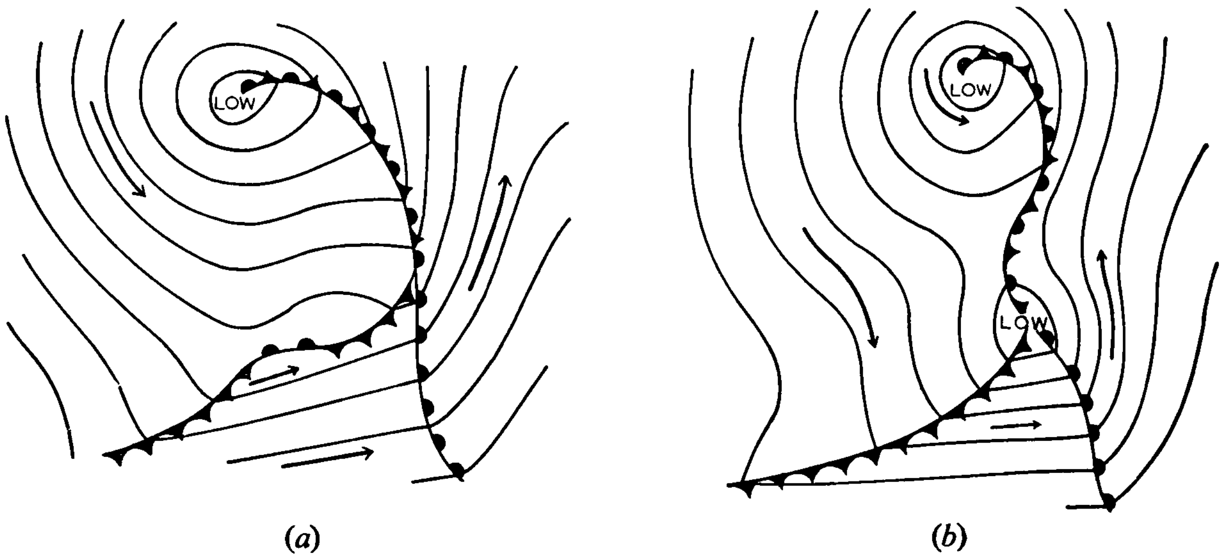


FIG. 96.(*a*) *Secondary depression well within the primary circulation.* (*b*) *Secondary depression at the tip of the warm sector of a partly occluded depression*

Other types of secondary

The processes of surface heating and instability and the disturbances due to orographical features are each liable to produce secondary centres. On almost all occasions when the circulation of a large depression passes across both land and sea or over mountain barriers, secondaries are formed. Thermal secondaries are most common in polar air when the winds are not very strong. For example, in summer, with light southerly winds circulating round an occluded depression off north-west Ireland, secondaries form over France and later affect south-east England.

Winds and weather in secondary depressions

It is a general rule that secondaries are likely to form within a large occluded or inactive primary which may itself be responsible for little or no bad weather. Except where the secondary is definitely frontal, it is impossible to generalize about the associated weather although there is almost invariably some local deterioration; many secondaries are responsible for much low cloud and copious rainfall. Shallow secondaries over the land in summer are particularly likely to give thunderstorms. In dry climates they may occur with no rainfall, for example those caused by heating over the African coasts of the Mediterranean when a depression lies to the north, but may be responsible for duststorms. The winds in the path of a secondary are subject to rapid variation as the centre moves over and, when the low is intense, strong winds or gales may occur. The strongest winds are generally to be found on the side remote from the primary centre, where the circulation of the secondary intensifies that of the primary; calms and variable winds occur in the col between the two centres. When the secondary is small its wind circulation is generally quite shallow and has little effect on the winds at heights above 10 000 feet, these being determined by the primary circulation.

118. TROPICAL REVOLVING STORMS

Intense depressions occur in some tropical regions during certain periods of the year. They are known as cyclones, typhoons or hurricanes according to locality.

The lowest pressure is frequently about 960 millibars but is often much less; one record from the Philippines gave a pressure of 887 millibars. Often the depth is as great as that of a deep extratropical depression but, whereas the latter may measure as much as 2000 miles across, the isobars of the tropical cyclone are crowded into a diameter generally less than 500 miles, or even less than 100 miles in its early stages. With such pressure gradients the winds regularly reach hurricane force. Weather is violently disturbed with torrential rain occasionally accompanied by thunder and lightning. After formation of the storm, a central region develops, a few miles in diameter, where the sky is broken or cloudless and the wind usually (but not invariably) calm; this is the 'eye' of the storm and there the air is subsiding. The storms always originate over the sea and gradually increase in area but if they move over land they quickly degenerate and are usually destructive only near the coasts. The rate of movement of the centre is generally less than 15 knots. If the storms move outside the tropics they enlarge further and subsequently behave like extra-tropical depressions, perhaps acquiring a frontal structure. Occasionally depressions reaching the British Isles have been traced back to a tropical origin.

Details of the localities and tracks and the seasons most liable to disturbance are given in Section 159.

Theory of tropical revolving storms

These storms most frequently originate near the intertropical front (see Section 159) and appear to result from the marked instability which is often characteristic of this region. Much of the energy is derived from the latent heat set free by condensation of water vapour; in this connection it is pertinent to note that the storms occur mainly over the western parts of the tropical oceans where the trade winds have had a long passage over the sea, and often in areas where the air on the equatorward side of the intertropical front has crossed over from the other hemisphere and has become almost saturated. Once a storm has formed, the circulatory velocity is so great that no frontal structure can persist and it becomes an almost symmetrical circular depression.

The mechanism by which the energy of latent heat released by condensation is converted to intense circulatory motion appears to be mainly that already described for thermal depressions but it cannot be claimed that the process is as yet fully understood. One factor contributing to the intensity of the storm is undoubtedly the high humidity and degree of instability present beforehand; another factor is the latitude, since for a given pressure gradient the strength of the associated winds increases as the equator is approached (Section 24). However, no storms are found within five degrees of the equator, the explanation being that in those latitudes the deviating force of the earth's rotation is too small to produce any circulation. Over land, tropical cyclones do not form even when conditions are extremely unstable. It seems that most of the inflow to a storm takes place near the bottom and that the amount of water vapour so taken in over the land is likely to be less than over the sea where the air currents are very moist. Possibly an additional factor is the increased surface friction over the land; this not only reduces the wind speed as compared with the value it would have over the sea but thereby also further reduces the input of water vapour.

Flying conditions in relation to tropical revolving storms

Tropical revolving storms have been successfully traversed by aircraft on many occasions – indeed, they are flown through regularly on certain meteorological reconnaissance flights. These flights have shown that conditions within the storms up to at least 10 000 feet are extremely unpleasant and even hazardous. In general the only reasonable course is to avoid the storms as far as possible but if for some particular reason, such as reconnaissance, it is necessary to enter a storm, this should be done only after the fullest pre-flight preparation. The localities and the seasons of the year subject to the disturbances are well recognized and any pilot contemplating a flight in an area likely to be affected should make a study of their local characteristics so that he may recognize the signs of their formation. In most parts of the world navigational warnings are adequate to ensure that no aircraft in communication with the ground need be caught unawares. The position and probable movement of the centre are determined as closely as possible, in some areas with the help of reconnaissance aircraft, and this information is widely distributed. It is however necessary to anticipate the possibly of an aircraft out of touch with base, or the occurrence of a cyclone without previous warning having been received, so that a careful watch should be kept for any conditions likely to indicate the existence of a storm. The following notes describe the premonitory indications both at the ground and aloft and the flying conditions within the storms.

Ocean swell. The strong winds of a cyclone often set up a heavy swell which spreads out far from the storm centre. When seen this should serve as a warning but the absence of swell should not be taken to indicate that no storm exists in the area.

Surface pressure. The characteristic diurnal variation of pressure in the tropics shows up clearly on a station or ship's barograph as a double wave with maxima at 1000 and 2200 local time and minima at 0400 and 1600 (Section 6). Any tendency to a general fall of pressure, or even a departure from the regular oscillation, should be looked on as a likely indication of an approaching or developing storm.

Use of altimeters. The pressure altimeter is of little use in maintaining constant altitude in the immediate neighbourhood of a cyclone because the rapid fall of pressure towards the centre causes serious over-reading. However, if the pressure altimeter is used in conjunction with the radio altimeter, it is possible to find out whether the aircraft is moving towards or away from the centre since both the pressure at a given height, and the height at a given pressure altitude, decrease towards the centre. This decrease is gradual in the outer parts of the storm but becomes more rapid as the centre is approached.

Wind. Since a cyclone develops in a region where winds are light any marked increase of wind may be taken as a reliable warning, while the direction of the centre may be inferred from that of the wind by the usual rule. If a cyclonic circulation of strong winds is observed, then the existence of a cyclone is certainly established even though its subsequent intensification remains uncertain.

Cloud and precipitation. The instability conditions usually associated with the intertropical front become intensified in the earliest stage of the development of a cyclone, when compact masses of convective cloud are present with heavy showers and thunderstorms. After the cyclonic circulation has started, the cloud formations tend to become arranged in bands along the wind and more or less concentric with the centre but spiralling inwards so that the central area of the storm (outside the eye) forms an extensive unbroken cloud mass. Subsidence takes place between these cloud bands in the outer part of the storm, as well as in the eye. The vertical extent of the clouds appears to be greater over the inner core of the storm – apart from the eye, which may be cloudless – and commonly reaches to cirrus levels. If flight is above cloud, the approach to a storm may be indicated only by thick bands of cirrus, but while the existence of a storm may on occasions be revealed by the upper or lower cloud formations, these are apt to be misleading as a guide to the position of the centre. The cloud base is often about 1000 feet and occasionally on the surface in heavy rain. Visibility below the cloud base is poor in strong winds on account of spray.

Use of radar. Airborne radar is a valuable help to the detection of a cyclone, especially at night when visual indications are lacking. The spiralling cloud system can be easily recognized on the radar screen (Plate XXXI) at distances of some 30 to 40 miles and the position of the eye accurately fixed. A disadvantage is that this method gives little or no indication of the intensity of the storm.

Turbulence. This occurs in two ways. Severe turbulence is present in the active convective clouds within the storm and in low-level flight it can often be avoided by flying below the clouds. The other type is the frictional turbulence generated in the lowest layers by the strong winds. This of course increases in intensity with the strength of the wind; it becomes severe when the wind exceeds 50 knots and is considered dangerous if the wind exceeds 80 knots. Low-level frictional turbulence can best be avoided by increasing the flight altitude to about 5000 feet and experience shows that this is usually possible, in these regions of strongest wind, without entering the clouds. Strong, large-scale, updraughts and downdraughts may also be encountered in various parts of the storm.

Vertical extent. When a new cyclone forms, the circulation of wind probably comes into existence first of all in the lowest layers but the height affected increases rapidly during the first day or two, sometimes reaching 30 000 feet and possibly exceeding 50 000 feet. Most of the penetrations hitherto made by reconnaissance aircraft have taken place at about 1500 feet since accurate navigation requires the sea to be kept in view most of the time so that the wind may be determined at frequent intervals. If a flight is continuously in cloud, navigation becomes uncertain because of the absence of precise wind information. In fact the greatest danger in flying blind through these storms lies in getting lost because of a combination of the strong variable winds and a failure—due to 'static'—of radio aids. It cannot be said to what height it would be necessary to ascend in order to fly 'above the weather', but the minimum safe height no doubt tends to increase with the age of the storm. On some occasions the sky has been found clear at 10 000 feet but that is exceptional.

Avoidance of a storm while in flight

When the pilot of an aircraft realizes that he is flying in the vicinity of a cyclone, there are various courses which may be adopted according to circumstances and much must be left to the navigator's discretion. A change to a reciprocal course before conditions become seriously adverse may be advisable—this will ensure flying out of the storm and a landing may be effected at any accessible base. If circumstances permit of reliable wind measurement the storm may be negotiated by an intelligent application of Buys Ballot's law. On whatever compass course the storm is approaching strong winds from the left, or port (in the northern hemisphere), indicate that the centre lies somewhere ahead; if therefore course is altered until the wind is from starboard the centre will be avoided. As the centre always moves quite slowly it may be regarded as stationary to an aircraft flying in the vicinity. The following three courses of action may be considered when conditions indicate the proximity of a revolving storm.

(i) If a strong wind springs up on the port bow change course to the left sufficiently to put the wind on the starboard bow and continue until conditions improve before returning to the proper course (course AA of Fig. 97(*a*)).

(ii) If a strong wind springs up on the port quarter change course to the right sufficiently to put the wind on the starboard quarter and proceed as above (course BB of Fig. 97(*a*)).

(iii) If a strong wind springs up on the port beam take advantage of the following winds by turning to the right so as to put the wind on the starboard quarter as an (ii) (course CC of Fig. 97(*a*)).

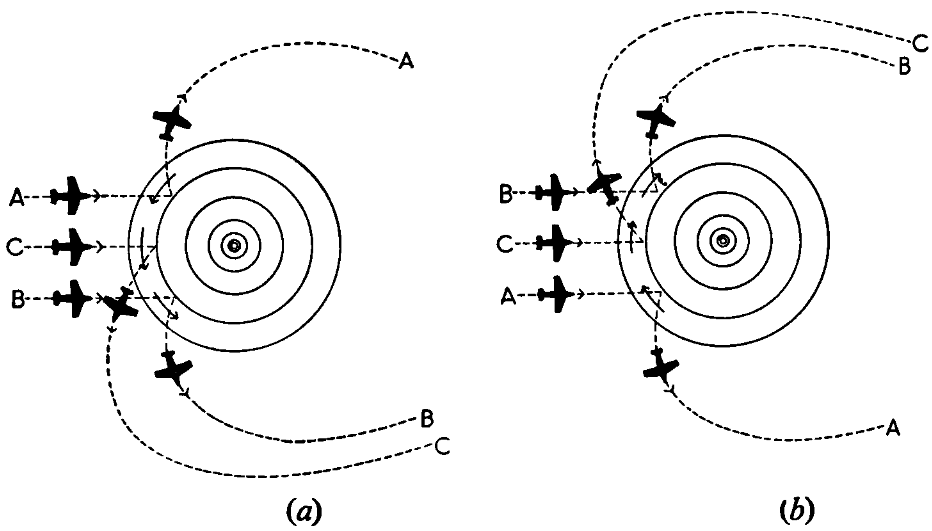


FIG. 97. *Avoiding the storm* (*a*) *northern hemisphere* (*b*) *southern hemisphere*

The broken lines show courses taken to avoid flying near the centre of the storm.

In the southern hemisphere the terms port and starboard, left and right, should be interchanged as illustrated in Fig. 97(*b*). The rules are based on the wind at a flying height of say 1000–2000 feet, which may be taken as circulating round the centre without in-draught. If the wind is estimated at the surface from the sea disturbance, the wind at flying height should be taken as veered from the surface wind in the northern hemisphere, backed in the southern hemisphere, by some 30°.

119. TORNADOES

The name tornado is applied to two distinct phenomena. In West Africa, particularly in the countries bordering the Gulf of Guinea, tornadoes are thunder squalls generally accompanied by heavy rain but occasionally without precipitation. They often advance across the country from east to west as a line squall and are most frequent at the beginning and end of the wet season, that is in March, April, May, October and November.

The other kind of tornado takes the form of a violent whirlwind and it is this type of disturbance which is signified by the tornado as generally understood. They occur most frequently in the United States of America east of the Rocky Mountains, especially in the central plains of the Mississippi region; in other parts of the world they are usually less intense. The whirl or vortex has a diameter of some hundreds of feet and so is much too small to be represented by a system of isobars on a weather map, although the central pressure is extremely low. It forms in conditions of great instability in association with a trough of low pressure, particularly when V-shaped, where the interaction of winds of widely different temperatures gives opportunity for violent vertical convection combined with cyclonic rotation. Tornadoes are therefore usually associated with thunderstorms, heavy rain and perhaps hail. The winds are of hurricane force and extremely destructive within the narrow track affected; moreover the large and rapid decrease of pressure experienced as a tornado passes overhead causes the air inside buildings to expand with an almost explosive effect.

120. WATERSPOUTS

The structure of a waterspout is similar to that of the tornado except that the occurrence takes place over the sea and the phenomenon does not reach the same intensity. It is associated with heavy instability cloud. In the incipient stage a funnel-shaped extension, point downwards, projects from the base of the cloud and at the same time the sea below becomes agitated by a whirling vortex of air, so forming a cloud of spray. Sometimes this stage is all that is seen, at other times the so-called 'cloud pendant' develops downwards until it forms an unbroken column of cloudy whirling air joining sea and cloud; this is the waterspout. The dimensions of the vortex are quite small; the figure of 20 or 30 feet is sometimes quoted for the diameter and 200 to 300 feet for the height, but the maximum possible dimensions are no doubt considerably greater. In the early stages the column is generally vertical but later it becomes distorted and breaks away at the base, after which it quickly disappears. The whole phenomenon rarely lasts more than half an hour. Waterspouts are most common in low latitudes although they are occasionally seen off the British coasts. They frequently occur in groups.

121. TROUGH OF LOW PRESSURE

All sharply V-shaped troughs are associated with fronts (cold, warm or occluded) and as such they have already been dealt with. Not infrequently however troughs appear with no front along them, in which case there is no sharp kink in the isobars. Being regions of convergence and rising air, troughs are often the scene of much bad weather, and over land in summer are particularly liable to thunderstorms. They are also likely places for the development of a secondary depression or a secondary cold front. A non-frontal trough is typical of the cold air mass behind an occluded depression and the increase of pressure gradient associated with it may be a source of renewed gales after a depression has passed.

While a front, and therefore a frontal trough, is displaced by the wind component normal to the front, the wind in a non-frontal trough tends to follow the curved isobars without producing displacement. The movement of this type of trough can be determined only from the pressure tendencies; it is directed towards the side of most rapid fall (or of slowest rise, if pressures are generally rising).

CHAPTER 18

ANTICYCLONES

122. TYPES OF ANTICYCLONE

A convenient classification divides anticyclones into two types. In the first the high pressure is accounted for by the low temperature and high density of the air in the surface levels and through the lower troposphere; it is known as a 'cold anticyclone'. In the second type or 'warm anticyclone', the air in the troposphere is warmer than the average; the high pressure then cannot be explained by the greater weight of the lower atmosphere and must be due to an excess of air at high levels.

A subdivision of each type may be made into 'permanent' (or 'quasi-permanent') and 'temporary' or migratory anticyclones. A permanent anticyclone constitutes a persistent feature of a given area over a period of months. For example the subtropical belts of high pressure are regarded as permanent features even though in places they may be occasionally interrupted or displaced by low-pressure systems; similarly the Siberian high is a more or less permanent feature throughout the winter months. On the other hand the temporary or migratory highs are transient features which often pass over any one place within a day; occasionally one remains stationary for several days or even weeks, but then it is usually found to form an extension of a permanent high.

123. GENERAL PROPERTIES OF ANTICYCLONES

An anticyclone is a region of high pressure with the winds circulating in the direction given by Buys Ballot's law. It may be shown on theoretical grounds that the wind speeds over the central regions must in general be weak compared with those possible in a depression. An anticyclone is, then, a region of light winds although there is no reason why on the outskirts, away from the centre, winds should not be strong.

When surface pressure is rising within a mass of air, divergence is set up and the air must subside (Section 27). The development of an anticyclone therefore involves subsidence and even when the system has reached its maximum development there is still slight outflow near the surface due to surface friction (which causes the surface winds to drift across the isobars from high pressure to low) and slight subsidence continues. As a necessary condition for rain is ascending motion, it follows that an anticyclonic region is generally dry.

In broad terms an anticyclone is, then, a region of quiet weather.

124. COLD ANTICYCLONES

Permanent or quasi-permanent cold anticyclones

Maps of average pressure over a long period show weak anticyclones over the polar regions but daily charts show that these regions are frequently invaded by travelling depressions so that the high pressure is not a permanent feature. Nor are the polar anticyclones, when they do form, necessarily of the 'cold' type – they can also be of the 'warm' type.

Perhaps the only example of a permanent cold anticyclone is that over Siberia in the winter, and even this is not immune from disturbances. There is, however, a predominating tendency for high pressure to be maintained throughout the season and the same is true to a less extent over the North American continent. In the Siberian high the air subsides, surface cooling more than compensates the adiabatic warming and the air in the lower atmosphere is maintained at a very low temperature with a marked surface inversion. There is no precipitation except when the anticyclone is temporarily displaced by travelling depressions and fronts, but conditions are very favourable for fog or low stratus when the moisture content of the air is adequate. In the central region the air is generally too dry for fog and brilliant frosty weather is common, but in the outer regions, particularly near the sea coasts, dull foggy weather is common. Flying conditions, apart from the extreme cold, are good; even when low cloud and fog obscure the ground, clear skies and smooth air may be found at a height of a few thousand feet.

These seasonal anticyclones play an important part in controlling the general wind circulation of the atmosphere, which is described in Part V. The area affected by the circulation varies from time to time. For example, the British Isles occasionally come under the influence of the Siberian anticyclone towards the end of the winter. The wind direction is then easterly and either the clear and frosty or the dull foggy type of weather may occur. Frequently the air leaves the continent clear and dry but picks up considerable moisture over the sea; as the sea is warmer than the air, fog formation is prevented but a layer of stratus or stratocumulus may be formed. Often in south-east England the easterly winds have not picked up sufficient moisture in their short sea-passage, particularly if the wind is fresh or strong, and the weather remains bright; further north, however, easterly winds in winter almost invariably bring cloudy skies to the east coasts – heating over the sea may even be sufficient to give local instability showers of snow or sleet.

Temporary cold anticyclones

Cold anticyclones of a temporary nature are common features of the changeable weather of middle latitudes. When a family of frontal depressions travels eastwards, each member is necessarily separated from its successor by a ridge of high pressure or a small anticyclone in the cold air which moves along between the two centres. The ridge brings a short break of dry weather lasting perhaps 24 hours between the instability showers behind one depression and the warm-front cloud and rain of the next; it is a transitory pressure feature which usually collapses as the new depression advances. When polar air finally breaks through and terminates the family (see Section 111), the cold air builds up into an anticyclone of considerable size. Such developments may occur anywhere in temperate latitudes; in winter over the land the anticyclone may merge into, or become an extension of, the seasonal continental anticyclone, being maintained as a cold anticyclone by continuous radiative cooling at the surface. Over the sea, and over the land in summer, a cold anticyclone is never of great persistence; it either collapses within a day or two or is gradually transformed into a warm anticyclone by the adiabatic heating associated with subsidence.

The type of weather depends on many factors. In the summer, polar air is subject to surface heating over both land and sea. Fog is, therefore, exceptional and clouds of a cumulus type may develop. In the winter the air may undergo surface cooling over the land and radiation fog becomes probable where the winds are light, and stratus or stratocumulus clouds where the wind is stronger. The cold anticyclone is, however, associated with a large measure of bright and dry weather. In accordance with the general rule that a region of low temperature tends to become a region of low pressure in the upper atmosphere (see Section 7), the cold anticyclone is not deep and the easterly winds on its equatorward side are generally shallow, decreasing with height and changing to westerly at moderate altitudes.

125. WARM ANTICYCLONES

Permanent warm anticyclones

The subtropical oceanic belts of persistent high pressure are areas of subsiding air which does not undergo any appreciable surface cooling. They are subject to a seasonal variation in position, moving north and south with the sun, and in the winter season they extend over the land areas where they are partly intensified by some degree of surface cooling. They are also subject to temporary variations in sympathy with the development and movement of depressions and anticyclones in higher latitudes; and lastly they are liable to be invaded in certain localities by tropical revolving storms originating nearer the equator. In general, however, they are remarkably stable systems, being composed of warm and dry subsided air; the weather is fine with usually excellent visibility. These anticyclonic areas are the main source regions of tropical maritime air masses which give much low cloud, fog and drizzle on moving to higher latitudes.

Temporary warm anticyclones

In temperate latitudes, warm anticyclones are a temporary phenomenon and it is possible to distinguish two varieties:

Extension from the subtropical high. The air mass in this type of anticyclone is tropical. In summer it gives fine warm weather over the land but in autumn, whenever the air is of maritime origin, radiation fog at night becomes almost certain where the winds are light. Sea fog may occur at any season but especially in spring and summer. Above the fog or low cloud there is invariably an inversion of temperature with a clear upper sky. This type of anticyclone seldom affects temperate regions in winter since the subtropical high itself is then shifted towards the equator, but in other seasons the anticyclone when once established may persist for periods up to a few weeks.

Development from a temporary cold anticyclone. The most common type of temporary warm anticyclone is evolved from a temporary cold anticyclone by the adiabatic warming which results from subsidence. When one develops over the land in summer the air becomes dry and the result is bright dry weather. Over the sea the air absorbs moisture by evaporation while an inversion of temperature builds up as the result of subsidence. Fog may then develop over the western side where the air is moving towards lower sea temperatures and also over the land as a result of cooling by radiation. Alternatively, a layer of stratocumulus often forms within the inversion and if the air moves from the sea over land during the summer, day-time insolation may be sufficient to break the cloud into small cumulus or even to clear it completely. Long spells of settled weather are often associated with this type of anticyclone.

126. RIDGE OF HIGH PRESSURE

A ridge may extend in any direction outwards from an anticyclone and may be almost stationary or fast moving. The only reliable indication of the motion of a ridge is the barometric tendencies; the direction of motion is, of course, towards the rising pressures and away from the falling pressures. A ridge is generally a passive feature and when it projects between two depressions its motion is controlled by that of the depressions. The weather is generally good and particularly fine and clear in the ridge behind a frontal depression, a consequence of subsidence of the polar air.

127. COL

The col is not an entity in the same way as a depression or anticyclone – the winds do not circulate round it but flow towards and away from it. If as in Fig. 98 a quadrilateral is drawn centred at the col, the air flows in across two opposite sides

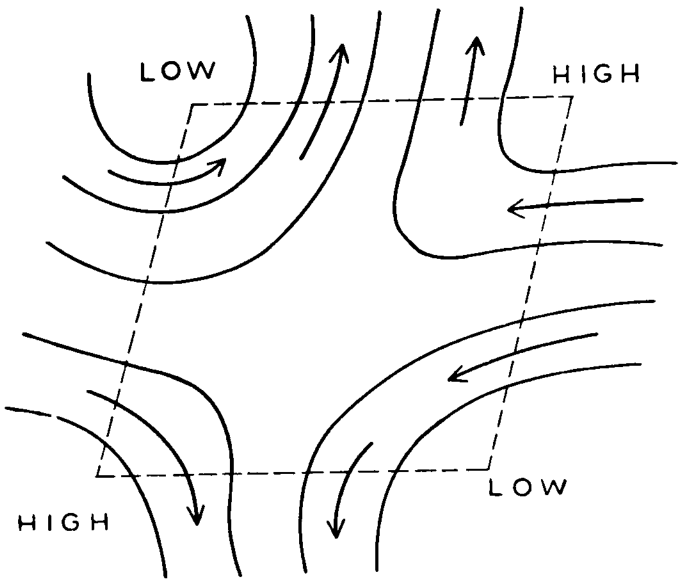


FIG. 98. *Isobars and geostrophic winds in a col*

and out across the others, remaining stagnant in the centre. The associated weather is very much a matter of circumstances and almost any phenomena can occur except for strong winds at low levels. This does not of course mean that the probable weather cannot be anticipated but only that each col needs to be considered on its merits. A col is frequently traversed by a front, particularly when one of the two lows is secondary to the other, and as the front is more or less stationary any rain from it may be long continued. If, however, pressure is rising in the col (a frequent occurrence) the front will usually degenerate. On the other hand if pressure begins to fall the inflowing currents from opposite directions may tend to develop a new front or intensify an old inactive one, and the convergence may then cause rain. In autumn and winter over the land the col is associated with low cloud and fog; in summer thunderstorms are frequent although the likelihood of their occurrence on any particular occasion can be determined only by studying the upper air observations.

CHAPTER 19

ELEMENTS OF FORECASTING

128. INTRODUCTION

It is not possible within the limits of one chapter to give more than a very general outline of forecasting methods. Indeed, since the synoptic chart is the main tool of the scientific forecaster, no course or book of instruction on practical forecasting can claim to be comprehensive unless it is accompanied by reproductions of numerous actual charts. That good treatises on practical forecasting are so few is mainly because of the difficulty in providing, either within the text of the work or in folder or album form, a sufficiently large number and variety of charts bearing all the necessary observations and analyses. For the purposes of the present work Chapter 20 has been devoted to the presentation of small-scale reproductions of surface and upper air charts for several situations and these enable a number of points of general and particular interest to be studied. No more than this can be done within the compass of the present volume, but it should be sufficient to demonstrate how weather charts are constructed and used, and to illustrate some of the points discussed in the present chapter. Only by long experience and close examination of successive weather charts and particularly of current ones, which can be related by the forecaster to the existing weather in his own area, can ability be acquired in applying to actual weather forecasting the physical principles and other meteorological information which have been discussed in this book.

However, a pilot or navigator is not required to become an accomplished forecaster, though it has long been recognized that he should have sufficient acquaintance with synoptic charts and with the methods of aviation forecasting to enable him to derive the maximum benefit from meteorological briefings by professional forecasters; moreover in cases of necessity or emergency he himself should be able to make good forecasts during flight on the basis of all available information. The syllabuses for the technical examinations for civil pilots' and navigators' licences and for corresponding qualifications in the Royal Air Force have accordingly provided for these requirements. Further, by international agreement through the International Civil Aviation Organization (ICAO) regarding the supply of detailed meteorological information for long-distance flights, it is necessary that pilots and navigators should be competent in using the various aids provided. These aids include the supply of actual and forecast charts for the surface and certain upper levels, together with forms giving information about the weather distribution along the route to be flown.

It has been the aim of ICAO to secure the maximum possible uniformity of practices in the meteorological services of the world in regard to facilities provided for aviation, so that pilots and navigators may readily understand the flight briefing and documentation which they receive in any country that is a member of ICAO. Some differences in procedure between meteorological services in equatorial regions and temperate latitudes must remain, however, particularly in regard to the attention paid to frontal analysis on surface charts and to modes of displaying the distribution of winds at high levels. Variations there will also be in the methods of forecasting; in some regions seasonal and diurnal effects must be given particularly careful consideration, whilst elsewhere the trends of movement and development of pressure systems will figure prominently in the technique of forecasting. Nevertheless, in what follows in this chapter the main emphasis will be on practices and procedures in temperate latitudes in general and in the British Meteorological Office in particular.

129. ANALYSIS OF THE SURFACE CHART

By analysis is meant the dual process of locating the fronts and drawing the isobars. Before either of these steps can be taken it is essential for the analyst to be so familiar with the symbolic method of plotting the observations that the meaning and significance of each entry is at once evident except perhaps for a few symbols of infrequent or rare occurrence. It is also valuable if he is readily able to discern any reports which are clearly erroneous, since errors, although infrequent, may arise at any stage from the making of the original observations to the final decoding and plotting. In addition to the obvious errors there are the reports which appear inconsistent with neighbouring reports or with previous observations from the same station. On these suspected reports an open mind may have to be maintained until a confirmation can be obtained from the station originating the report or until the report can be compared with a subsequent one which may enable its accuracy to be decided.

The best procedure to follow in analysing a chart varies with circumstances. An experienced forecaster working against time and being interrupted by requests for forecasts is normally handicapped in following a systematic method and in giving concentrated attention to every detail. The pilot or navigator who is learning analysis mainly for examination purposes should accustom himself to a sequence on the following lines:

(i) If possible, obtain an analysed chart for the previous hour of observation. For an anlysis to be satisfactory it must be consistent with that of the preceding chart. This is the fundamental principle of 'historical sequence'. It is usually an advantage if the previous movement of depressions and anticyclones has been indicated on the earlier chart. This is commonly done by marking the positions of the centres of the systems at the 6-hourly chart intervals and joining the successive positions by a broken line. From a brief inspection of the resulting tracks a preliminary idea of the main features of the synoptic situation is gained.

(ii) If an illuminated tracing or 'light' table is available, place the chart to be analysed on top of the previous chart, ensure correct registration and lay the two on the light table. By extrapolation from the tracks shown on the previous chart deduce approximate positions for the centres of the depressions and anticyclones and mark them lightly on the top chart. Also estimate and lightly mark approximate positions for the main fronts; these too must follow on logically from the previous chart, their displacement being estimated by means of the geostrophic rule (Section 104). Should no previous chart be available, the positions of fronts and pressure centres must be determined solely by a careful scrutiny of the plotted observations.

(iii) Sketch in the isobars roughly and lightly, starting where the analysis is simplest. In general it is easier to draw isobars in areas where winds are strong than where they are light and variable. It is often a good plan to begin by inserting the isobars for the lowest pressure in the depressions and the highest pressure in the anticyclones.

(iv) Insert the fronts as accurately as possible by making any necessary adjustments to the estimated positions with the aid of the plotted reports, applying the knowledge about frontal and air mass characteristics given in earlier chapters. From the weather conditions at the front, the characteristics of the air masses on the two sides of the front as shown by the plotted reports and from the previous history (if available), decide on the type of front (warm, cold, occlusion).

(v) Revise the roughly drawn isobars, carefully fitting the pressure and wind observations and making appropriate kinks where they cross a front, as for example in Fig. 103 (page 262).

(vi) Finally mark the fronts in the standard colours or by the conventional markings. The standard colours are as follows:

warm front: red
occlusion: purple (or contiguous red and blue)
cold front: blue
stationary front: broken line alternating red and blue.

The conventional markings can be seen on Fig. 107 (page 281); the projections are placed on the side of the front towards which it is moving, but for a stationary front they alternate from side to side.

Drawing of isobars

Isobars are normally drawn for every one, two or four millibars according to the scale of the chart; in all cases the isobar of 1000 millibars is drawn. If then, for example, the lowest pressure reported in a depression is 989 millibars, the central isobar of the system to be drawn on a chart with isobars at two-millibar intervals will be that for 990 millibars, whilst that on a chart with isobars at four-millibar intervals will be 992 millibars. By inserting the intermediate isobars between the highest and lowest pressures whether at one, two or four-millibar intervals, a start may be made anywhere on the chart and the isobar drawn continuously in such a way that the pressures at stations near the isobar on one side of the line are always above and on the other side always below that appropriate to the isobar itself; for example the 1020-millibar isobar will pass between two stations if one has a pressure of 1018·7 millibars and the other 1020·9 millibars and will be rather nearer the latter than the former. When stations are far apart and there is a substantial difference of pressure it is useful to determine how many isobars lie between them and then to divide the distance between them so as to space the isobars out more or less uniformly. If however the wind force at one of the stations is considerably greater than at the other the spacing of the isobars should be correspondingly closer near the station with the stronger wind than near the other. The following fundamental properties of isobars should be borne in mind:

(i) Isobars are simple curved lines with loose ends at the edges of the chart only, or simple closed curves.

(ii) Isobars never cross, touch or join (except where two parts of the same isobar join to make a closed curve).

(iii) Everywhere along an isobar the higher pressures are on one side and the lower pressures on the other, and the sides must never be interchanged on passing along the isobar. For example, in Fig. 99 the isobar for 1012 millibars must not be drawn either as AB or CD, for in either case the sides of high and low pressure are interchanged on passing along the line. The isobar must therefore be in two parts, as shown by the full lines. These mark a col – it is in such a region that the beginner is most likely to get into difficulties.

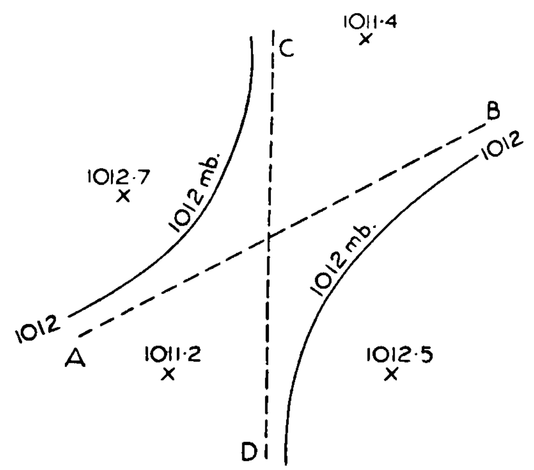


FIG. 99. *Drawing isobars in a col*

(iv) The pressures on consecutive isobars always differ by the same standard interval except at a col where they have the same value in the direction across the col.

(v) Isobars should be drawn to fit the wind direction as nearly as possible, in accordance with Buys Ballot's law, allowing for a slight flow across the isobars from high pressure to low. There should, however, be no slavish insistence on this point for there are many exceptions to the rule particularly where winds are light and in coastal and hilly districts. Isobars must also be drawn in general conformity with the wind strength, being near together where the winds are strong and well separated where they are light. In regions of somewhat sparse observations where a good deal of interpolation may be necessary, it is important in passing from an area of strong winds and concentrated isobars to one of light winds and few isobars that the isobars should be drawn to show a gradual transition and should not be unduly crowded in the one region at the expense of the other.

(vi) When the isobars are finally adjusted they should be as smooth as possible with sharp bends only at frontal troughs and possibly over mountain ranges. Small irregularities in the isobars should be avoided unless it is certain that they are caused by secondary disturbances. It is often better to assume a slight error in a pressure observation and a local cause for an irregular wind (such as a sea-breeze or a mountain or valley wind) than to fit one such single observation by drawing a dubious trough in the isobars which could be mistaken for a significant secondary disturbance. Experience normally enables practised forecasters to distinguish the real from the local or spurious irregularity.

This can be said to complete the usual analysis of the chart, but on occasions when pressure is falling or rising at a substantial rate it may be convenient to draw isallobars (lines joining places having the same barometric tendency) since these show clearly the areas where the main changes of pressure and movements of pressure systems are taking place. A concentrated area of falling pressure appears as a system of closed curves similar to the isobars of a depression and is known as an isallobaric low. Likewise an area of rising pressure is known as an isallobaric high.

130. ANALYSIS OF UPPER AIR CHARTS

No study of the weather situation can be complete unless it is three dimensional. The surface charts do in fact take some account of this aspect of the problem since the plotted reports include features of weather, for example, cloud and rain, which may originate far above the surface layers; but a detailed analysis of observations of pressure, temperature, humidity and wind in the upper air is an essential part of the diagnosis of any synoptic situation.

The most practical way of displaying the patterns of circulation in the free atmosphere is by the construction of contour charts for selected pressure levels. The basic theory of these charts has been dealt with in Section 26. In analysing these charts it is not sufficient to consider each selected level independently, but rather is it necessary to analyse the atmosphere layer by layer and through the superimposition of successive layers to ensure that the analyses of the selected levels are mutually consistent. The importance of doing this is related to the sparseness of upper air observations in space and time as compared with surface reports, so that the layer-by-layer process secures an accuracy at any given level which is substantially greater than would be achieved by analysing each level independently.

The centralization in Bracknell of upper air analysis by computer (see Section 133) has removed the necessity to make similar analyses manually at various forecast centres. But apart from the possibility that adequate computer output may not be readily available everywhere, a practising forecaster needs to appreciate the 'thickness' and 'gridding' techniques, and while details of the method of analysis may vary between individuals the general principles are the same throughout.

Standard pressure levels for which contour charts are normally drawn are 700, 500, 300, 200 and 100 millibars and the contours are usually drawn at 60-metre (200 feet) intervals. The first step is the drawing of a chart of contours of the 1000-millibar surface; these contours are nearly the same as the mean-sea-level isobars at 8-millibar intervals but depart somewhat from them by small amounts depending mainly on the temperature. An experienced analyst can draw the 1000-millibar contours directly from the isobars; they are traced from the surface chart on a 'light' table.

Next the chart for 700 millibars is analysed. Firstly the thickness lines of the layer 1000–700 millibars are drawn to fit the observed thicknesses and the thermal winds for the layer as plotted. The contours of the 700-millibar surface are then obtained by graphical addition of these thickness lines and the 1000-millibar contours which have previously been traced on to the chart. This process of addition is known as 'gridding' and is illustrated in Fig. 100 in which the set of curves A represents the 1000-millibar contours and curves T are thickness isopleths of the layer 1000–700 millibars. The 700-millibar contours, curves B, are obtained by addition. The addition is not, however, a purely mechanical process; the contours must be in as close agreement as possible with both the observed contour heights and the winds at 700 millibars as plotted on the chart, and the pattern of the contours must follow on consistently from that of the preceding chart. In adjusting

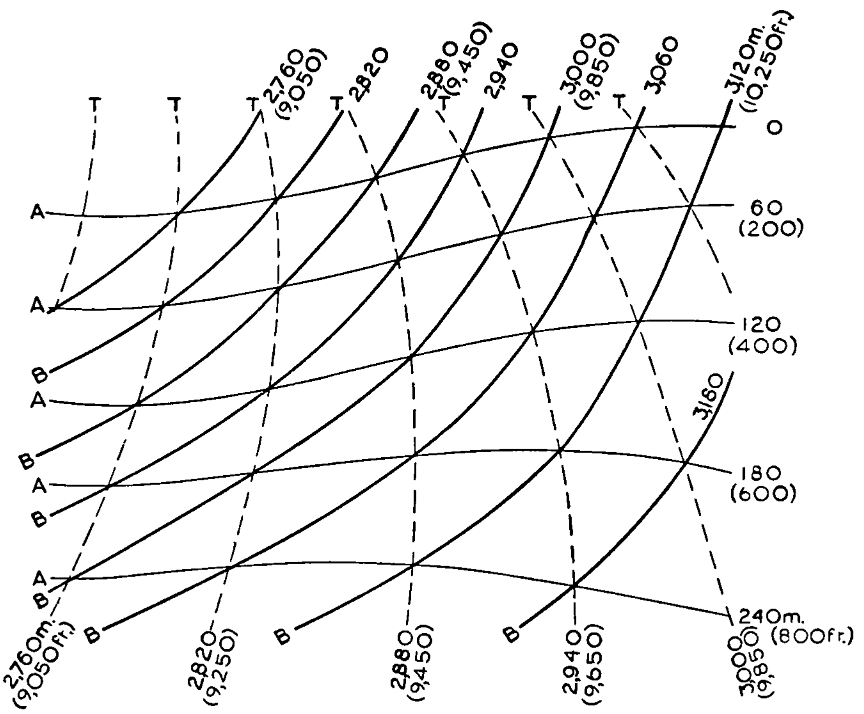


FIG. 100. *Construction of* 700-*millibar contours by gridding* 1000-*millibar contours with* 1000–700-*millibar thickness lines*

A ——— 1000-millibar contours B ——— 700-millibar contours
T – – – 1000–700-millibar thickness lines
At each intersection it will be seen that:
1000-millibar height + thickness = 700-millibar height

the contours to the observed winds it is usual to assume that the air movement is geostrophic, that is, in a direction parallel to the contours and with a speed inversely proportional to their distance apart. It is necessary, however, to be prepared for substantial departures from geostrophic flow particularly in regions where rapid pressure changes are taking place or where the contours are strongly curved and close together; considerable judgment and experience are called for, at times, to obtain a field of contours which can be regarded with fair confidence as a true representation of the wind flow at the level in question.

The chart for 500 millibars is obtained similarly by gridding the 700–500-millibar 'partial' thicknesses with the 700-millibar contours, with adjustments as before (the thickness of the layer from one standard level to the next is known as the 'partial' thickness of that layer). An additional criterion for consistency of analysis is obtained from the drawing of thickness lines for 1000–500 millibars by graphical subtraction of the 1000-millibar contours from the 500-millibar contours; these thickness lines must be in agreement with the thicknesses and thermal winds for the 1000–500-millibar layer as plotted on the chart. The isopleths of 1000–500-millibar thickness have also a special application in forecasting changes in the surface pressure distribution. The practice at some forecasting centres is to draw the 500-millibar contours directly from the 1000-millibar contours and the 1000–500-millibar thicknesses.

Repetition of the technique of building up from one standard level to the next produces the contour charts for 300, 200 and 100 millibars. The number of observations of contour height and wind normally decreases above about 300 millibars and is liable to be very small at 100 millibars; moreover the contour heights based on radiosonde observations at 200 and 100 millibars are less reliable than the direct wind observations. Consequently the contours at these levels are drawn to fit the winds rather than the observed contour heights and it becomes more necessary than ever to ensure that the contours are properly related to those at the level next below as well as to those on the preceding chart. The tropopause frequently occurs between the 300-millibar and 200-millibar levels and it is usual to draw on the 200-millibar chart the intersection of the tropopause and the 200-millibar surface, thus separating the areas where the 200-millibar surface is above the tropopause from those where it is below. The axes of jet streams may be marked by coloured lines on the 200- or 300-millibar chart.

The study of contour charts soon reveals that moving patterns of airflow are as characteristic of upper levels as they are of the surface. Whereas, however, closed circulations in the form of depressions and anticyclones are common at the surface such circulations tend to disappear in the upper air where they are frequently replaced by troughs and ridges. These troughs and ridges are often major features of the general circulation in the sense that a single one of them may cover a substantial sector of the northern hemisphere. A good idea of the scale of these features is given by Figs 108 and 109 (Section 140) which are specimen contour charts for 500 and 200 millibars for the North Atlantic. These show troughs over eastern Europe and the eastern seaboard of North America and a ridge over western Europe, the general shape suggesting that they are part of a wave-like flow pattern. A rough measure of the amplitude of the wave pattern is given by the 5400-metre contour line on the 500-millibar chart which on the western Atlantic is as far south as 45°N and over Scandinavia is at 65°N – an amplitude of 20 degrees. Troughs and ridges on this scale are common features of the upper flow patterns; they are known as 'long waves'. The long-wave pattern is often distorted by smaller-scale features such as the trough west of Ireland in Fig. 108. The long-wave features are often persistent and slow-moving and a particular trough or ridge may dominate the weather of a large area for several days. It will be seen in a later section how this tendency for persistence can be a useful forecasting aid. On charts of the 1000–500-millibar thickness there are patterns broadly similar to those on the 500-millibar contour chart but, whilst the most common features are likewise troughs and ridges in the thermal pattern, there are sometimes closed areas of warm or cold air known as warm and cold pools. These are commonly very slow-moving or quasi-stationary. Similar pools are also to be seen in the patterns of partial thicknesses of the various layers. Since the temperature in the lower stratosphere varies inversely with the height of the tropopause and low tropopauses occur over well-developed depressions at the surface, it is usual to find warm pools in the lower stratosphere (for example, in the partial thickness pattern on the 200–100-millibar chart) over cold pools in the troposphere.

131. PREPARATION OF FORECAST SURFACE CHARTS

With the advent of the computer many of the laborious processes needed to produce analyses and forecasts have been superseded, and this is a development which will continue. This section is included, however, in order that the methods underlying the complicated process of producing forecast charts without the assistance of a computer may be appreciated.

We have seen that the large mass of data received by the forecaster in the form of individual reports is presented in systematic form on surface synoptic charts. On these charts the main emphasis is on the pressure systems, fronts and air masses and from these alone a tolerably good idea of the broad distribution of weather can be obtained even without the detailed observations. It is clear therefore that in preparing a forecast it is necessary first of all to consider how the pressure systems and fronts shown on the current chart are likely to move and change during the period for which the forecast applies. The most natural way to develop and formulate these ideas is to sketch the forecast isobars and fronts on a blank map. This is the method adopted in most forecast offices and it is strongly recommended to the student. A forecast, or prognostic, chart may be drawn to relate to any future time within those limits for which useful forecasts can be prepared. For general purposes in this country it is usual to prepare the forecast chart for 24 hours ahead of the current synoptic chart. Since some 3 hours must be allowed for the plotting and analysis of the current chart, the 24-hour forecast chart can usually be completed about 21 hours before the time to which it applies. For special purposes the forecast chart may be drawn for other time intervals as required.

The main considerations in drawing a forecast chart are as follows:

(i) the movement of existing pressure systems,

(ii) the evolution (formation, development and decay) of pressure systems,

(iii) the changes of pressure in areas remote from the main centres of activity,

(iv) the movement of fronts, and

(v) the formation and disappearance of fronts.

It must be emphasized that this separation into distinct steps is made solely to avoid confusion in the written description. In practice the process must be considered as a whole, each step affecting the others to a greater or lesser degree. The aim is to build up a coherent and self-consistent picture which follows on logically from the current charts. Moreover, not only the surface charts but also the upper air charts must be studied in the process of forecasting. The considerations listed above will now be discussed in turn.

Movement of pressure systems

Extrapolation. In this connection extrapolation means the continuation of the recent movement of the system for some further period. The motion of each anticyclone or depression over the past 24 hours will have been noted on the analysed charts and the positions of the centre at 6-hour intervals marked as already described. A first rough approximation to the position at which the centre might be expected 24 hours hence is given by assuming that its speed and direction of motion over the next 24 hours will be the same as over the past 24 hours. This would be an obvious conclusion if in fact over the past 24 hours the depression had moved in a constant direction at a steady speed, but such is not often the case. Usually the track of the centre as marked on the charts is curved to a greater or lesser degree and the distances moved in successive 6-hour periods are variable. Sometimes these changes of speed and direction of the depression are irregular and then extrapolation affords no help. Often, however, the track of the depression is a smooth curve and the speed increases or decreases steadily over the successive 6-hour intervals. With judgment the curved track may then be continued by eye and allowance made for the changing speed; this is likely to give a better approximation to the future position of the centre than the mere repetition of the movement of the previous 24 hours.

With a trough or a ridge there is no point centre and a track cannot be drawn. The axis of the system can however be drawn in and the process of extrapolation can be applied using the position of the axis on successive 6-hour charts if the changes thus indicated have been sufficiently regular.

It should be pointed out however that extrapolation consists essentially of continuing past trends and can take no account of new factors which may arise during the forecast period. It should therefore be regarded as a rough approximation to be modified or even disregarded in the light of other considerations.

Tendencies. An important clue to the movement of pressure systems is provided by the pressure tendency or isallobaric field. The following rules apply broadly but it must be borne in mind that the tendency field is often affected by the changes of intensity of systems even more than by their movement and it is often difficult to separate the effects of these two factors.

(i) A depression or anticyclone moves roughly parallel to the line joining the largest positive tendencies with the largest negative tendencies. The depression moves towards the area of falling pressure, the anticyclone towards the rising pressure.

(ii) If the tendencies are symmetrical with respect to the centre, the system is stationary.

(iii) A ridge or trough usually moves in a direction more or less at right angles to its axis: the ridge moves towards that side on which the pressure is rising and the trough towards that on which it is falling.

(iv) Taken in conjunction with the pressure gradient the tendencies afford some indication of the speed of the pressure system. Assume for simplicity that the pressure system is moving bodily without change of intensity. Then if the isobars drawn at (say) 1-millibar intervals are 50 miles apart a tendency of 30 (that is, a change of 3 millibars in 3 hours or 1 millibar per hour) would correspond to a speed of 50 miles per hour for the movement of the pressure system. If however the 1-millibar isobars are only 25 miles apart the speed of the system corresponding to the same tendency would be only 25 miles per hour. Thus for a given tendency, the stronger the pressure gradient the smaller the speed of the system and vice versa.

Steering. Experience shows that a depression or anticyclone often moves in the same direction as the upper winds, that is, along the contours at 10 000 feet or more above its centre; this is called the steering effect. Theory leads to a similar result except that the movement of the pressure system is related to the thickness lines instead of to the contours. Since however there is often little difference between the directions of the upper contours and of the thickness lines, it follows that the two results are in agreement. For practical purposes the thickness lines for the layer 1000–500 millibars are used in applying the steering principle and the effect is therefore often referred to as 'thermal steering'. This is however largely a matter of convenience, the 500-millibar level being one for which upper air data are regularly available. The speed of the surface pressure system is only loosely related to the speed of the thermal (or upper) wind over the centre; the two increase and decrease together but the speed of the surface system is usually less than that of the upper wind and no precise relation can be formulated.

The steering principle is most useful in regard to small systems without vigorous circulations since these can, so to speak, move through the upper patterns without disturbing them appreciably. Large and vigorous surface circulations on the other hand tend to twist up the warm and cold air into spiral-shaped tongues thus distorting the thickness pattern and also the upper contours so that the steering pattern is destroyed. The movement of such a system then becomes slow and irregular.

Miscellaneous. A depression with an open warm sector moves in the direction of the isobars in the warm sector. The ratio of the speed of the depression to the geostrophic wind is variable, its average value being about four-fifths; this ratio tends to be least with the strongest gradients. Some small wave depressions move a little faster than the geostrophic wind in the warm sector.

During the process of occlusion the depression slows down and a fully occluded depression is usually slow-moving or stationary. In applying the warm-sector rule just described it is obviously important to consider how soon it will be before the depression becomes occluded.

If in any sector of a depression the wind is markedly stronger than in the remaining sectors, the depression tends to move in the same direction as that of the strongest wind (that is in the direction of the isobars where they are nearest together) (see Fig. 101).

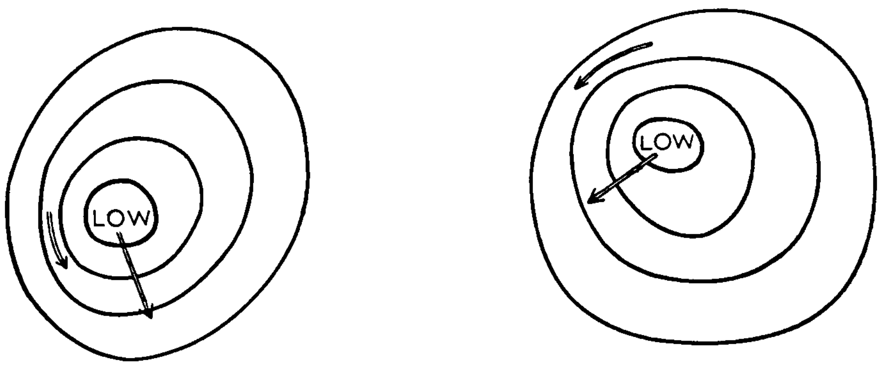


FIG. 101. *Movement of a depression in the direction of the strongest winds circulating round it*

Depressions tend to move round large anticyclones in the same direction as the anticyclonic circulation of winds.

Secondary depressions often move in the direction of the circulation round the primary. This applies particularly to small secondaries on the cold front of the primary and to secondaries forming within the polar air. On the other hand secondaries on the warm front or at the point of occlusion often move along the warm front more or less radially away from the primary; an indication of the movement is then usually given by the tendencies and also by the thermal steering.

An anticyclone or ridge separating successive depressions of a family normally moves with the depressions.

An anticyclone forming in an outbreak of polar air behind the cold front of a frontal depression moves with the mass of cold air (generally towards lower latitudes).

A warm anticyclone moves slowly and tends to become stationary.

The movement of pressure systems may be influenced by topographical features. For example, a depression approaching a high plateau or mountain range is often brought to a stop or it is deflected so as to avoid the high ground; occasionally, if the ground is not too high, the upper part of the disturbance crosses the barrier and the surface depression then appears to re-form on the lee side. It is however difficult to lay down general rules regarding topographical influences; such influences are usually introduced into forecasting practice indirectly by means of the local indications such as pressure tendencies or through a knowledge of the tracks habitually followed by pressure systems in given situations.

Evolution of pressure systems

It has been found that surface pressure systems are likely to develop where there are certain characteristic differences between the contour patterns near the surface and those in the upper levels of the troposphere—that is when certain marked characteristics are present in the thickness pattern. In the application of these ideas the 1000–500-millibar thickness is normally taken as being representative of the atmospheric layer in which the development takes place. The effects occur where there are sharp changes in the direction and spacing of the thickness lines, and the development is proportional to the speed of the thermal wind. Accordingly the forecaster examines the thickness chart for sharp troughs and ridges and for spreading out (diffluence) or closing together (confluence) of the thickness lines, coupled with close spacing of thickness lines. These may occur in a variety of combinations and only a few of the simpler models can be considered here. These are shown in Fig. 102. Development is to be looked for in those parts of the thickness pattern marked A (divergence and anticyclonic development) and C (convergence and cyclonic development).

Cold thermal trough (Fig. 102(*a*)). Cyclonic development tends to occur on the forward side and anticyclonic development on the rear side of the trough. A depression in a position such as X will tend not to develop; rather it will weaken and may disappear. It may however be steered along the thickness lines until it rounds the trough when it will begin to deepen in the position C. An anticyclone at X on the other hand will be steered towards A where it will settle down and intensify; it will not readily come round the trough but if it does so it will weaken in the cyclonic development area C.

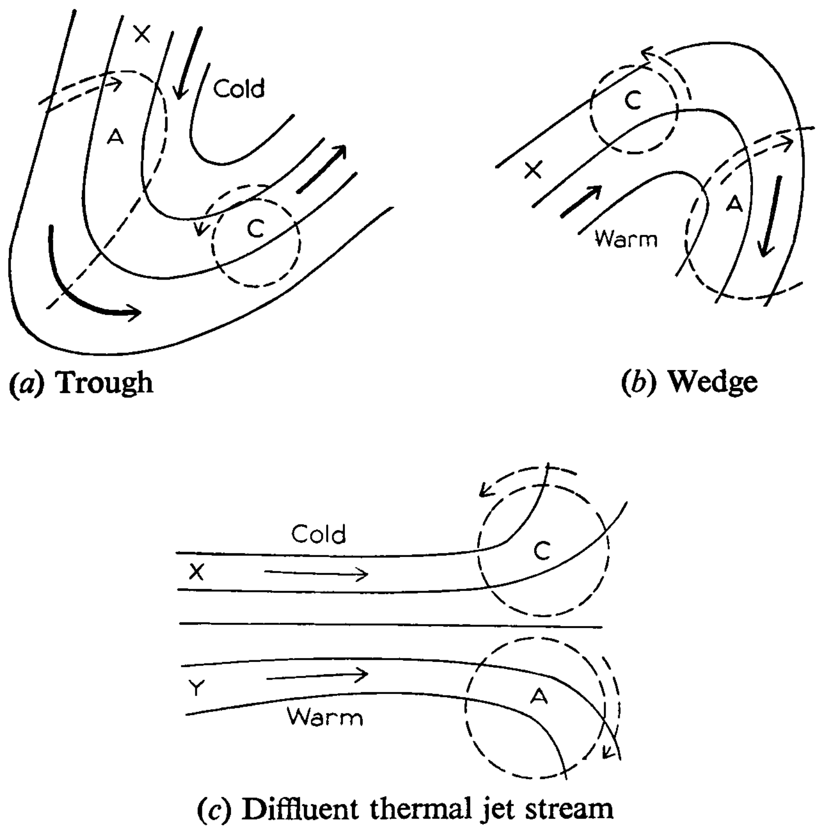


FIG. 102. *Cyclonic* (C) *and anticyclonic* (A) *development in relation to the thickness pattern*

The thickness lines and thermal-wind direction are shown by full lines and arrows; the surface isobars and wind direction are shown by broken lines and arrows.

Warm thermal ridge (Fig. 102(*b*)). This is the opposite of the cold trough. Cyclonic development occurs at C, behind the crest of the ridge, and anticyclonic development at A, ahead of it. A depression at X is not easily steered round the ridge but moves towards C where it settles down and often deepens.

Diffluent thermal jet (Fig. 102(*c*)). The thermal jet (not to be confused with the jet stream) is represented by the closely spaced thickness lines running from XY to CA: at C and A the lines spread out and the pattern is said to be diffluent. Cyclonic and anticyclonic development occur respectively at C and A and since the thermal wind over C and A is weak the systems developing here are often slow moving. A depression at X will be readily steered towards C where it slows down and deepens. A depression at Y on the other hand tends to weaken as it is steered towards A. Frequently however such a depression also moves across the thickness lines towards C where it may become slow moving and deepen.

Tendencies. The following rules though almost self-evident are given here in some detail owing to their frequent application in forecasting.

(i) A depression is deepening if pressure is falling all round the centre or if the rate of fall on one side is greater than the rate of rise on the other.

(ii) A depression is filling up if the pressure is rising all round the centre or if the rate of fall on one side is less than the rate of rise on the other.

(iii) The intensity of a depression or anticyclone does not change if the tendency is zero all round the centre or if a fall on one side is balanced by an equal rise on the other.

(iv) An anticyclone is intensifying if pressure is rising on one side more rapidly than it is falling on the other. Conversely it is weakening if pressure is falling at the centre or falling on one side more rapidly than it is rising on the other.

New depressions and secondaries. In addition to estimating the movement and development of features already on the chart, it is always necessary to consider the possibility of new formations. The various ways in which depressions may form have already been discussed. The first sign of a new disturbance is usually a local fall of pressure not accounted for by the movement or development of the systems already present and for this reason the closest attention should always be given to the pressure tendencies. If the tendencies indicate the development of some recognized type of depression the forecast chart would be modified accordingly. If for example pressure begins to fall locally on a slowly moving or stationary front a new frontal depression is indicated; if the fall occurs in a broad current of polar air a non-frontal depression or secondary is indicated, and so on. But experience alone will enable the probable degree of development to be estimated and much depends upon the part of the world and the season of the year.

The following points are also useful:

(i) A warm-sector depression usually starts as a wave on the cold front of an existing (primary) depression. It may remain in wave form for some time without deepening but on moving into a favourable position, usually forward of a thermal trough, it may deepen rapidly for a time until it is well occluded.

(ii) A wave disturbance on a warm front usually moves fairly quickly along the front away from the main depression; it is unusual for such a system to deepen.

Miscellaneous. A fully occluded depression rarely deepens. It usually fills up slowly and since the filling process may last for several days the depression is a rather persistent though weakening feature. An old depression often fills up rapidly when a new deepening centre moves into its circulation.

It is rare for two neighbouring depressions to deepen at the same time.

Instability of the warmer air mass within a depression tends to favour its deepening, while stability has a damping effect.

A temporary cold anticyclone formed behind a frontal depression is liable to collapse rapidly on the approach of a deepening depression. The weakening of such an anticyclone from any cause is usually followed by its complete collapse.

A warm anticyclone is relatively stable and approaching depressions tend to be deflected round it.

A warm anticyclone is liable to collapse if invaded by a cold front. A cold anticyclone may then develop behind the cold front and is said to absorb the warm anticyclone.

Changes of pressure in areas remote from the main centres of activity

In preparing a forecast chart it usually happens that there are some areas, remote from pressure centres and fronts, in which the existing features of the synoptic chart afford little guide to the likely pressure changes. It is true that these pressure changes are usually small so that no great errors are introduced if the forecast pressures are made equal to the current ones; occasionally, however, the slow pressure changes in these regions are part of a general trend which if continued over 24 hours or more may become important. It is best to estimate the probable pressures 24 hours ahead at a few selected points and to draw the forecast isobars to fit these values, joining them smoothly with the isobars already drawn for the more active regions. For estimating the pressures 24 hours ahead the plotted 3-hour tendencies are obviously useful but they should be used in conjunction with the pressure trends (changes in the past 24 hours) which can be read off the appropriate charts. In low latitudes and to a less extent in temperate latitudes the normal diurnal variation of pressure (Section 6) needs to be allowed for especially when the tendencies are small. For example a tendency of —20 (fall of 2 millibars in 3 hours) at the Azores is not to be regarded as significant if it occurs on the afternoon chart since this is about the normal diurnal fall of pressure; in the absence of a near depression or front it is usually followed by a corresponding rise during the evening. On summer afternoons similar falling tendencies often appear over France and central Europe but if they are no greater than about —15 they are not as a rule synoptically significant.

Movement of fronts

The first approximation to the velocity of a front is given by the geostrophic component of wind at right angles to it, which may be measured by means of the geostrophic wind scale (see Section 104). The following additional rules are also of assistance:

(i) The speed of a warm front is generally considerably less than the geostrophic component (roughly two thirds of its value), particularly within an occluding depression or when the isobars in advance of the front are inclined at a small angle to it (see Fig. 103).

(ii) The speed of a cold front is generally about equal to the geostrophic value or a little greater.

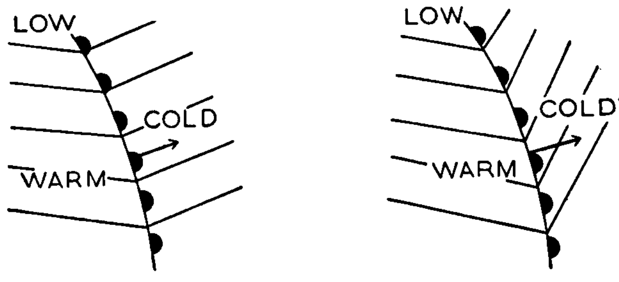


FIG. 103. *Movement of a warm front*

With the isobars as in (*a*) the front moves in the direction of the arrow with a speed of about 4/5 of the geostrophic component. With the isobars as in (*b*) the front moves at about 2/3 the geostrophic component.

(iii) When the geostrophic wind across the front is small the direction of motion may be determined by other considerations, for example, a flow of air across the isobars will occur from an area where the pressure is rising to one where it is falling and this may be sufficient to neutralize or reverse the geostrophic component on the front (see Fig. 104).

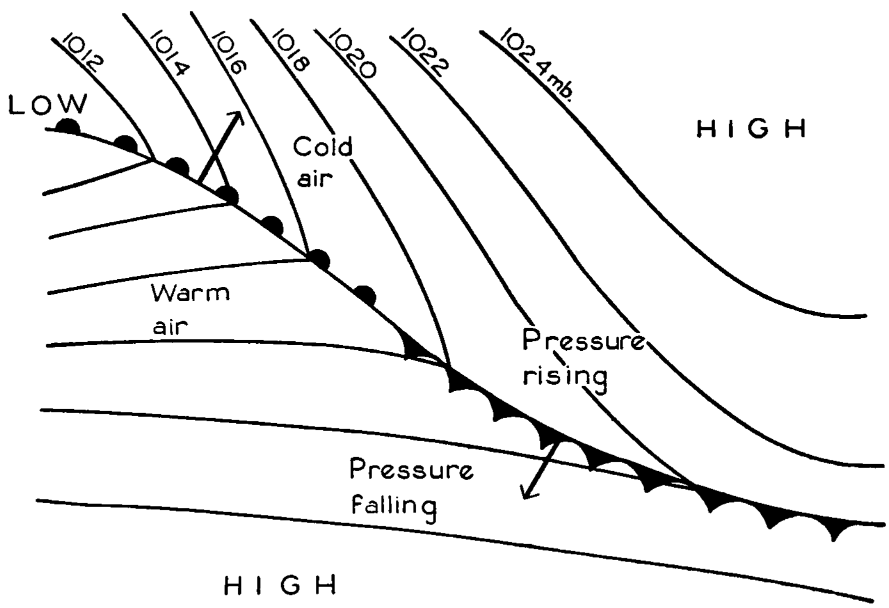


FIG. 104. *Isallobaric movement of a front*

A front always tends to move towards the direction of greatest falling pressure. When the pressure gradient is weak the motion may be against the direction of the geostrophic wind.

(iv) An occlusion moves roughly with the speed of the geostrophic wind component but there is no precise rule.

The speed of a front is often given most accurately by measuring its movement between successive charts. This should always be done if possible as a check on the foregoing rules. Appreciable departures from the movements given by the rules should be given due weight in estimating the frontal positions on the forecast charts.

The above rules can be applied in a simple direct way only if it is assumed that the geostrophic wind across the front is not likely to change much during the forecast period. In practice there are many complications of which only a few can here be mentioned by way of illustration. In the case of the straightforward deepening of a depression (that is when the main fall of pressure occurs in the central part of the system) the geostrophic wind across the front will increase and the front may be expected to speed up. On the other hand it often happens that a fall of pressure is concentrated along the front itself, leading either to an elongated trough along the front or perhaps to a small separate closed centre. In either case the geostrophic wind across the front will be very much reduced along a considerable length of the front and in the second case it will be reversed over a limited distance so that the front will become very slow-moving or possibly retrograde. Lastly, with an old frontal depression the occlusion often spirals outwards from the centre. The part of the occlusion remote from the centre gradually tends to become parallel to the isobars and, in consequence, almost stationary.

Formation and disappearance of fronts

For forecasting for periods up to about 12 hours it is as a rule sufficient to consider only those fronts which are already present on the current synoptic charts. The question of the formation of new fronts as well as the strengthening, weakening and disappearance of existing ones is hardly likely to arise; it does however present a problem to the professional forecaster concerned with longer periods and the student should at least be aware of the possibilities. It may generally be assumed that fronts will weaken in anticyclones and pressure ridges. They will weaken in those parts of the thickness patterns favourable to anticyclonic development and strengthen in the parts favourable to cyclonic development (see page 259). For short-period forecasting the weakening of a front is shown by rising pressure and general decrease of rain and cloud in its vicinity, while changes of the opposite type indicate a strengthening of the front. Generally speaking a major front is associated with a close spacing of thickness lines; if it can be seen that because of the circulation of winds the thermal gradient is likely to become strong over a considerable distance the formation of a new front should be anticipated. A front which is drawn crossing the thickness lines at a large angle should be regarded with suspicion; if it is a true front its effects are usually confined to a comparatively shallow layer of the atmosphere near the surface.

Secondary cold fronts are often a feature of an outbreak of cold air. The showers in unstable polar air often occur along an extended line more or less at right angles to the air stream, sometimes merging into an unbroken elongated rain area, and since a shallow trough of low pressure sometimes develops along the rain belt the synoptic chart shows all the appearances of a minor cold front. It is not uncommon for a series of such secondary fronts to develop and be carried round in the circulation of a large slow-moving depression after the main cold front has passed. It is not usually possible to forecast the location of such a front before its formation but once it has appeared on the current synoptic chart it should be included in the forecast chart with due allowance for its probable movement.

Drawing the forecast chart

The various considerations involved in the preparation of the forecast chart have been discussed in some detail in the preceding sections but it is perhaps desirable here to summarize the actual procedure. First the expected positions of the centres of high and low pressure and the troughs and ridges should be sketched in lightly, using extrapolation for the first approximation with modification in the light of tendencies and steering. Next the forecast central pressure should be estimated for each system. After this the forecast frontal positions should be lightly sketched in; these will be based on the geostrophic winds of the current synoptic chart and any change of speed can be allowed for only rather vaguely at this stage. A few estimated forecast pressures at points remote from active pressure systems and fronts should now be inserted and the isobars sketched in. Generally speaking the isobars should run in evenly spaced and more or less smooth curves except for the kinks where they intersect the fronts. If the predicted pressures lead to a local crowding of the isobars or imply unaccountably large pressure changes in any part of the chart they should be smoothed out unless there is a good reason not to. Finally each frontal position should be adjusted so that the frontal movement is not inconsistent with the appropriate geostrophic wind at both the beginning and the end of the forecast period. For example if the geostrophic wind across the front is 40 knots on the current synoptic chart and 20 knots on the forecast chart then a predicted frontal movement corresponding to a steady speed of 30 knots throughout the 24-hour period would be reasonable, whereas a sustained speed of 40 knots would be inconsistent.

132. PREPARATION OF FORECAST UPPER AIR CHARTS

Most upper air charts, whether analyses or forecasts, are prepared using a computer (see Section 133) but it is still useful to know something of the methods used when computer forecasts are not available.

Basically, in preparing forecast upper charts, the forecaster applies the same technique as is used in the analysis of the current charts. Starting with the forecast surface pressure chart he builds up the forecast contour charts at the upper levels by the successive addition (gridding) of the forecast charts of appropriate partial thickness. The element of forecasting is involved mainly through forecasting the partial thickness lines for the various levels. The elaborate technique employed at forecasting centres can only be acquired through long experience and no more will be attempted here than a brief descriptive outline.

It will be recalled that the movement and development of surface pressure systems is influenced by the thickness pattern while the thickness pattern is itself modified by (among other factors) the surface circulations. It is evident therefore that the preparation of forecast surface and upper air charts are interrelated problems and ideally they should proceed hand in hand by a process of modifying one chart in the light of the others to bring the whole set of charts into reasonable conformity.

The changes in the various thickness patterns are forecast mainly by extrapolation of the troughs and ridges on the current chart and by considering how the thickness lines will move under the influence of the wind circulations in the appropriate layers. Heating or cooling of the air as it moves over warmer or colder parts of the earth's surface will affect the partial thicknesses to a greater or less degree and the experienced forecaster can to some extent estimate and allow for these effects. Studies of past records have shown that there are more or less definite limiting latitudes to which air masses having particular thickness values can penetrate without becoming modified. Charts showing these limiting latitudes (which vary from longitude to longitude) which may be reached by the various thickness lines have been prepared for each month of the year and are useful aids to forecasting.

133. NUMERICAL FORECASTING

The idea of using equations that describe the motion of the atmosphere to calculate the changes in distribution of temperature and pressure, although suggested some time ago, only became feasible with the advent of high-speed computers to perform the large number of calculations required. The movement and development of atmospheric systems are complex and even a large computer can only deal with a simplified model of the atmosphere. At present at Bracknell, developments at three levels—1000, 500 and 200 millibars—are examined. The area analysed is divided into regularly spaced grid points and the computer, fed with the observations made simultaneously from land stations, ships and radiosondes, assigns a value for the pressure and temperature to each grid point for the hour of observation. The machine solves the equations to calculate how the contour height and thickness at the upper levels will change at each grid point during the next hour, and using these new values it carries the calculation forward in steps of one hour until eventually it produces a forecast for 24 hours ahead of the contour heights at each of the three levels and of the thickness of the 1000–500-millibar layer. The whole operation from feeding in the observations to the final forecast is completed in just over one hour.

The forecast values are printed in their correct geographical positions by a high-speed printer and isopleths may be drawn in by hand. Alternatively the print-out may be directly in the form of isopleths, numbers being used to delineate the patterns, or better still, a line-drawing machine may be coupled to the computer output. Examples of these computer-produced charts are shown in Section 140, Figs 108, 109 and 111.

Charts for any intermediate levels are produced by the computer by a regression interpolation from the calculated levels.

Upper air forecast charts at the Central Forecasting Office at Bracknell are now all produced by the computer. Eventually all forecast charts will be based on the numerical charts produced at Bracknell and distributed by facsimile to subsidiary forecasting offices where the forecaster will add the details of local weather.

134. PREPARING THE FORECAST

The construction of forecast charts as described in the preceding sections is only a means to an end – the preparation of the forecast. To a certain extent the forecast is implicit in the forecast charts; the forecast fronts and depressions indicate the areas of bad weather, the winds are indicated by the forecast isobars and contours, and the general character of the weather away from pressure systems and fronts can usually be inferred from the broad air-mass properties. But there are other considerations which enter into the forecast which are not implicit in the forecast charts but dependent upon the more detailed structure of the pressure systems, fronts and air masses. These considerations are important for the forecasting of such features as fog, thunderstorms, turbulence, showers, the distinction between snow and rain and so on. A physical account of these phenomena has been given in Part I and by using that knowledge in conjunction with the forecast charts the student should be able to prepare useful forecasts in some of the simpler cases. But most meteorological situations are more or less complex to a degree which cannot be adequately dealt with in a textbook. Only long experience can show how they should best be handled and so the pilot or navigator has perforce to rely on the professional forecaster for his advice. In this book nothing more than a brief outline can be given of the simpler considerations involved in translating synoptic and forecast charts into a weather forecast.

Types of forecast

The formulation of a forecast is a matter which depends upon the purpose for which it is intended. The various types of aviation forecast may be classified as follows:

General inference. A general description of the distribution of pressure and of the position of the fronts based on the current synoptic chart, followed by the probable movements and developments of the pressure features and fronts based on the forecast charts. In the days before forecast charts were drawn the general inference was the only means of indicating the expected changes in the pressure field; but now it is regarded rather as supplementary to the charts; thus it amplifies the information which they portray, particularly in regard to the activity of fronts and the properties of air masses which cannot easily be represented cartographically, and indicates degrees of confidence and possible alternative developments.

Area forecasts. General indications of flying weather conditions to be expected over substantial areas for a stated period with little attention to detail.

Route and flight forecasts. The expected conditions along a specified route for a given period of time (route forecast) or for the duration of a particular flight (flight forecast).

Local forecasts. The expected conditions in some detail for the immediate vicinity of a flying centre for a stated period.

Aerodrome forecasts. Area, route and flight forecasts need to be supplemented by forecasts for terminal aerodromes and alternates for appropriate stated periods.

The broad principles employed in the preparation of these forecasts will now be discussed. It must however be repeated that every situation shows some variations from these broad principles which cannot be dealt with in a textbook and which only the experienced forecaster can handle.

The area, route and flight forecasts will be considered together since they are prepared almost entirely from the current and forecast charts. Local and aerodrome forecasts require in addition a study of the way in which the meteorological situation is modified locally by geographical peculiarities and they are therefore dealt with separately.

135. AREA, ROUTE AND FLIGHT FORECASTS

Frontal weather and cloud

The pilot's first concern will probably be the extent to which the area or route will be affected by the bad weather and extensive cloud associated with fronts. The areas likely to be so affected can be seen at a glance from the current and forecast charts and allowance can be made for the frontal movements during the forecast period. Reference should be made to Chapter 16 for a detailed description of the weather to be expected on a flight through a frontal depression, and for a tabular statement of normal frontal characteristics. The expected changes in intensity of the fronts must also be taken into consideration. In summer, especially, the activity of a front may be increased by the presence of a moist unstable air mass and any front, but more particularly a cold front, advancing into such an air mass is liable to give rise to an extensive outbreak of thunderstorms. With weakening fronts the cloud sheets tend to break up and the normal cirrostratus, altostratus and nimbostratus assume the forms of cirrocumulus, altocumulus and stratocumulus respectively.

Air-mass weather

Away from the fronts the broad features of the expected weather will be inferred from the characteristic air-mass properties. These, as they apply to the temperate regions of the northern hemisphere, are described in Chapter 15. It will be recalled that the air masses are broadly classified into warm and cold types according to their respective origins in tropical or polar regions, with a further subdivision into maritime and continental according to their recent location over ocean or land areas. In any given situation however the standard air-mass properties are modified to a greater or lesser degree by various factors such as heating or cooling at the earth's surface and the evaporation of moisture from the sea, vegetation or wet ground. The resulting changes in air-mass properties may to some extent be estimated but to deal with them properly representative observations of upper air temperature and humidity within the air mass are essential; these should be plotted on a tephigram.

Clouds in relation to air masses

The broad distinction is between cumuliform clouds characterized by vertical development and layer cloud occuring in extensive horizontal sheets. The former occur with unstable air masses and are most characteristic of maritime polar air which usually contains adequate moisture for cloud formation. Layer clouds occur with stable air masses, most often with tropical maritime air. But there are widely varying degrees of stability and instability for a given type of air mass and the forecasting of cloud in a given situation is greatly helped by the tephigram. The reader is referred to Chapter 6 for a full discussion of cloud formation in relation to lapse rate; only a broad summary is given here. In general if the lapse rate is steep, clouds are mainly cumuliform; the height to which the clouds grow is roughly the same as the height to which the steep lapse rate continues. Deep unstable layers are therefore associated with large cumulus or cumulonimbus clouds; shallow unstable layers with only small cumulus. Stable layers and inversions on the other hand are associated with layer cloud which may be formed either by turbulence or by the spreading out of cumulus clouds formed in rising currents in unstable layers below. Over land, except in winter, the cumuliform clouds usually build up during the day and disperse at night; stratiform clouds on the other hand tend to increase at night and sometimes they disappear by day; these changes are clearly associated with the changes in lapse rate caused by the variation of surface heating. In this connection it is often useful to estimate the changes in the lapse rate in the lowest layers caused by the surface heating and cooling. Over the sea there is no appreciable diurnal variation of lapse rate; cumulus or stratocumulus clouds may occur at all times of the day in a polar air current whose surface temperature is lower than that of the sea. If on the other hand the air is warmer than the sea the tendency is for stratus to form.

Precipitation in relation to air masses

In areas away from fronts there are two principal types of rainfall: convectional and orographic.

Convectional rainfall occurs with a steep lapse rate and high moisture content in at least the lower layers of the atmosphere. The principal aid to forecasting this type of rainfall is the tephigram, as described in some detail in Chapter 6. In general the more unstable the lapse rate and the higher the humidity the more vigorous is the convection and the heavier is the rain associated with it; in extreme cases violent thunderstorms and heavy hail occur. Much depends also upon the surface pressure distribution. Within a depression or region of cyclonically curved isobars the convergence caused by inflow of air across the isobars not only aids the vertical motion but also increases the available moisture and the vigour of the convection is thereby increased. A shallow depression is more favourable than an intense one for vigorous convection, for a weak circulation of wind prolongs the contact of the air with the warm surface while strong winds tend to shear off the rising columns of heated air; the most intense convectional effects occur therefore in summer in shallow depressions moving slowly over land areas. Within an anticyclone or ridge there is usually subsidence coupled with the outflow of surface air across the isobars. This not only tends to diminish the instability but also brings down drier air towards the surface, and these two factors tend to damp out the convection.

In temperate latitudes convectional rainfall occurs mainly in polar maritime air, chiefly in the form of showers from detached cumulonimbus, the individual showers often being separated by areas of clear sky. Over the sea this type of rainfall is most prevalent in winter when the cold air moves over relatively warm water and surface heating leads to instability. Windward coastal regions also are subject to these showers and, with diminishing frequency, the showers may penetrate considerable distances inland especially through gaps in mountains or hills; thus in the Midlands of England with north-westerly winds showers often penetrate via the Cheshire gap, and with south-westerly winds via the Bristol Channel. In summer the sea is relatively cool and it is over the heated land that convection chiefly occurs, the frequency of showers often increasing with the distance from the windward coast. Owing to the diurnal variation of temperature, convectional rainfall over land in summer occurs chiefly in the afternoon; on the other hand, over the sea convectional rainfall may occur at any time, day or night. There is however a tendency for showers in coastal areas to occur most frequently in the early morning.

Orographic rainfall. The conditions associated with orographic rainfall and the factors which cause variations in its intensity have been discussed in Chapter 6. We may summarize by saying that orographic precipitation should be forecast whenever an airstream which is saturated or nearly saturated blows against hills or mountains or a high coastline. If the lapse rate is stable and there is no front, only drizzle or light rain is probable; if a front is present or there is instability (including convective instability) the intensity of the rainfall will be increased. To the lee of the high ground there is a region of diminished rainfall, or even none at all; this is the so-called 'rain shadow'.

Drizzle is precipitation in the form of very small drops. It occurs in association with a stable lapse rate and stratiform clouds. The processes which give rise to drizzle are turbulence and the less-marked orographic effects. It is especially liable to occur when warm moist air passes over a cold sea surface; if there is a sufficient degree of turbulence the drizzle may occur over the open sea but it is more usual on the coasts and for some distance inland where turbulence increases as the airstream passes from sea to land. It frequently occurs in the warm sector of a frontal depression but may equally well occur in the outer regions of an anticyclone if the airstream is damp enough.

Snow should usually be forecast when the surface temperature is below about 4°C in situations which would otherwise give rain. The exact critical temperature however depends upon circumstances and there is no hard-and-fast rule. If the wet-bulb temperature is low, precipitation which starts as heavy rain may soon turn to snow or a mixture of rain and snow even though the dry-bulb temperature is a few degrees above the critical value of 4°C to begin with; under these circumstances evaporation from the raindrops cools the air to its wet-bulb temperature; melting snow also is liable to lower the dry-bulb and wet-bulb temperature by a few degrees in the critical region just above the freezing-point. On the other hand drizzle or very fine rain may occur with the temperature nearly down to freezing-point.

Visibility in relation to air masses

The factors which affect visibility and give rise to fog, mist and haze are discussed in detail in Chapter 9 and it is necessary here to consider these factors only in relation to the synoptic situation. It will be recalled that fog arises in the main from two causes: radiation and advection. Radiation fog occurs chiefly over land at night and in the early morning, with clear skies and light winds; the fog develops where the ground cools by radiation to the clear sky. The central regions of anticyclones and ridges often give rise to the requisite conditions and indeed the advantages arising from the dry quiet weather of an anticyclone are often nullified by its liability to fog, especially in autumn and winter. For dense fog the air must be cooled below its dew-point not only at the surface but also up to a height of a few hundred feet. Some of the worst fogs begin under the special conditions of light wind and clear sky at night after the air has been made damp by rain falling through it. This occurs most frequently in a depression or trough with weak pressure gradients but the pressure distribution does not matter so long as the pressure gradient is weak.

Advection fog occurs when moist air flows over a cold surface of land or sea; the temperature of the cold surface must be below the dew-point of the air. This type of fog occurs at any season of the year and at any time of the day or night and is not restricted to occasions of light winds or clear skies. Over land it is especially liable to occur in winter with the incursion of mild damp air over a snow-covered surface or after a spell of frosty weather. Over the coastal waters of the British Isles it occurs chiefly in late spring and early summer while the sea is still cold. Tropical maritime air from the Atlantic then frequently gives rise to fog near the south-west coasts of the British Isles. Similarly, heated air from the continent, after cooling over the North Sea, may give rise to fog on the east coasts.

In general, visibility over land and sea is best when the lapse rate is steep; a small lapse rate and especially an inversion is often associated with poor visibility. Visibility is, however, affected also by dust, smoke and other solid impurities in the atmosphere. A steep lapse rate with its associated upward air currents readily disperses the impurities while a stable lapse rate prevents this dispersal and causes the impurities to accumulate near the surface. Under these conditions the visibility may be adversely affected not only near the source of pollution but for a great distance downwind.

Ice accretion

The meteorological conditions leading to ice accretion on aircraft are fully discussed in Chapter 8. Any clouds containing water droplets will probably give rise to airframe icing if the temperature is between 0°C and —40°C (or even below on occasions) though as a general rule the risk is smaller as the clouds become colder. Cumuliform clouds and particularly cumulonimbus are the most dangerous. Airframe icing is heaviest in the more 'active' types of cloud, that is, in clouds associated with active fronts, in cumulonimbus associated with vigorous convection and in clouds intensified by orographic uplift. Forecasting airframe icing is thus a matter of forecasting clouds – their horizontal and vertical extent as well as their type – in conjunction with the distribution of upper air temperature over the area or route concerned. Therefore a good selection of recent upper air ascents plotted on tephigrams is essential in addition to the synoptic chart. The degree of airframe icing is classed 'light', 'moderate' or 'severe' according to the expected severity; alternatively the 'icing index' may be said to be 'low', 'moderate' or 'high'. Further, when rain ice is expected it must be specifically mentioned in the forecast.

Since engine icing can occur over a wide range of conditions (including clear air and temperature above 0°C) and since its occurrence is affected by the design of the engine, forecasts of engine icing are not normally provided.

Winds

For area, route and flight forecasts the upper winds constitute perhaps the most important item because of their direct bearing on navigation, duration of flight and pay-load. Surface winds on the other hand are important for take-off and landing at particular aerodromes and they are more appropriately dealt with under local and aerodrome forecasts.

A good approximation to the existing wind at an upper level is given by the geostrophic wind measured from the appropriate contour chart. The standard levels for which upper contour charts are drawn are listed in Section 130; for intermediate levels it is not difficult, as a rule, to estimate the winds by interpolation. The forecast upper winds will likewise be estimated from the forecast upper air charts. In most cases the geostrophic wind may be used without modification but where the contours are sharply curved and where rapid changes are taking place in the contour field some adjustment is necessary as described in Section 24. Cyclonic curvature of the contours indicates that some reduction of the geostrophic wind is necessary; anticyclonic curvature requires an increase. When rapid changes are taking place in the contour field there is a tendency for the winds to blow at an angle across the contours outwards from areas where the contour height is rising and into regions in which it is falling. There is also a tendency for the change of wind speed to lag behind the change of gradient. Thus when the pressure gradient is increasing quickly the wind speed is usually less than the geostrophic value while for decreasing pressure gradient the wind speed tends to be greater than the geostrophic.

Above the level of 200 millibars (approximately 39 000 feet) analysis and forecasting of winds is made difficult because of the sparseness and reduced accuracy of the observations. It is useful at these levels to bear in mind the usual variation of wind with height – speed increasing with height towards the tropopause and thereafter decreasing in the stratosphere. Thus at 100 millibars the winds may be expected to be light compared with those at 300 and 200 millibars; for example, in summer over the British Isles the wind at 100 millibars rarely exceeds 30 knots. Moreover the changes of wind with time at very high levels are less than they are at lower levels so that in the absence of any other guiding principle a forecast of 'no change' may well be appropriate.

Surface pressure

The expected surface pressure at any position on the route can be read off directly from the forecast chart. Its importance lies in its effect on altimeter readings;

in particular when flying into an area of low pressure the altimeter if uncorrected will read too high, perhaps by many hundreds of feet. Allowance will be made for this of course in the flight plan but the forecast chart is useful as indicating the areas in which possible danger might arise.

136. LOCAL AND AERODROME FORECASTS

The problems of local and aerodrome forecasting are at the same time more restricted and more intensive than those of area and route or flight forecasting; for while the region considered is small, the precision demanded is much greater. Where the area forecaster is content to forecast 'local fog' or 'squally showers', to take two common phenomena, the local forecaster is expected to state if and when fog will occur, how bad the visibility will become, at what time of day the showers are expected and how severe the squalls may be. The local forecast of course is based on the forecast chart but special attention must be given to local peculiarities. The main factors causing local variations in conditions have been mentioned in connection with the individual elements but it is convenient now to present a comprehensive summary.

Causes of local variations in meteorological conditions

Surface wind

Winds stronger over the sea than over the land for the same pressure gradient.

Diurnal variation of wind; for the same pressure gradient the surface wind is stronger for large than for small lapse rate.

Sheltering effect of high ground to windward.

Deviating effect of hills on wind direction.

Funnel effect through valleys.

Katabatic winds down hillsides at night.

Diurnal land- and sea-breezes.

Variations in turbulence due to ground contour or variation in lapse rate; in particular the difference between turbulence over land and over sea.

Cloud

Orographic clouds over high ground.

Lowering of cloud base to windward of high ground and clearance in the lee.

Differences in turbulence as affecting low cloud.

At coastal stations the effect of wind direction on cloud formed by convection or turbulence. It is the wind direction at cloud level which is important. Thus with an on-shore wind stratus or other low cloud which has formed over the sea may drift in over the coast, or cloud may form at the coast where there is generally a little lifting because of the slight orographic effect and the increased turbulence.

Precipitation

Orographic precipitation or intensification of general rain by orographic features; rain-shadow effects in the lee of high ground.

Near the coasts, the differences between temperature conditions over sea and land as affecting instability showers and drizzle; diurnal and seasonal influences are important in this connection.

Tendency for thunderstorms to favour certain regions.

Visibility

Valley fogs.

Hill fogs (low cloud on hills).

Differences between land and sea.

Smoke effects (according to wind direction).

Effect of local clouds and local winds on the formation of radiation fog.

Effect of the nature of the ground on radiation fog.

Possible drift of fog from adjacent areas, especially from lower ground soon after dawn. Fog sometimes drifts from a direction backed by 45° or even a little more from the geostrophic wind when this is weak.

Temperature

Effect of altitude.

Low minima in valleys on clear nights.

Effect of local clouds or fog on diurnal heating or cooling by radiation.

Föhn effect in the lee of high ground.

Differences between land and sea.

The local forecaster may require long experience before he can deal with these problems with the desirable accuracy. It should not, however, be presumed that local forecasts are his sole preoccupation. The local variations depend in part on the general weather conditions and must be studied in relation to carefully analysed synoptic charts; moreover the vagaries of the local weather should be correlated with the general characteristics of the air masses and fronts which affect the locality.

137. DOCUMENTATION

The documents available at forecast offices for the briefing of aircrews varies from from time to time and from place to place, but a selection from the following charts will be available:

(i) A forecast of the surface chart for the area and time appropriate to the flight.
(ii) Contour charts for certain pressure levels.
(iii) Aerodrome forecasts for the destination and suitable alternatives.
(iv) Significant-weather chart.
(v) Tropopause chart.
(vi) Vertical cross-section of the weather expected along the route.

Significant-weather chart

Certain phenomena are considered to be of great significance to aircraft flights. These include turbulence, moderate or severe icing, marked mountain waves, etc. The significant-weather chart shows these phenomena by means of symbols plotted on the chart, which also shows the surface frontal analysis for the appropriate time. In areas where significant weather is expected, the amounts and types of cloud are shown with the heights of bases and tops of the various layers. A full list of weather considered significant is given in the meteorological section of the *Air Pilot* and in the ICAO document *Procedures for air navigation services – Meteorology* (see Appendix III).

A specimen significant-weather chart is included in Section 140.

Forecast chart for the tropopause height and vertical wind shear

This chart indicates the pressure altitude of the tropopause and is usually drawn at intervals of 50 millibars. Maximum winds are shown as arrows on the chart with the heights of the level of maximum wind marked along the arrow. Isopleths of wind shear, in knots per thousand feet, are also marked on the chart; these values are used when wind speeds at different heights are interpolated from the charts supplied. The values of wind shear are taken to apply both above and below the maximum wind levels and they should be added or substracted from the speeds read from the appropriate contour chart – usually for 300 millibars – supplied with the documentation.

Vertical cross-section

The route forecast for a long distance flight may include a simple vertical pictorial representation of the weather expected along the route. This vertical cross-section may be regarded as a profile along the route and is usually divided into zones defined by five-degree intervals of latitude or longitude.

Clouds are indicated by a pictorial representation of the general structure showing the heights of their bases and tops and their horizontal extent. The cloud outlines are often filled by shading or colouring, green being used for portions of the cloud in which the temperature is expected to be above 0°C and red for portions below 0°C. The isotherm for 0°C is shown as a heavy green line. Frontal surfaces are shown as sloping lines in the colour appropriate to the type of front as used on a surface chart (see Section 129). Precipitation and airframe icing are represented by the appropriate symbols at the levels at which they are expected to occur. This symbolic information is supplemented by a plain-language description together with details of upper winds and temperatures at a few selected levels. Heights on vertical cross-sections are given as pressure altitudes.

138. FORECASTING IN THE TROPICS

Despite the fact that in some tropical regions, especially certain coastal and mountainous areas, extensive cloud and heavy rain occur quite frequently, on the whole the weather in the tropics is considerably less hazardous to aviation than in temperate latitudes. Over both land and sea the predominating weather in most areas is fine with well-broken cloud at a safe height, good visibility and light or moderate winds. Bad flying conditions in the tropics are mostly associated with one or more of the following circumstances: active portions of the intertropical front and other convective conditions giving rise to heavy thunderstorms with their dangerously turbulent clouds and liability to severe airframe icing, orographic effects producing or intensifying cloud and rain, the obscuring of high ground and occasionally low-level aerodromes by low cloud, duststorms in arid areas, and tropical revolving storms. Most of these phenomena are comparatively infrequent and limited in extent but they are often extremely violent.

A brief inspection of a tropical synoptic chart will show that it cannot be interpreted in the same way as the temperate-latitude chart. The pressure gradients are as a rule weak and irregular; the common isobaric models – depressions, anticyclones, troughs and ridges – are rarely recognizable and certainly they cannot be regarded as discrete entities which move in more or less regular fashion and carry their characteristic weather with them, a property which is fundamental for temperate-latitude forecasting. Fronts likewise are exceptional and when they do occur they are as a rule weak and indefinite and difficult to trace. The geostrophic wind relation does not apply in very low latitudes and winds frequently appear to be haphazard in relation to the isobars; isobars and upper contours are of limited assistance for estimating the surface and upper winds. The synoptic chart does not 'hang together' as it does in temperate latitudes and perhaps is more truly regarded as a collective representation of local weather at a large number of stations than as an ordered picture of an over-all weather situation. This absence of large-scale predominating weather systems does however make for some advantages in forecasting, for at any particular place the weather shows a high degree of regularity in its behaviour since the local, diurnal and seasonal effects come into full play and are hardly at all masked by over-all influences. This means that at many places in the tropics the weather is more dependent on the time of day than on any other single factor and that changes from one day to the next are slight – the well-known daily sequence of fine clear nights and mornings and afternoon showers or thunderstorms provides a good example. It follows that a good knowledge of local weather is an important part of the tropical forecaster's equipment. Adverse weather on the other hand, since it can seldom be tied to moving depressions and fronts, is a much more difficult problem for the forecaster than it is in temperate latitudes.

It is beyond the scope of this book to give a detailed account of forecasting in the tropics and only a broad outline of the main considerations will be attempted. They may conveniently be dealt with under the following headings:

Diurnal effects and convection
Orographic effects
Fronts in the tropics
Tropical revolving storms.

These are not necessarily distinct from one another and in any given situation a number of them will probably have to be considered together.

Diurnal effects and convection

At many inland places in the tropics the most significant feature of the weather is the daily build-up of cumulus or cumulonimbus cloud often culminating in afternoon and evening thunderstorms; the forecasting problem then consists mainly in estimating the incidence, time of occurrence and intensity of these phenomena. The greatest single aid is the morning tephigram. With experience the daily change of temperature in the lower layers due to surface heating can be estimated satisfactorily and the forecasting of convection then follows the usual method. It is particularly important for showers and thunderstorms, however, that coupled with the necessary unstable lapse rate there shall be adequate moisture both in the lower and middle levels. If the air aloft is dry the rising cumulus clouds readily evaporate.

Convection starts most readily over high ground and it is common in the afternoon to see the hills topped by massive cumulonimbus while over the adjacent plains the weather is bright and hot. For reasons not fully understood, but probably because of cool air flowing katabatically down the hillsides and undercutting the heated air over the plains, the showers and thunderstorms sometimes tend to spread to the plains during the evening and night. The sky usually clears before morning but radiative cooling of the air moistened by the rain often gives rise to early morning fog patches, which however clear quickly after sunrise.

Over the sea the diurnal variation of temperature is small and convection may occur at any time of the day or night. In the equatorial regions, roughly between latitudes 10° N and 10° S, the lapse rate is usually near the saturated adiabatic and the moisture content of the air is high. In the absence of surface heating convection is set off by other factors of which the principal one appears to be the convergence of winds in association with minor fronts (see below) – apart from the intertropical front. Near the coasts, especially on the seaward side, there is a marked tendency for convection cloud and precipitation to occur in the early morning and to disperse during the forenoon. This is possibly caused by the undercutting of the moist unstable air over the sea by the night-time land-breeze, for diurnal land- and sea-breezes are a prominent feature of the weather on tropical coasts. Outside the equatorial region, that is, in the trade-wind belts, convectional effects over the sea are small. There is usually moist air in the lowest few thousand feet in which the typical trade-wind fair-weather cumulus occur but above this there is usually a dry and clear layer.

Orographic effects

Extensive cloud and rain are formed orographically in the tropics wherever a broad current of air after flowing for a long distance over the warm oceans strikes an elevated coastline or mountain range. Such broad airstreams are seasonal features of the tropical circulation and are either monsoons or trade winds. Details regarding the incidence of these winds are given in Chapters 22 and 23. Here it is sufficient to note that heavy and persistent rain, often with intense cumulonimbus cloud and thunderstorms, occurs when maritime air rises over coastal hills and mountains. Inland the effects tend to become gradually less pronounced as the air dries out, although they may be accentuated at times by the convection due to solar heating.

Minor fronts in the tropics

The tropical regions occupy nearly one half of the earth's surface. Over this vast area which consists very largely of ocean the surface conditions are almost uniform and the air movement is on the whole rather sluggish. It is not to be expected therefore that distinct air masses will form in the tropics; on the other hand the area is perhaps best regarded as a single vast source region giving rise to but a single air mass, that is, equatorial air. Within this air mass the formation of fronts owing to differing surface influences is not possible to any marked degree and any fronts that may be carried into the tropics from higher latitudes tend to become smoothed out. Nevertheless, experience in the tropics, particularly flying experience, has shown that clouds and bad weather tend to occur along extended zones which can be represented as lines on synoptic charts with at least the superficial appearance of fronts. Surface and upper air observations have failed to detect significant changes of temperature and humidity across these lines such as would occur if they were true fronts. The conclusion is that they are, generally speaking, zones of convergence in which portions of the same air mass tend to flow together. The convergence leads to upward motion of the air within the zone and, in the moist and unstable conditions prevailing, convection cloud and precipitation follow; these can be violent if the convergence and instability are marked. But the front thus formed is a feature of the flow pattern; it is not to be identified with particular air masses and if the convergence 'relaxes' the front may disappear – sometimes to reappear simultaneously at a considerable distance away. In other words, the front is not necessarily an identifiable feature which can be followed from chart to chart as in temperate latitudes, and this lack of continuity is one reason why forecasting in the tropics is so difficult.

Other rather weak frontal zones appear from time to time in the tropics. The best known are the so-called 'easterly waves' which appear as shallow troughs of low pressure moving westward in the easterly wind streams on either side of the equator. As a rule they give rise only to increased cloud with a short period of rain or showers but on occasions squalls and thunderstorms develop. In West Africa they may give rise to heavy squalls and thunderstorms and are known as tornadoes.

These are of line-squall type and are not to be confused with the violent whirlwinds of the same name which occur in the United States of America and elsewhere (Section 119).

Intertropical front

The weather in the equatorial belt, within which the intertropical front is found, is described at length in Section 159, where it is stated that the active parts of the front at any one time consist of a number of detached sections. While the degree of activity of any part of the front on any particular occasion can be judged by a study of the synoptic charts, this affords little indication of future changes in the activity.

Tropical revolving storms

An account of these storms and the methods of detecting and avoiding them is given in Section 118 and the areas affected in Section 159.

Representation and forecasting of upper winds in the tropics

Since the geostrophic relationship between wind and pressure distribution is inapplicable near the equator, it is necessary to represent the wind field in the tropics

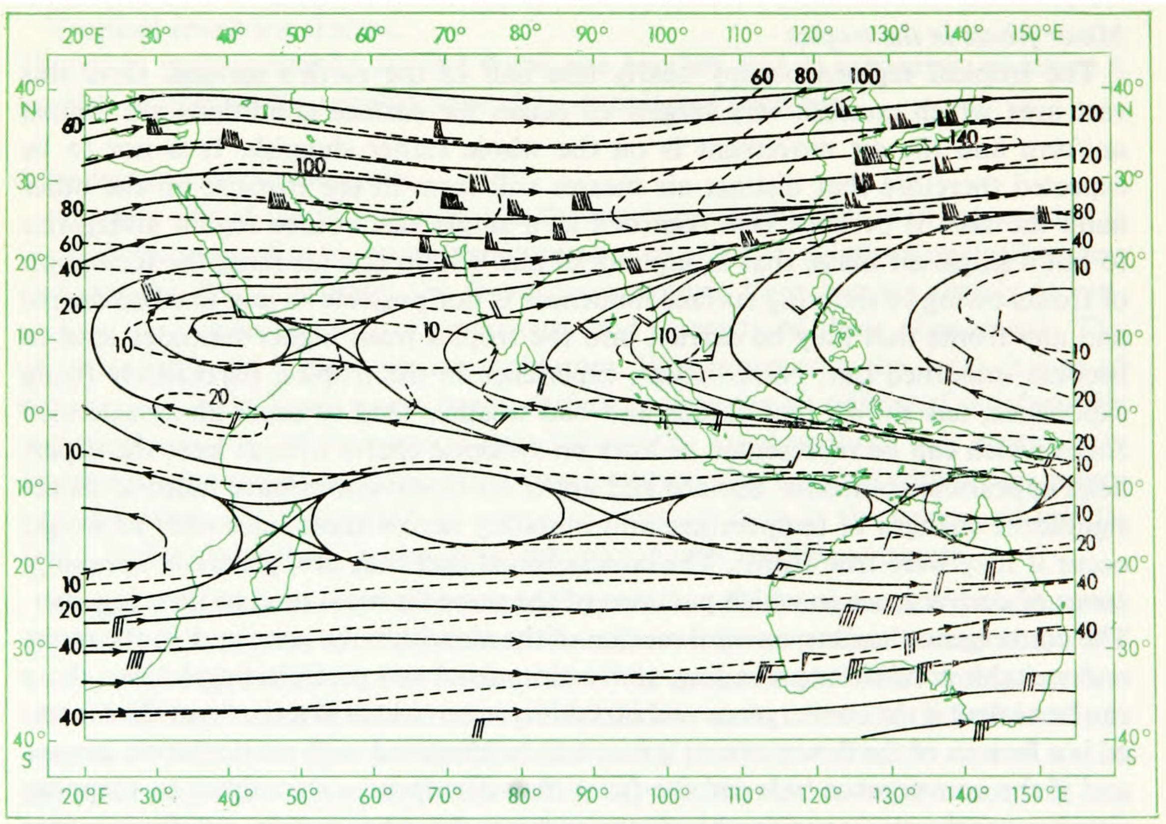


FIG. 105. *Streamlines and isotachs at 200 millibars, January (after Frost and Stephenson*)*

——— Streamlines — — — — Isotachs (knots)

Mean vector winds are represented by conventional symbols for wind. Wind arrows based on aircraft reports are pecked.

*FROST, B. A. and STEPHENSON, P. M.; Mean streamlines and isotachs at standard pressure levels over the Indian and west Pacific Oceans and adjacent land areas. *Geophys. Mem., London*, **14**, No. 109, 1965.

by means other than contours. The most useful method of mapping the wind field is by means of streamlines and isotachs. The streamlines are drawn so that the wind direction is everywhere tangential to them, as many or as few lines being drawn as are necessary to define the wind direction at every point on the chart. The isotachs are drawn to give the wind speed at any point. It is found that, unlike isobars which are always continuous curves, streamlines sometimes have to be drawn with free ends. They show vortex patterns with clockwise or counterclockwise flow indicating ridges of high pressure or troughs of low pressure respectively; neutral points appear and these correspond to cols in the isobaric analysis. Fig. 105 is an example of a streamline chart.

Forecasting is largely based on statistical similarity, but powerful or persistent trends may be extrapolated in the short term.

CHAPTER 20

EXAMPLES OF WEATHER ANALYSIS AND FORECASTING

139. INTRODUCTION

The greatest difficulty which arises in presenting in a textbook a course of practical synoptic meteorology is occasioned by the almost unlimited diversity of weather types with differences of real significance. The weather associated with depressions or anticyclones, with warm and cold fronts, with occlusions and with air masses, cannot be disposed of by single examples, but for any approximation to completeness numerous illustrations must be given. Indeed, nothing can take the place of years of experience with successive weather charts if the proper perspective is to be obtained, if general rules are to be applied with confidence, and if the indications of unusual developments are to be given their proper weight.

In this chapter a few short periods are discussed. In each case the general development is described and attention is drawn to the analysis of the charts and to any minor points of interest. The more complicated types of chart have been avoided but no attempt has been made to search through long records in order to find what are miscalled 'typical examples', a euphemism for those cases which conveniently illustrate the theories. Only very rarely, if ever, will a reader come across a weather map where the conditions associated with air masses and fronts show in every detail their 'characteristic properties', and should he do so the interpretation of the chart will offer no difficulty.

From any period chosen at random valuable lessons are to be learned; apart from showing conveniently placed fronts and important types of adverse flying weather, the cases here discussed are in no way peculiar. They will serve to illustrate the sort of material upon which the forecaster bases his work and the methods by which the information is analysed and interpreted.

There are obvious limitations to the possibility of reproducing synoptic charts on the printed page. Detailed charts, plotted by the full international symbolic method are accordingly given only for a restricted area, while the general situation and its development are indicated on supplementary charts of smaller scale where only isobars and fronts together with a few observations of wind, temperature and weather, can be reproduced. In accordance with the usual practice the pressure systems and fronts are marked with identifying letters which they retain throughout their history. Two letters are used; the first indicates the type of feature thus: L, depression (low); H, anticyclone (high); W, warm front; C, cold front; O, occlusion. The second letter is an identification symbol for the particular system.

Some of the situations are further illustrated by upper air charts, forecast charts and tephigrams.

The key map (Fig. 106) shows the names and positions of the stations from which reports are plotted in the detailed charts.

140. PERIOD: 17–18 JANUARY 1967

A deep depression in mid-Atlantic is moving east and an associated trough with warm and cold fronts moves across the British Isles.

FIG. 106. *Key to reporting stations*

Upper air stations are underlined.

The surface chart for 1200 GMT on 17 January 1967 (Fig. 107) shows a weakening trough just to the west of Scotland and Ireland moving east, followed by a ridge of high pressure with its axis at this time at about 20°W. A deep depression LX in mid-Atlantic is moving east. This centre is linked by an occlusion to another depression in the Davis Straits west of Greenland.

At 500 millibars the 1200 GMT chart on 17 January (Fig. 108) shows a ridge over Europe, a trough to the west of the British Isles and a marked trough over the western Atlantic.

The 200-millibar chart (Fig. 109) for the same time shows a similar pattern with very strong winds over the western Atlantic. Comparison with previous charts showed a general eastward progression of the trough with a strong westerly wind belt developing about 40°N. The upper flow pattern suggests that the eastward movement of LX will continue but the centre in the Davis Straits will remain stationary and fill up. Fig. 110 is the forecast surface chart for 1200 GMT on 18 January, 24 hours ahead of the actual chart. It was prepared by the forecaster at the Central Forecasting Office and issued at 1600 GMT on 17 January. The depression LX is at 53½°N 17°W and the trough with occluded fronts lies off the west coast of Ireland. The ridge, less well marked, has its axis over England. Fig. 111 is the forecast chart for the same period based on the same observations but produced by the computer. The actual chart for 1200 GMT on 18 January is shown in Fig. 112. The position of the centres of high and low pressure and the general pattern of the isobars on all three charts are in reasonable agreement, but the centre LX is rather too far east on the forecaster's chart. The depression is less occluded than was forecast.

The more detailed weather associated with the fronts can be followed on the charts for 1800 GMT on the 18th and 0000 GMT on the 19th (Figs 113 and 114). These show how the warm and cold fronts move across England with a general deterioration of weather, cloud base and visibility.

The anemogram for Valley (Section 25, Fig. 19) for the period 1200–2000 GMT on 18 January shows how the wind speed increased in the tightening gradient ahead of the warm front. Fig. 75 (Section 84) shows the cloud base as recorded at London/Heathrow Airport from 1900 to 2100 GMT on 18 January.

The significant weather chart issued by Heathrow for 0000 GMT on the 18th is shown in Fig. 115.

141. PERIOD: 9–10 FEBRUARY 1953

General development

The illustrative charts for this period, Figs 116–120, show the development of a deep depression over southern Ireland and its subsequent eastward movement across Wales and England. At 1800 on 9 February (Fig. 116) a large area of low pressure is situated to the north-west of the British Isles with two distinct centres LM and LN each with central pressure below 976 millibars. LM is an old frontal depression but the situation of LN far from any front suggests that it is a polar-air depression and this is confirmed by a reference to earlier charts. LM is well occluded and LN has no frontal structure so that a slow filling of these centres is to be expected. On the other hand there appears to be a developing secondary LA off western Ireland. It has not yet developed a separate circulation but the large falling tendencies in south and west Ireland suggest that it will soon do so. Further evidence of the approach of a secondary is furnished by the reports of rain in south-west Ireland

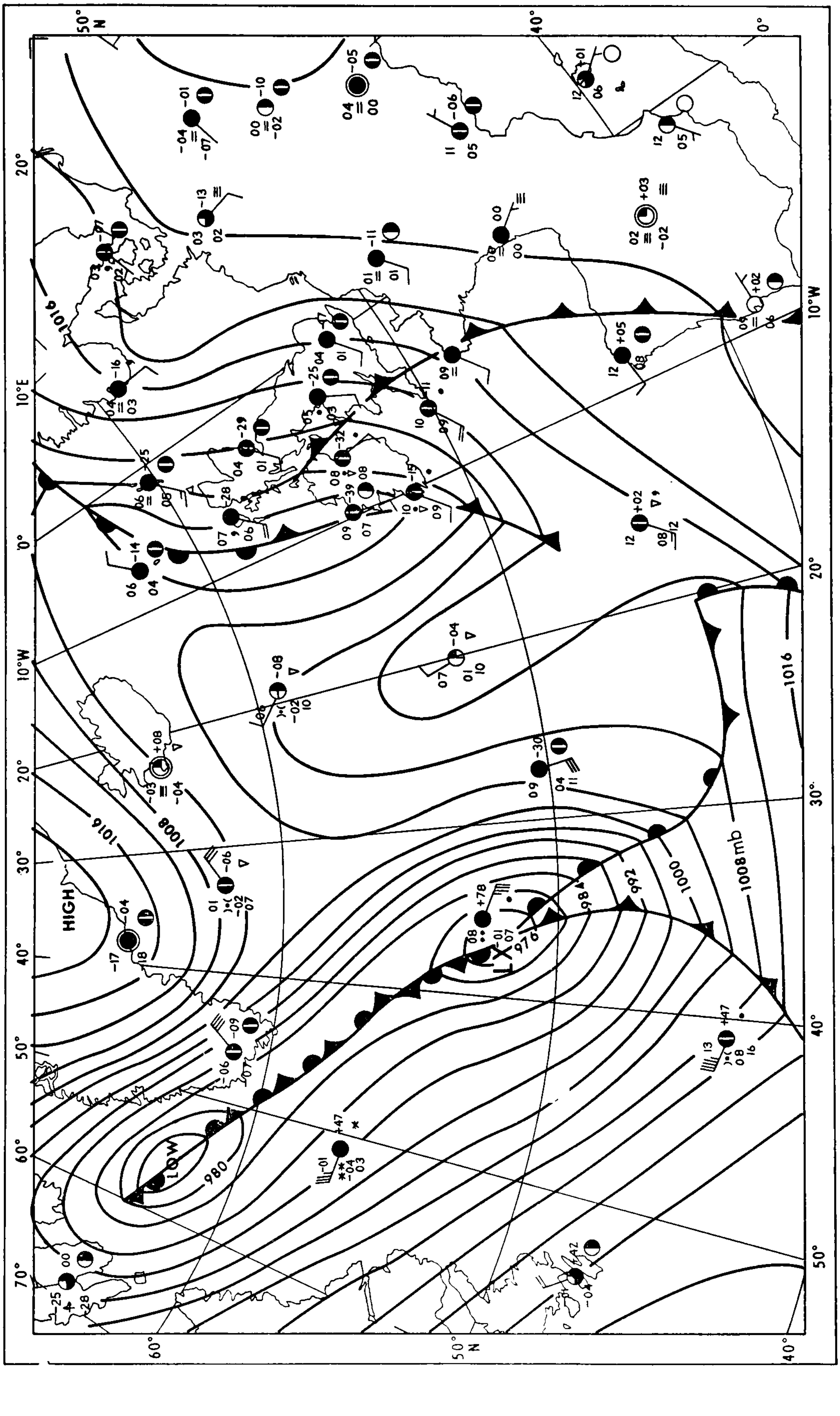


FIG. 107. *Surface chart for 1200 GMT, 17 January 1967*

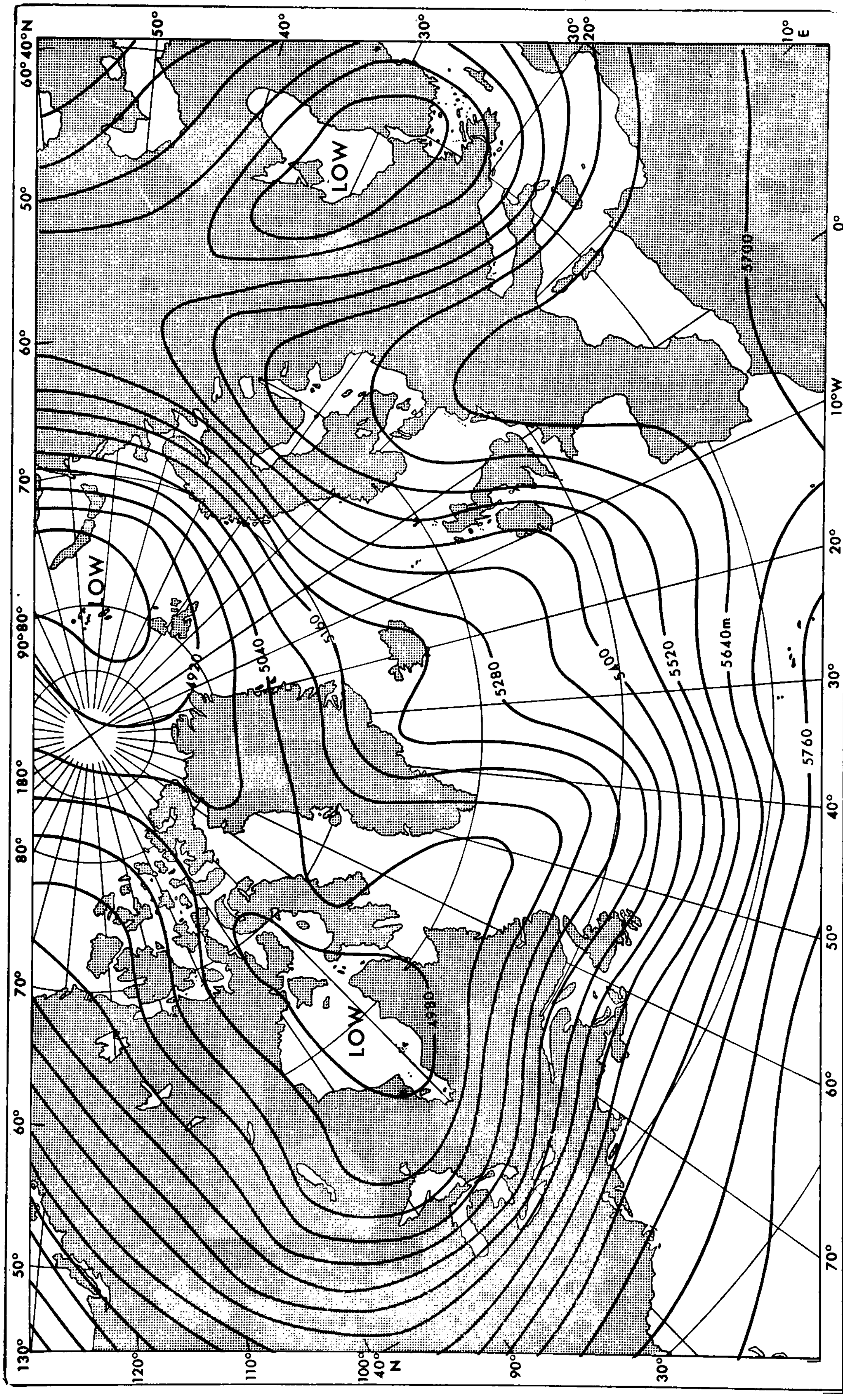


FIG. 108. *Contour chart for 500 millibars, 1200 GMT, 17 January 1967*

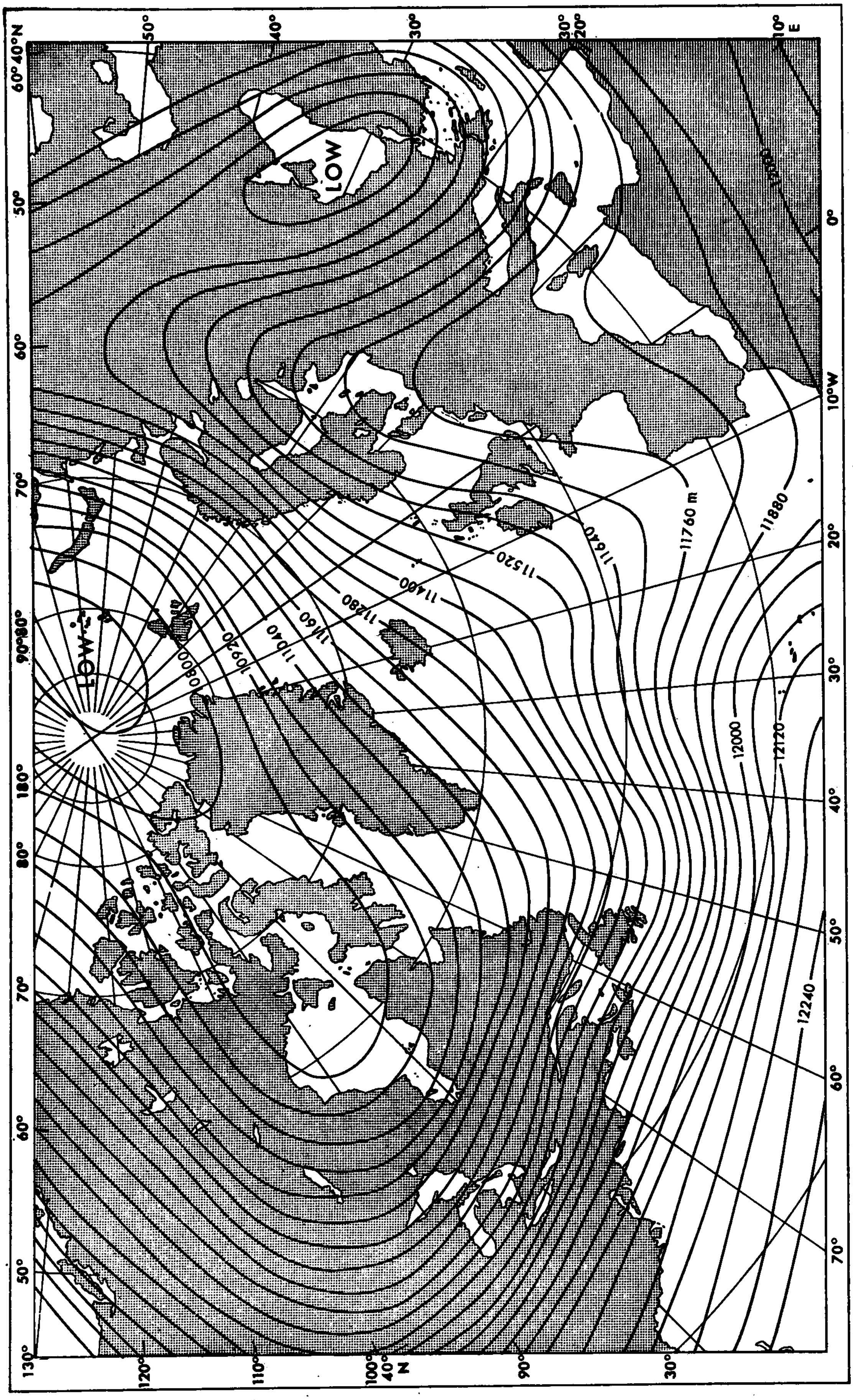


FIG. 109. *Contour chart for 200 millibars, 1200 GMT, 17 January 1967*

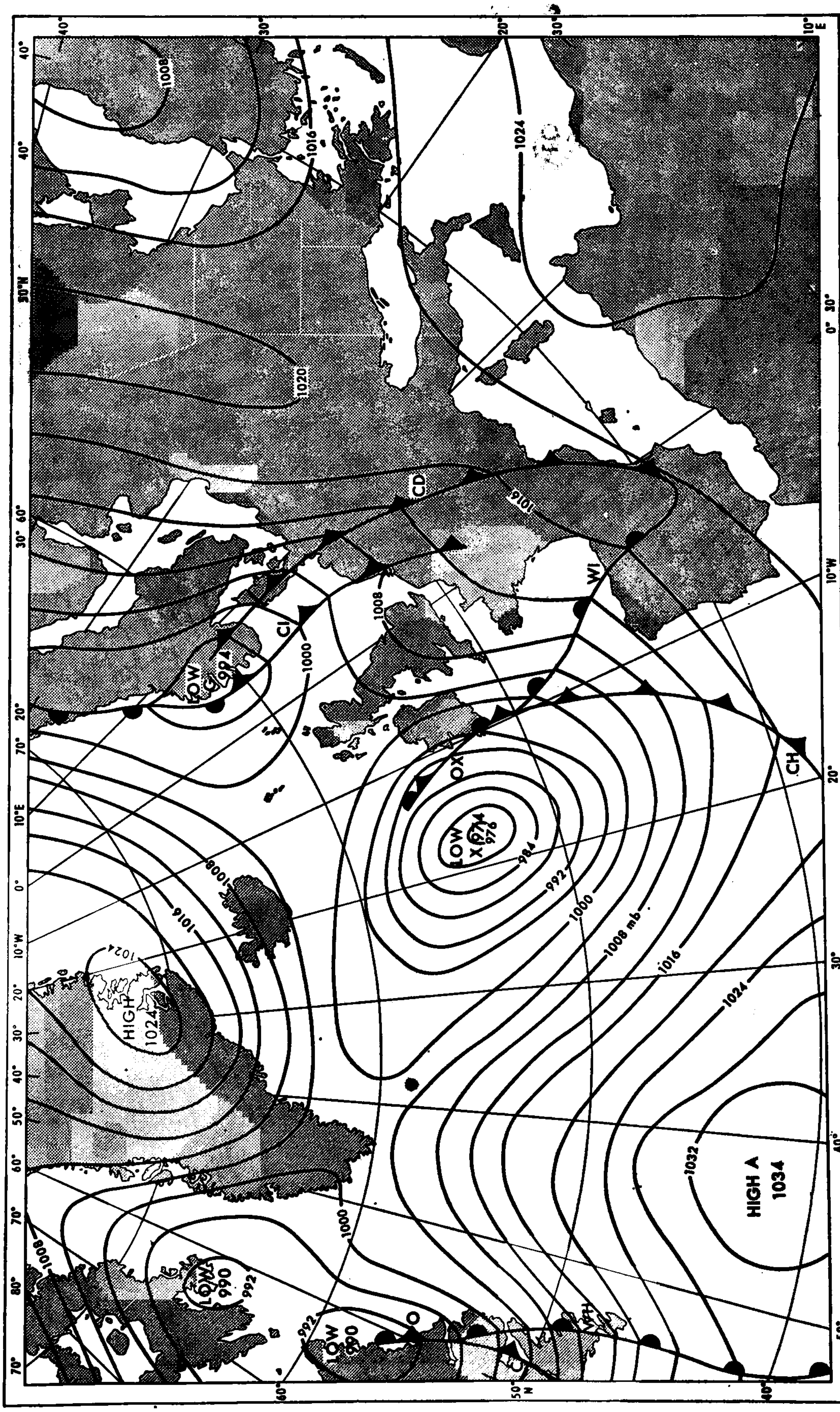


FIG. 110. Forecast surface chart for 1200 GMT, 18 January 1967

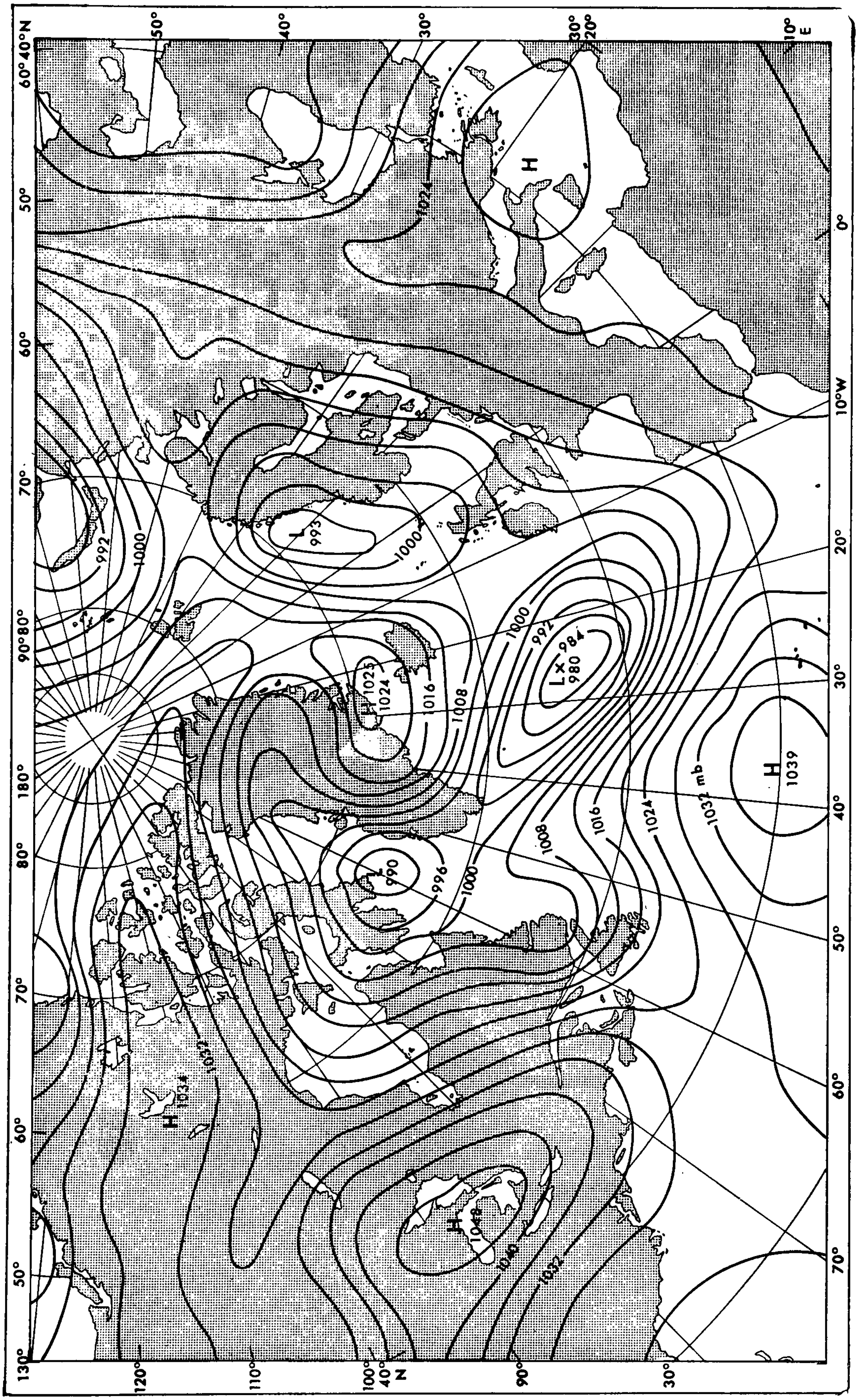


FIG. 111. *Computed forecast chart for 1200 GMT, 18 January 1967*

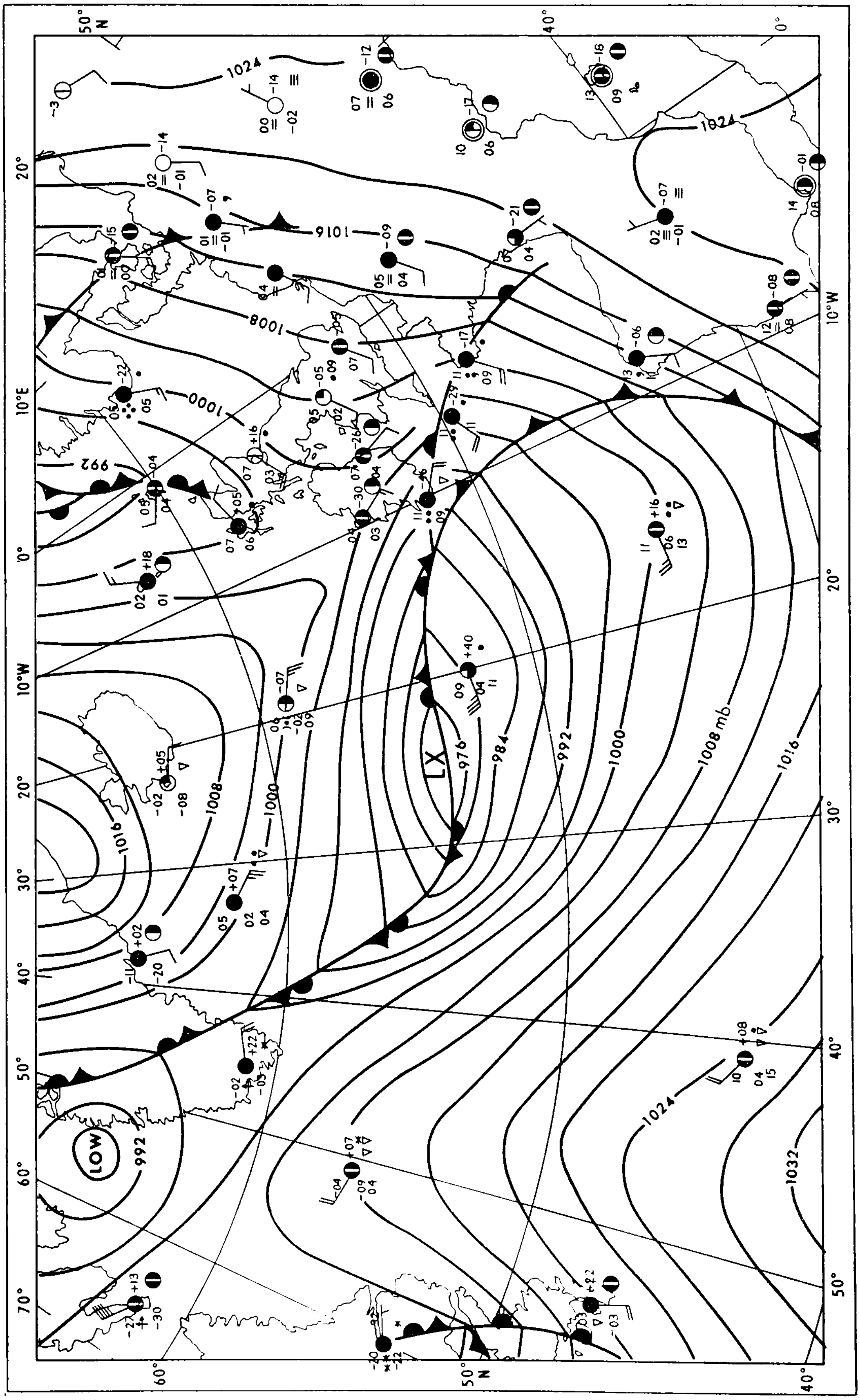


FIG. 112. *Surface chart for 1200 GMT, 18 January 1967*

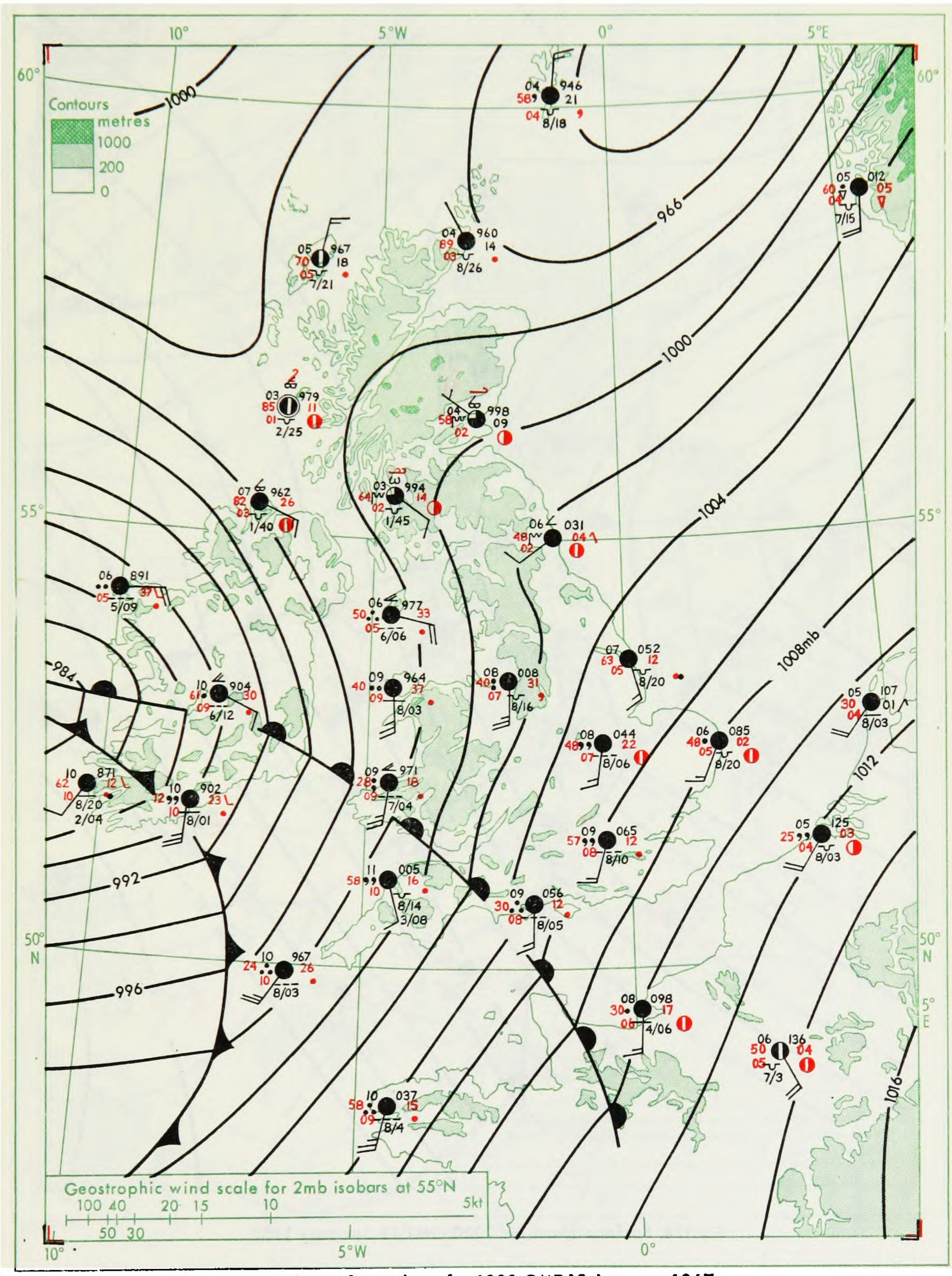


FIG.113. Surface chart for 1800 GMT, 18 January 1967

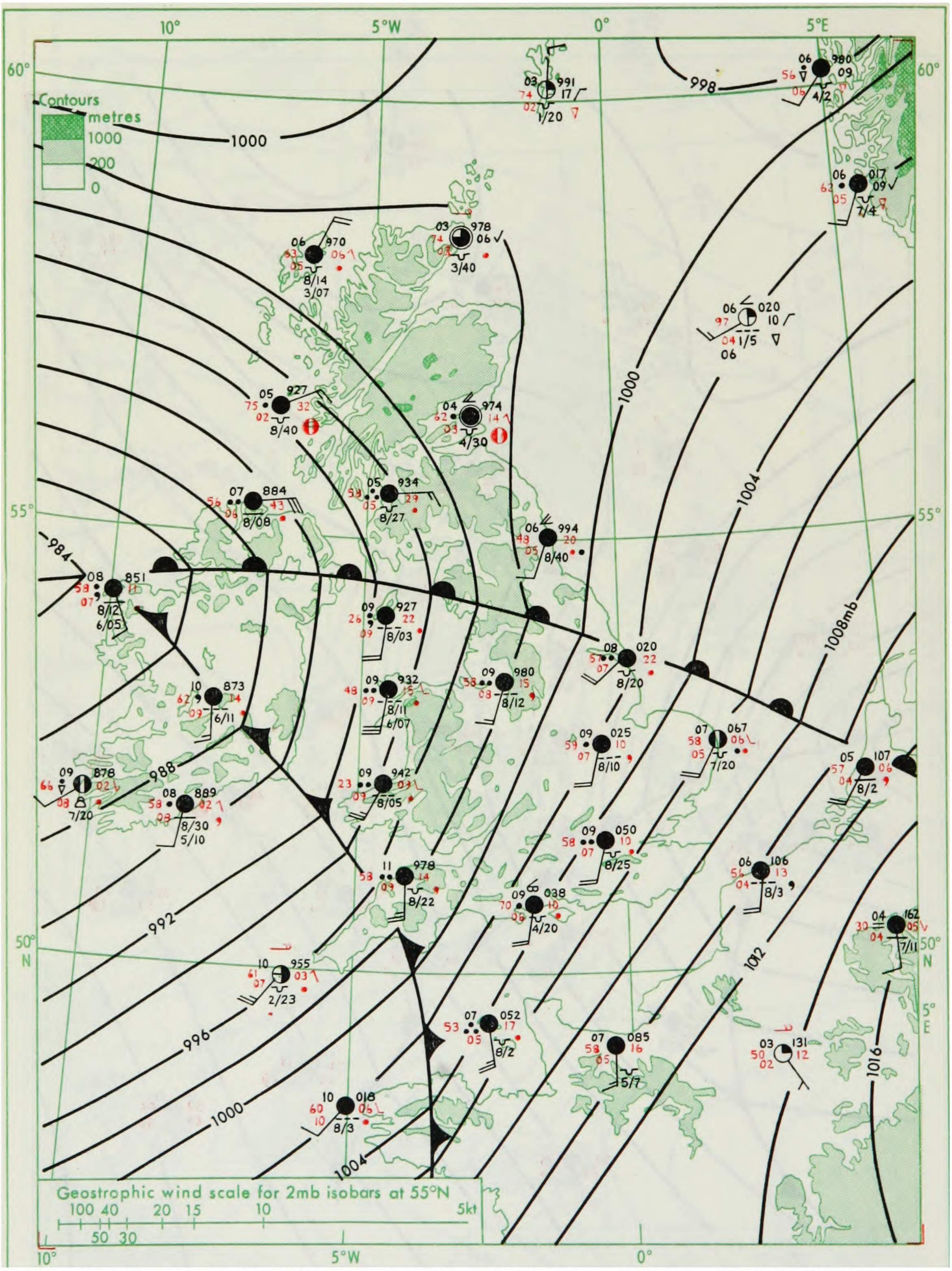
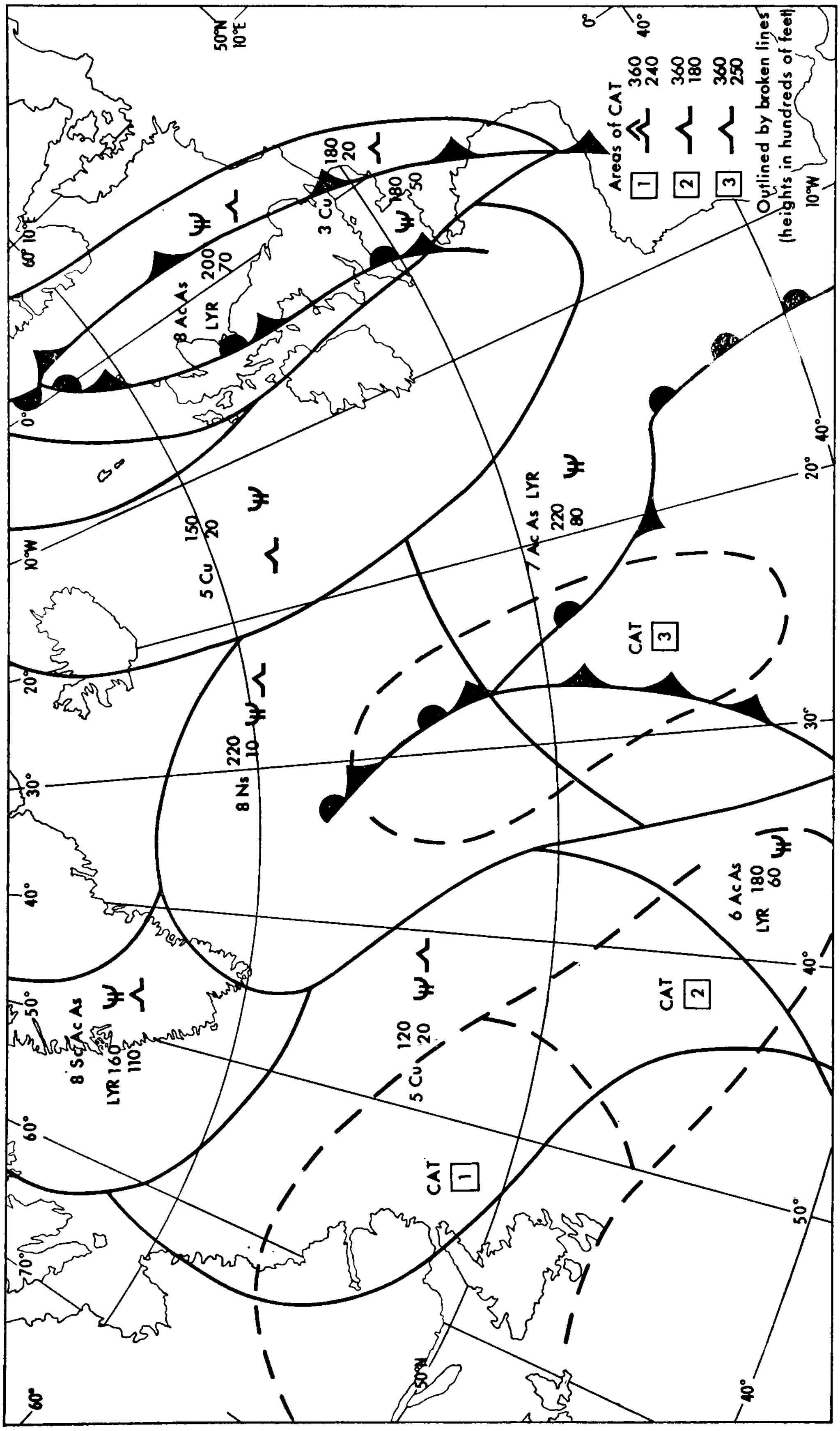


FIG.114. Surface chart for 0000 GMT, 19 January 1967



CAT = Clear-air turbulence LYR = Layer or layered ψ = Moderate airframe icing = Severe turbulence = Moderate turbulence

FIG. 115. *Significant-weather chart for 0000 GMT, 18 January 1967*

and the backing of the winds over Ireland in comparison with those over England and Wales. The new system LA is analysed with a warm sector extending right into the centre; the analysis is based largely on continuity with the preceding charts but is consistent also with Fig. 116 itself. The veering wind and higher temperature at Valentia as compared with Cork, Shannon and Belmullet place the warm front WA near or past Valentia; the change of wind direction places the cold front CB north-west of the ship P and south-east of the ships Q and R. North-westerly winds and showers with steadily lowering temperatures towards the north-west

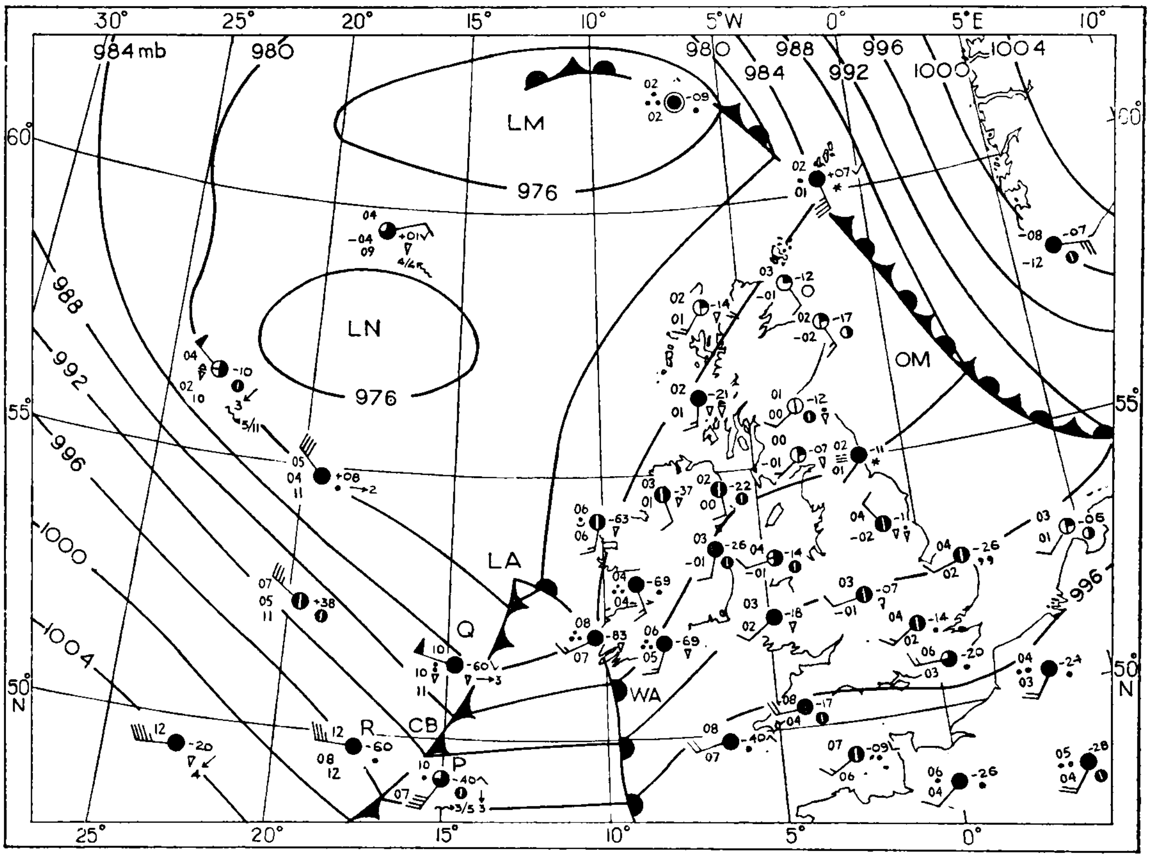


FIG. 116. *Surface chart for 1800 GMT, 9 February 1953*

are indicative of a polar airstream behind CB. Twenty-four hours later (Fig. 117) LA appears as the dominating system near the British Isles; it has in fact taken over the role of principal centre having deepened by about 15 millibars and developed an extensive and vigorous circulation of its own; meanwhile the centres LM and LN have filled and degenerated into a subsidiary trough. The warm sector of LA has been occluded and the occlusion OA spirals outwards from the centre; the temperatures show that LA is now completely surrounded by cold air.

The 500-millibar contours for about the middle of this period are shown in Fig. 120. The large rather flat area of low pressure to the north-west of the British Isles (Fig. 117) appears also at the 500-millibar level; this is usual with old occluded depressions which have cold air all round their centres. The small closed low over south-east Ireland in Fig. 120 is due to the upward extension of the deep surface low (Fig. 118). These small closed centres apart, the most prominent feature on this chart is the long-wave pattern with a ridge on the west Atlantic and trough over the British Isles and western Europe. From mid-Atlantic to the Bay of Biscay the close contour spacing indicates the probably existence of a jet stream but actual wind observations are scanty in this area and the only confirmation is the wind of 80 knots off north-west France. Over France and southern England the contours are diffluent and the deepening of the low LA at the left exit to the jet is in accord with the ideas of thermal development outlined in Section 131.

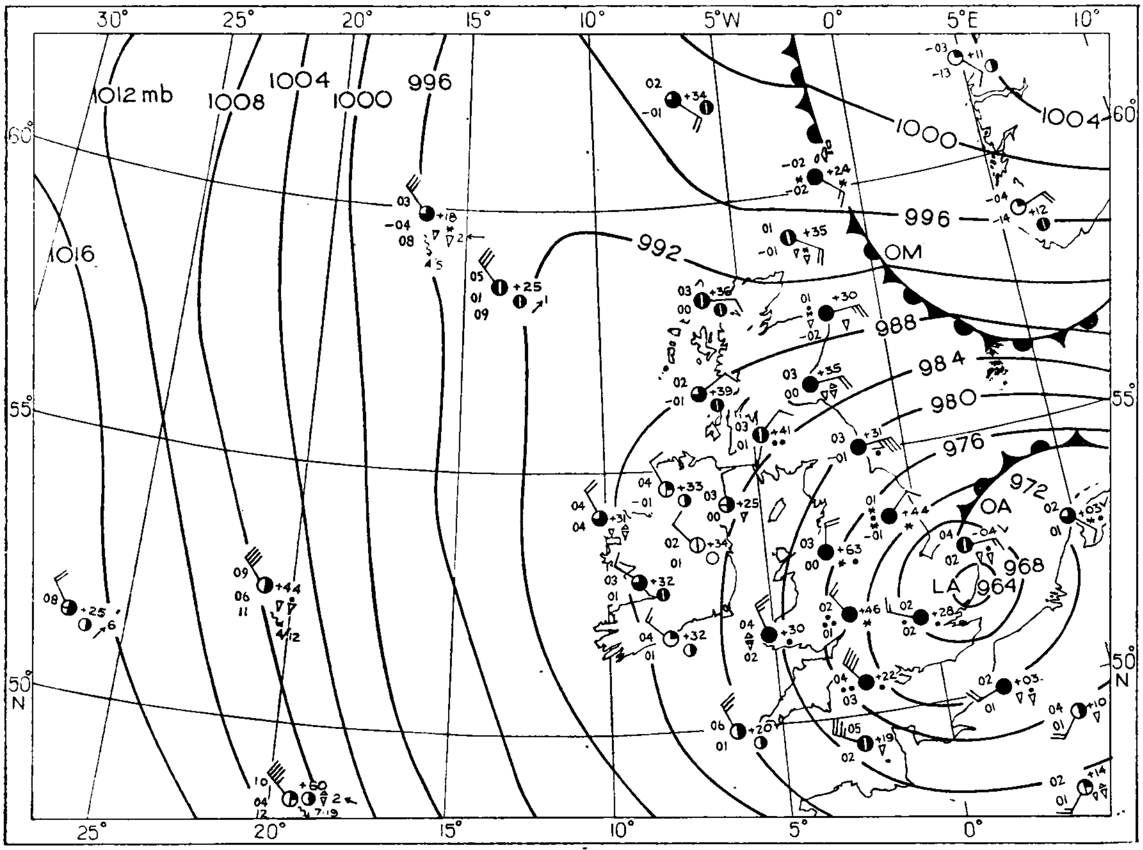


FIG. 117. *Surface chart for 1800 GMT, 10 February 1953*

Detailed analysis

The situation for 0000 on 10 February is shown in detail in Fig. 118. It is obvious from the reported pressures that a substantial depression is centred over south-east Ireland; it now has a closed circulation with a definite zone of easterly winds to north of the centre. There are large negative tendencies to north, east and south of the centre; the largest tendencies, including —114 at Pembroke Dock, occur over Wales and western England and indicate that the centre may be expected to move eastward. Behind the centre, in the extreme west of Ireland the tendencies show that pressure is just beginning to rise; as these positive tendencies are much smaller than the negative tendencies ahead of the centre, it is concluded that the depression is deepening.

The line of the warm front WA has moved into France and south-west England and continues into the centre of the low as an occlusion OA. Ahead of this front the winds at most stations have backed to south-east; there is a broad belt of rain (or snow) extending to Clones, Valley, Hurn, Abbeville and Paris. Far in advance of the front, as at Gorleston, Spurn Head and Tynemouth there is no low cloud or only stratocumulus; somewhat nearer the front there is cirriform cloud (at West Freugh) and altostratus (as at Blackpool, Cardington and Manston) while within about 100 miles of the front most stations report eight oktas of nimbostratus or low clouds of bad weather. Finally the tendencies, already referred to, complete a picture which illustrates in practically every respect the characteristic pre-warm-front conditions listed in Chapter 16.

The precise line of the warm front is located by the continuous rain at Pembroke Dock, Exeter, Guernsey and Nantes and by the wind directions, south-east at Pembroke Dock and Guernsey and south-west at Exeter, Rennes and Nantes. A rise of temperature, and still more, of dew-point is to be noted across the front from east to west. The low cloud and high dew-points reported at Brest and Le Talut are typical of a warm sector; the continuous rain and large negative tendencies at these stations are however less usual and point to continued deepening of the depression.

The cold front CB is not difficult to place; the cessation of the rain and the westerly wind and low dew-point at Scilly show that it has certainly passed that station; the rain, westerly wind and recent shower with thunder suggest that it is very close to the ship S. On the other hand St Eval still appears to be in the warm sector. The drawing of the front through ship S and between Scilly and St Eval requires it to link up with the warm front somewhere near Pembroke Dock, with the occlusion OA running from there into the centre of the depression.

At 0600 (Fig. 119) the tendencies show a roughly similar distribution with respect to the centre LA as at 0000, indicating continued eastward movement; this is confirmed by Fig. 117. The position of the centre over Wales is easily fixed by the reported pressures and winds. The line of the occlusion OA and warm front WA is also fairly easy to recognize; precipitation, cloud, wind direction, tendency, temperature and dew-point at Dishforth, Manby, Gorleston, Flushing, St Quentin and Paris place these stations ahead of OA–WA. Westerly winds indicate that the front has passed Lympne, Rouen and Tours. Of these three stations, however, the only one suggesting warm-sector air is Tours, with temperature 8° and dew-point 7°C; Lympne and Rouen both have dew-points 4°C or below indicating that the cold front also has passed them. The point of occlusion therefore appears to be somewhere near Abbeville, and the whole of southern England and north-west France are in the polar air which has swept round the western and southern flanks of LA.

Special points

Snow is occurring at a number of stations to northward of the centre LA in both Figs 118 and 119; it is associated with surface temperatures only a little above freezing-point and affords a good illustration of the rule that if the temperature is near freezing snow should be forecast in situations which would otherwise give rain.

In Fig. 119 some stations fairly near to the centre of the depression are still reporting continuous precipitation though they do not appear to be under the influence of the fronts, for example, Ronaldsway, Dublin, Pembroke Dock and Exeter. Farther away from the centre at Belmullet, Valentia, Cork, Scilly and Brest there are the broken cloud and showers typical of polar air. There is usually some such prolongation of the continuous rain in the rear of the centre of a moving depression; if the depression is moving slowly this rain may be very persistent. Its time of cessation is difficult to forecast since it has not the more or less clear-cut edge that is often associated with frontal rain areas.

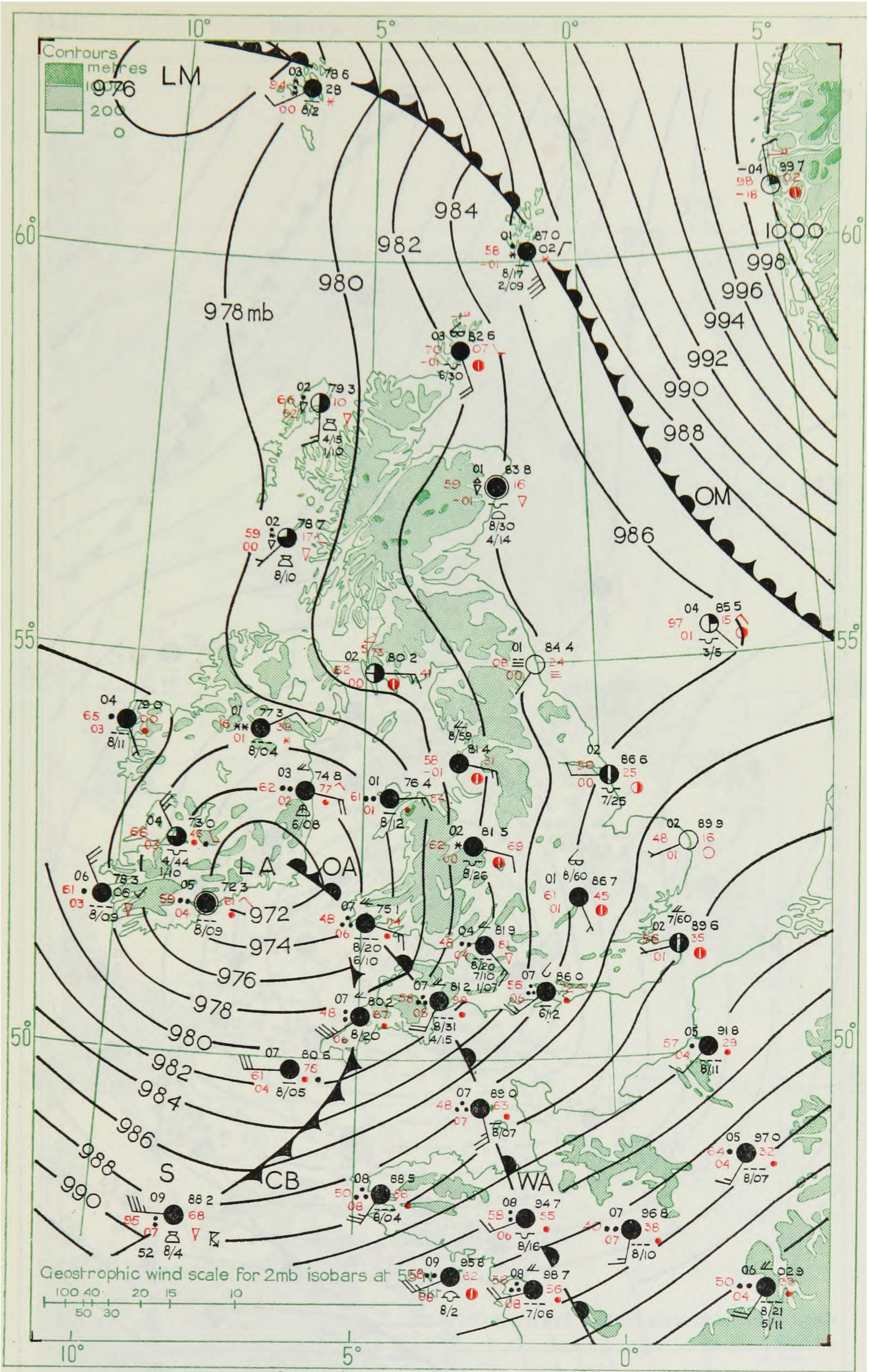


FIG.118. Surface chart for 0000 GMT,10 February 1953

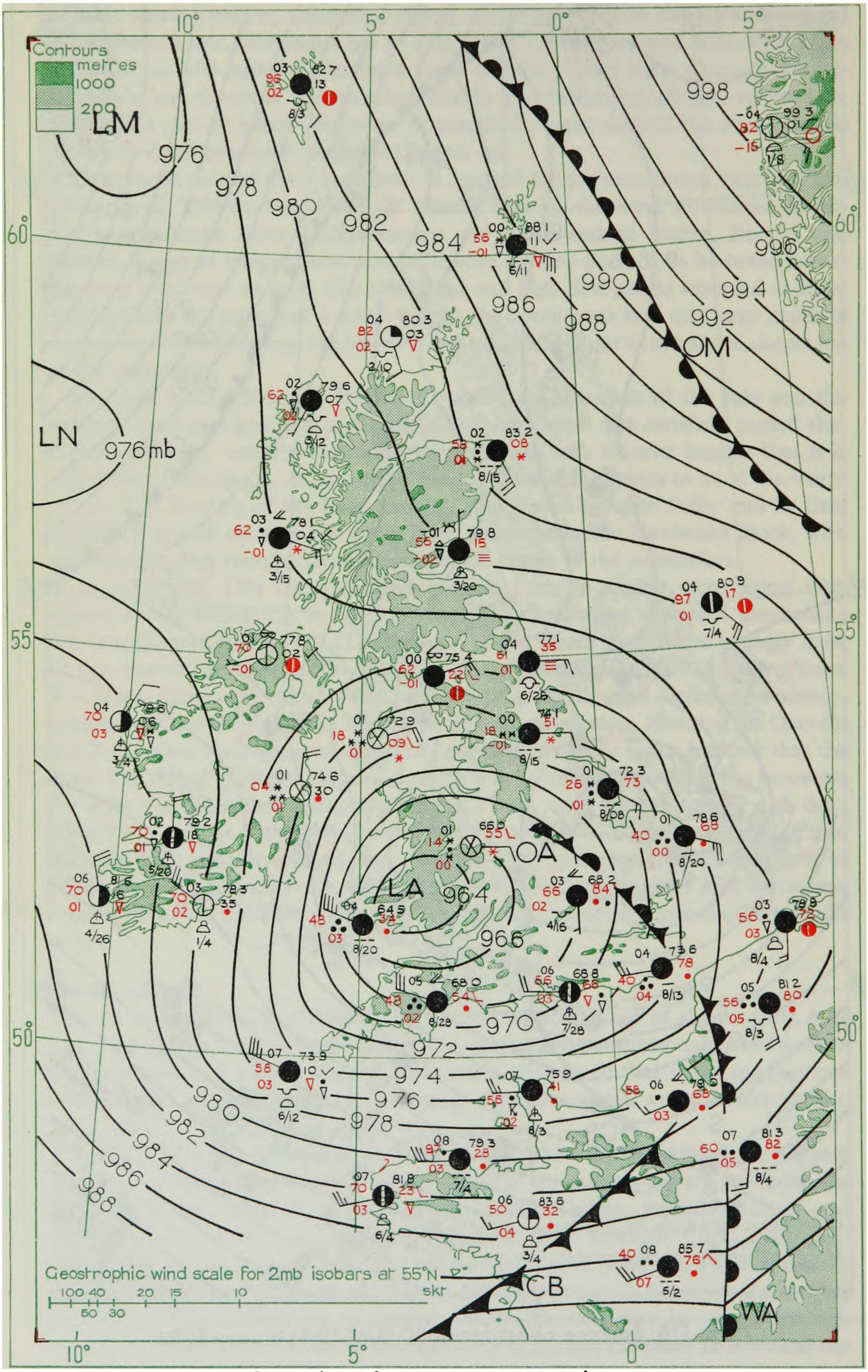


FIG.119. Surface chart for 0600 GMT,10 February 1953

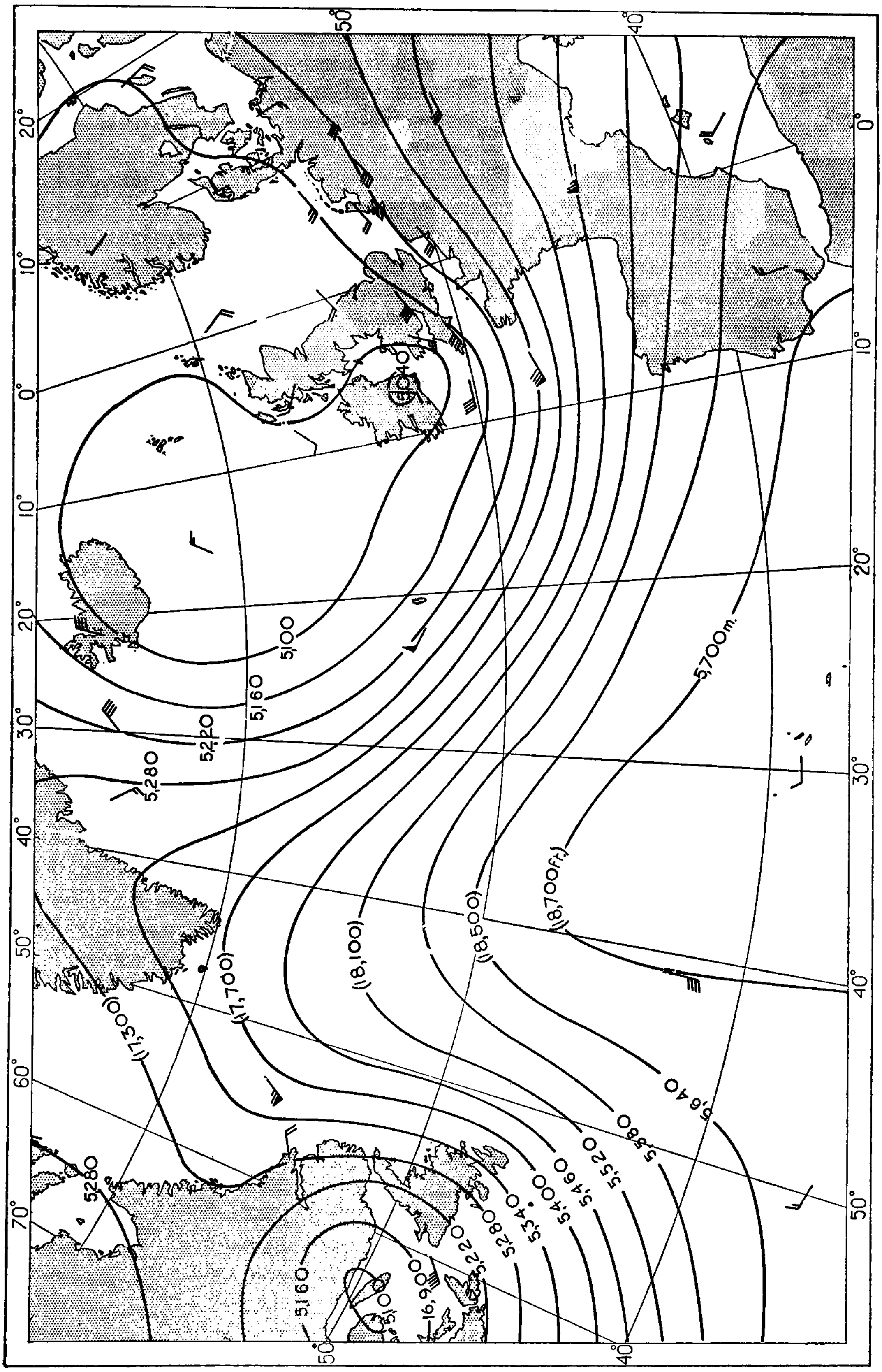


FIG. 120. *Contour chart for 500 millibars, 0300 GMT, 10 February 1953*

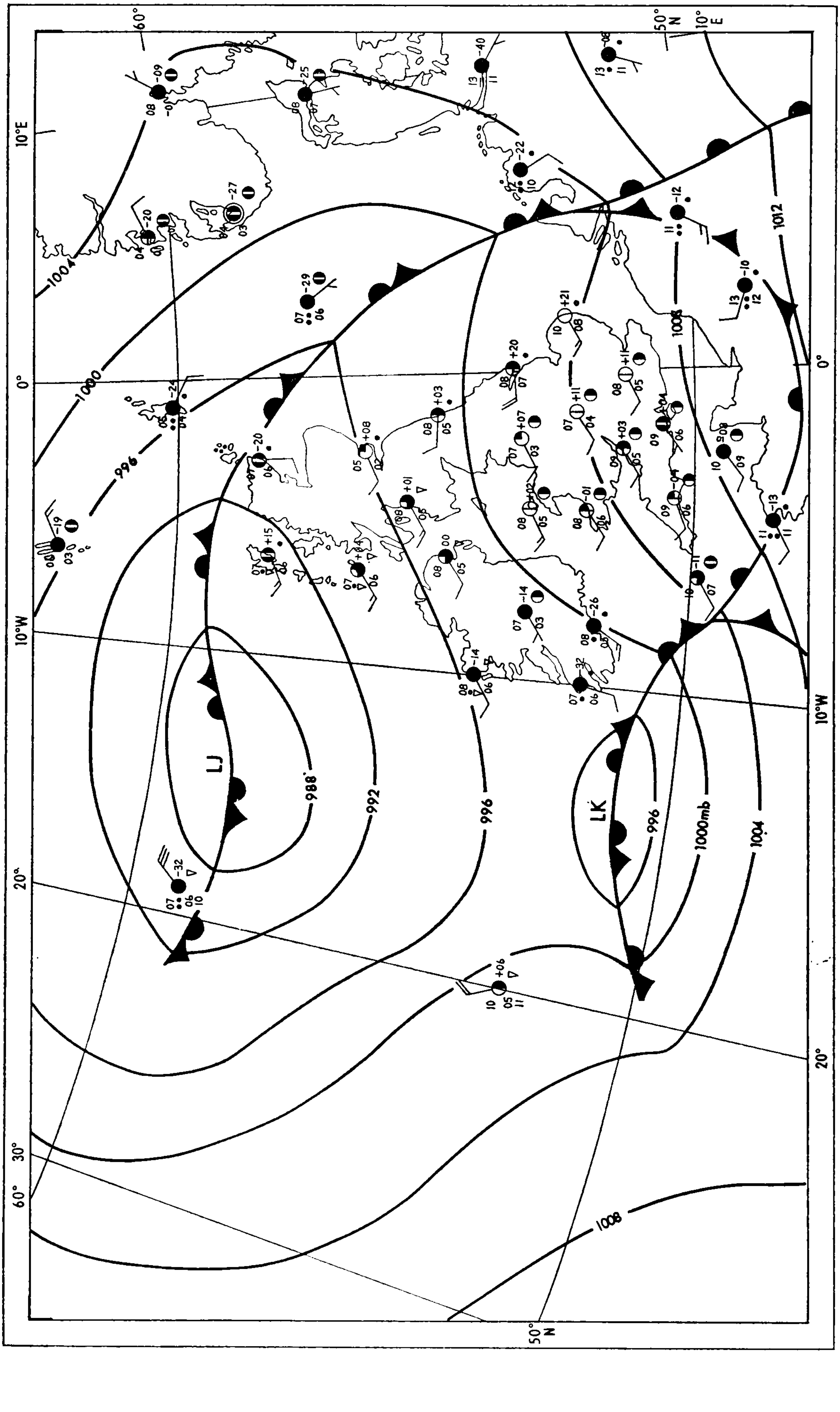


FIG. 121. *Surface chart for 0000 GMT, 5 May 1966*

142. PERIOD: 5 MAY 1966

This situation illustrates the contribution that satellite pictures can make to analysis and forecasting. It is described in greater detail in an article in the *Meteorological Magazine.**

The chart for 0000 GMT on 5 May 1966 (Fig. 121) shows off south-west Ireland a small secondary depression LK which is moving east. It is linked by fronts (occluded, warm and cold) to the main depression LJ off north-west Scotland. The 1000–500-millibar thickness pattern for 0000 GMT on 5 May (Fig. 122) shows a moderate thermal gradient over northern France and southern England but little or no gradient over Scotland and northern England. On the chart for 0600 on the

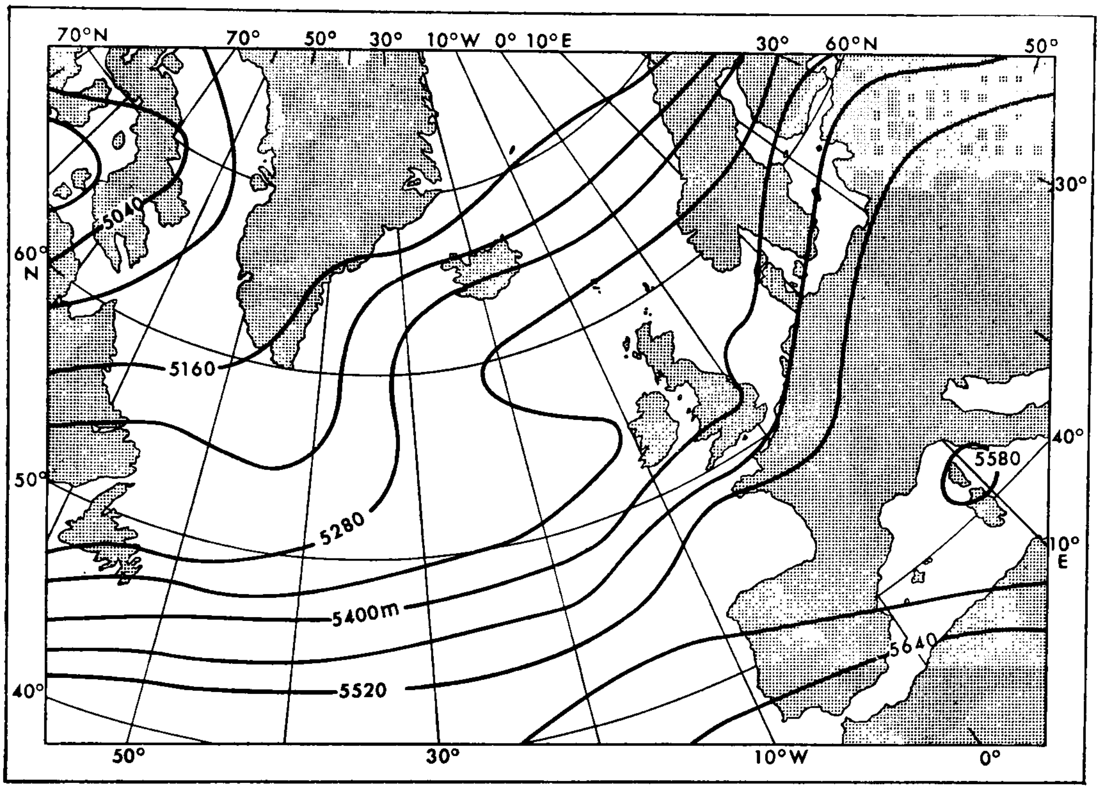


FIG. 122. *Thickness chart for 1000–500 millibars, 0000 GMT, 5 May 1966*

5th (Fig. 123) the occlusion has moved across south-west Ireland and south-west England; LJ has remained stationary, whilst LK has moved north-eastwards into south-west Ireland.

Fig. 124 shows a cloud analysis for 0950 GMT, 5 May, which is based on the satellite cloud picture centred over southern Scotland at that time (Plate XXX). The position of the cloud vortex and the general cloud pattern suggest that the main centre LJ was further south than had previously been drawn; consequently the chart for 1200 GMT on the 5th (Fig. 125) was drawn with the centre near the position of the vortex and with a tighter gradient on its western flank. A ship's report confirms the new position. The boundary of the cirriform cloud near the jet stream at the southern boundary of the satellite picture indicates that the jet stream shown on the 300-millibar chart, Fig. 126, at 46°N 12°W should probably be further south.

*See footnote on p. 297.

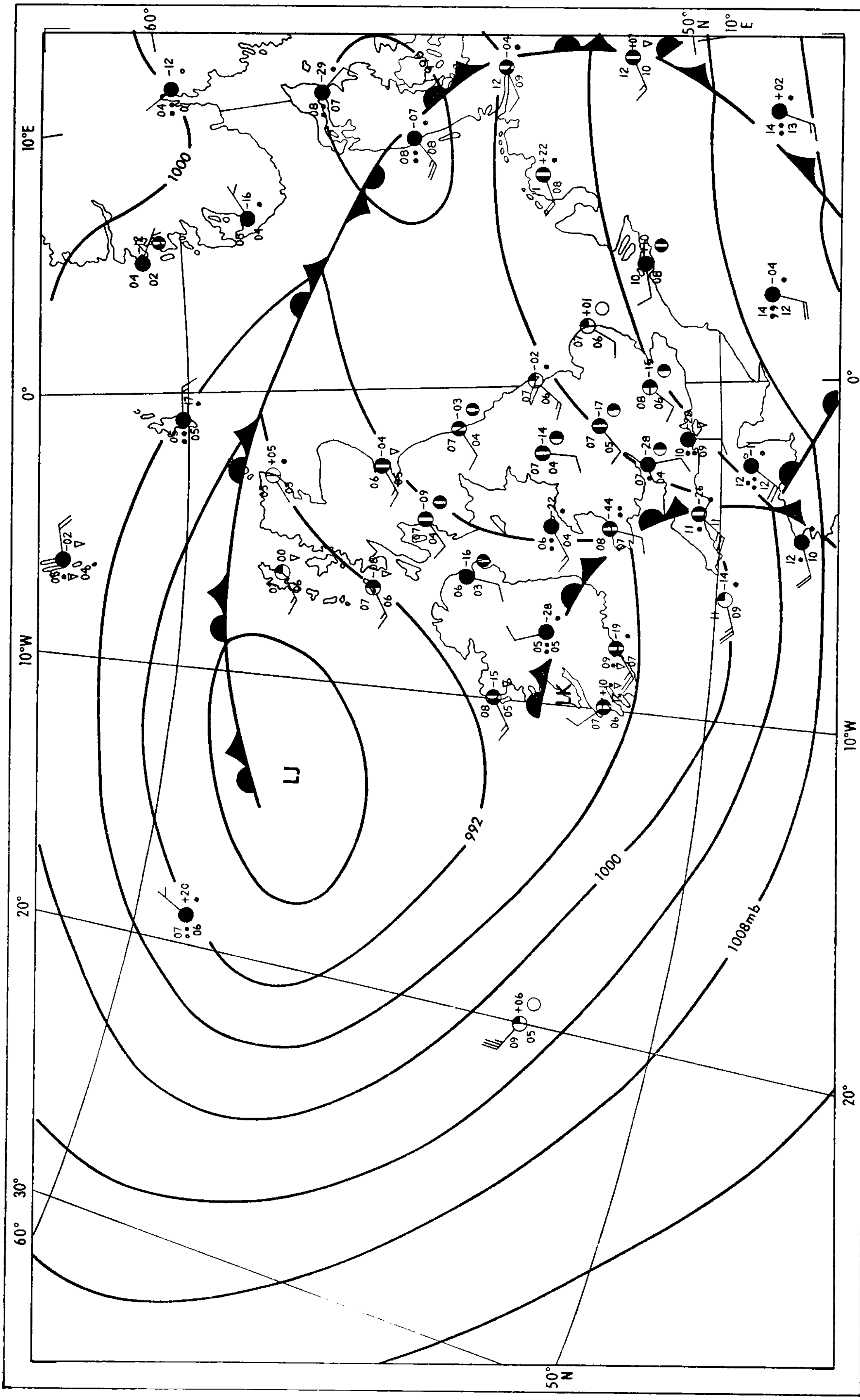


FIG. 123. *Surface chart for 0600 GMT, 5 May 1966*

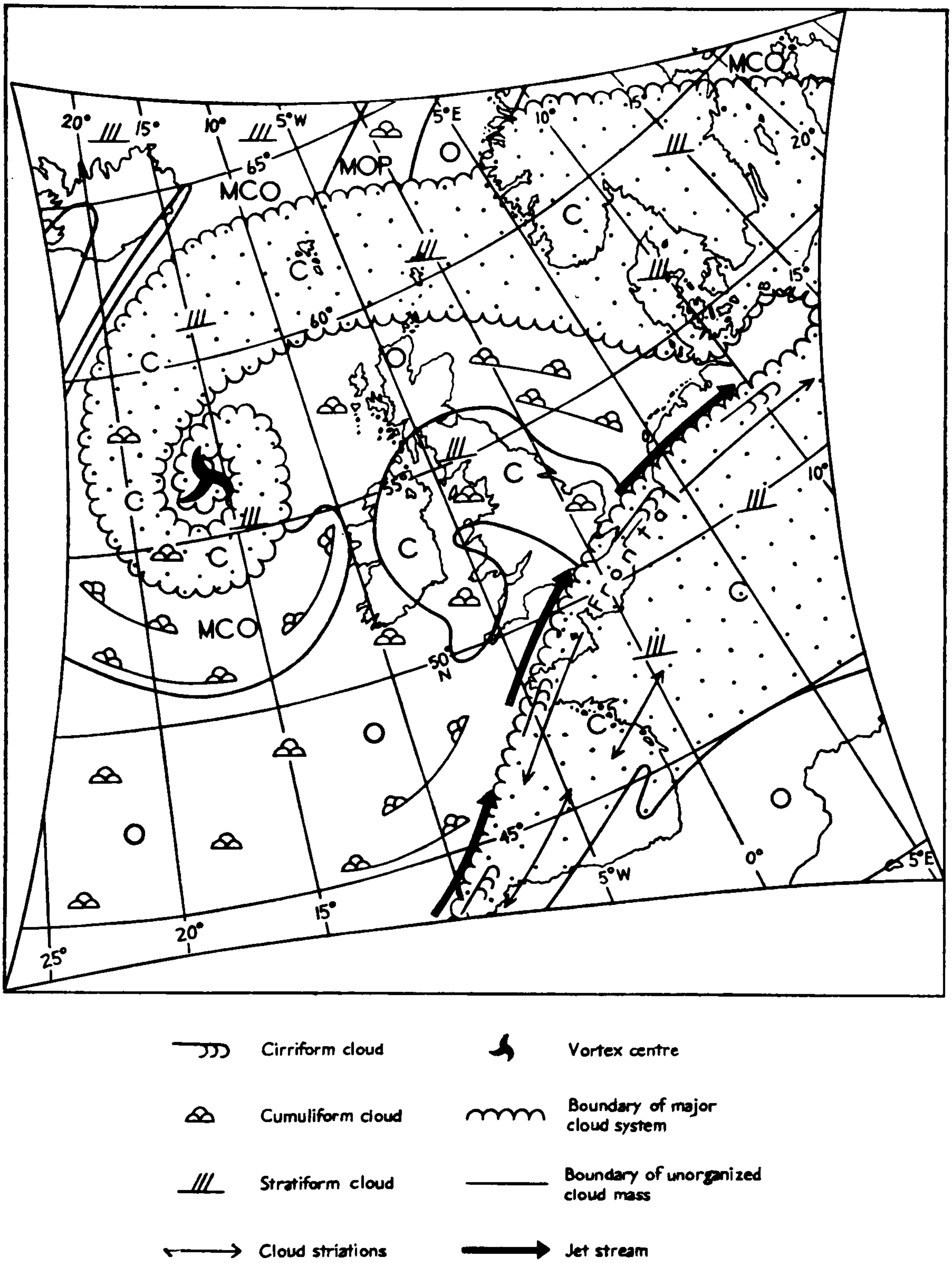


FIG. 124. *Nephanalysis based on satellite television picture, 0950 GMT, 5 May 1966* (after Pothecary and Ratcliffe*)

Stippling indicates cloud organization considered to be synoptically significant

O = < 20 per cent coverage MOP = 20–50 per cent coverage

MCO = 50–80 per cent coverage C = > 80 per cent coverage

*See footnote on p. 297.

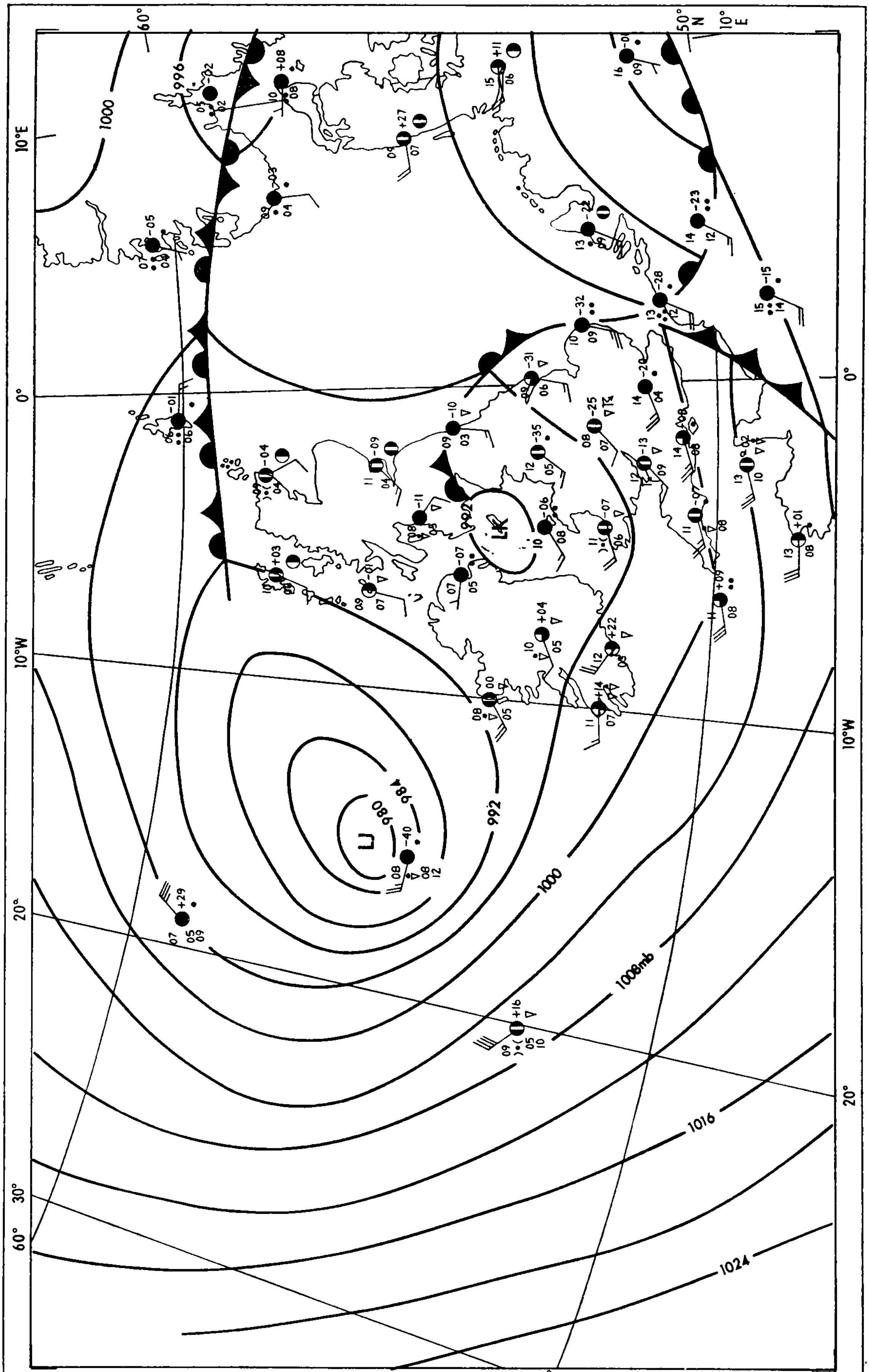


FIG. 125. *Surface chart for 1200 GMT, 5 May 1966*

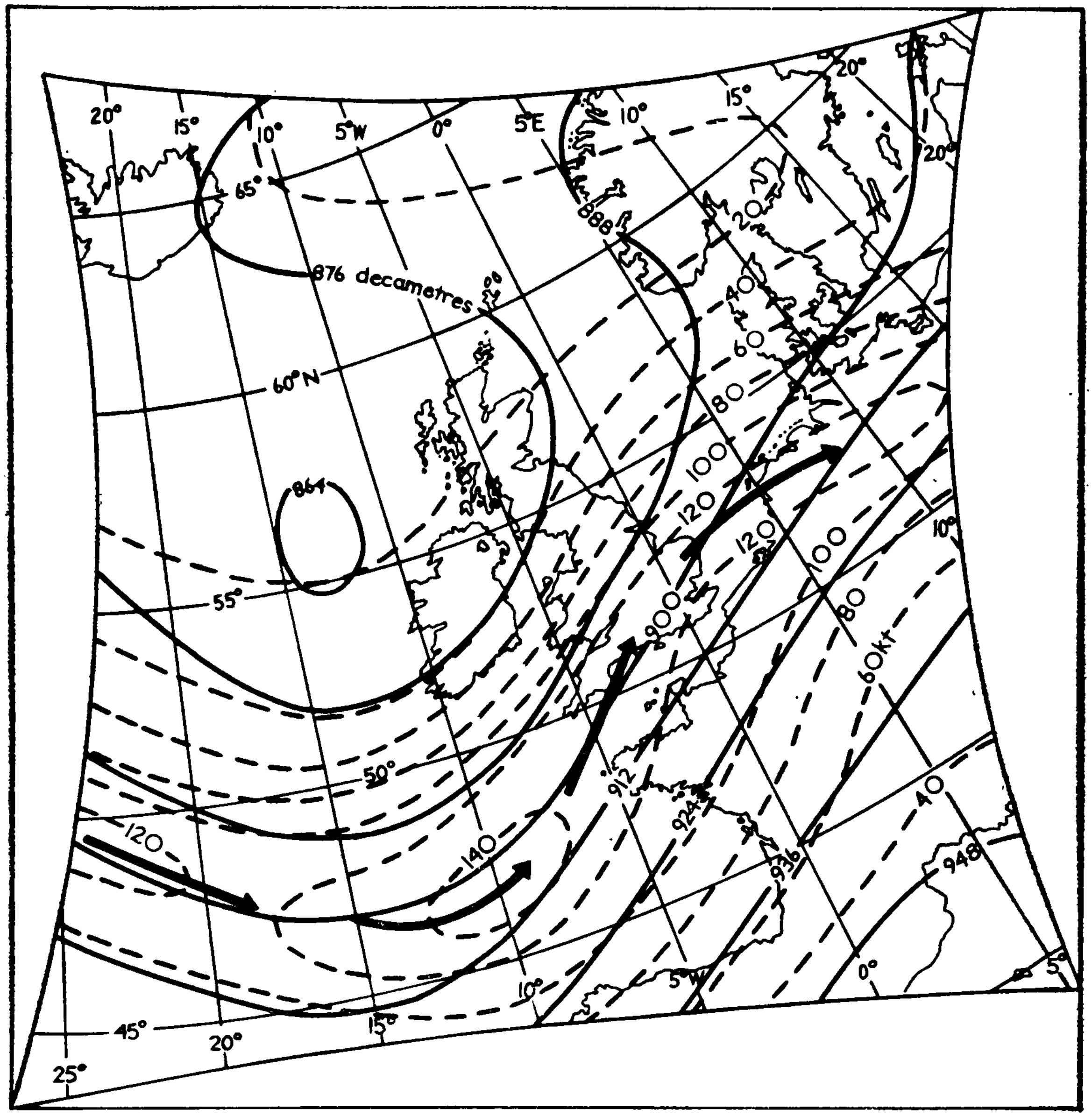


FIG. 126. *Contour chart for 300 millibars, 1200 GMT, 5 May 1966* (after Pothecary and Ratcliffe*)

——— Contours at 12-decametre intervals — — — Isotachs at 20-knot intervals
The jet stream is indicated by bold arrows.

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*POTHECARY, I. J. W. and RATCLIFFE, R. A. S.; Satellite pictures of an old occluded depression and their usefulness in analysis and forecasting. *Met. Mag., London*, 95, 1966 p. 332.

A vertical section for 1200 GMT on 5 May between Liverpool and Paris, Fig. 127, shows the position of the jet-stream core over south-east England at pressure altitudes around 300 millibars. The figure shows also the rapid change horizontally of the wind speed on the cold (north) side of the jet-stream core, contrasted with the gradual change on the warm (south) side, and the sharp decrease of wind speed upwards above the tropopause.

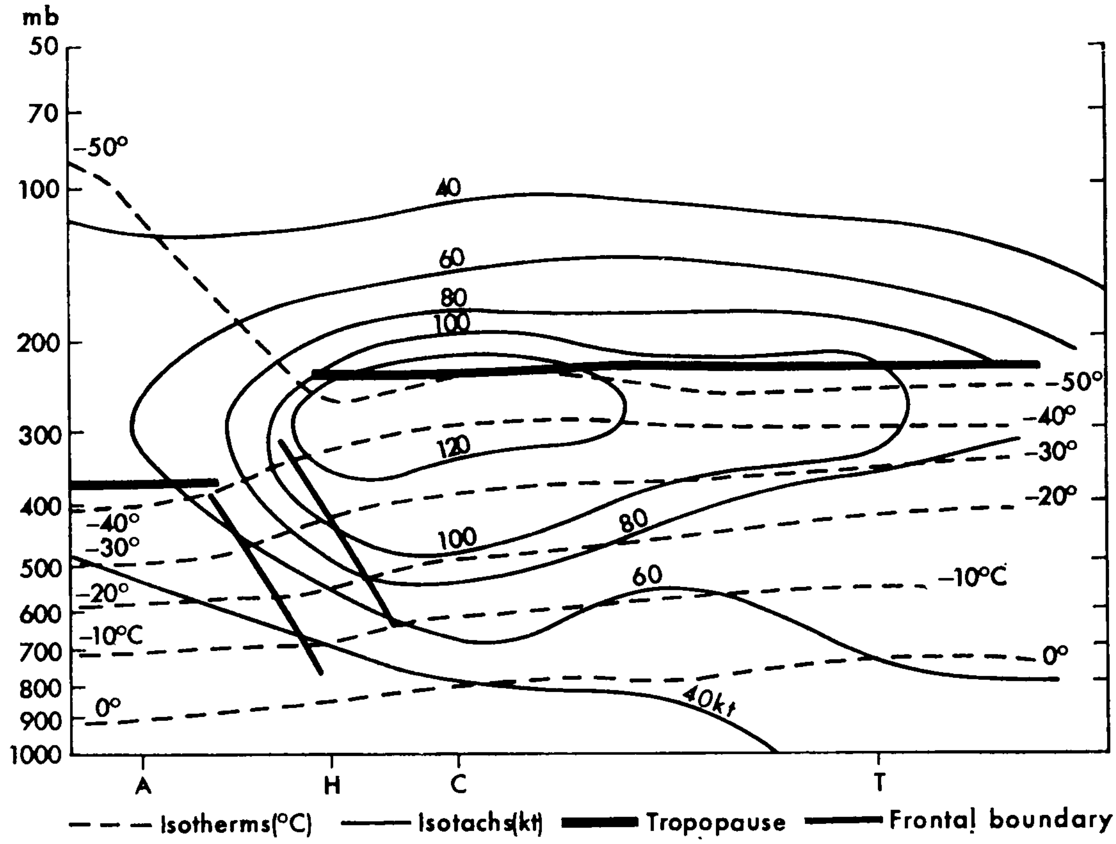


FIG. 127. *Vertical cross-section from Aughton to Trappes, 1200 GMT, 5 May 1966*

A Aughton H Hemsby C Crawley T Trappes

143. THUNDERY SITUATION: 18 JULY 1953

On 18 July 1953 there were widespread outbreaks of thunderstorms in the Midlands and the south and east of England. The thunderstorms were of the 'heat' or 'air-mass' type described in Chapter 7. The general synoptic situation over the British Isles, western Europe and the eastern Atlantic at 0600 on 18 July is shown in Fig. 128. A large depression LH with a weak circulation of winds is centred off the east coast of Scotland and extends over the whole of the British Isles and the sea area between Iceland and Norway. The nearest front lies from southern Scandinavia to eastern France and it is evident that the thunderstorms which developed in England were not of the 'frontal' type. Over England the curvature of the isobars is generally cyclonic though it is not possible to locate a definite trough. On the Atlantic a ridge of high pressure extends from westward of Spain towards Iceland. The weather over the British Isles and the adjacent areas is generally dry with variable cloud while in some places there is an almost clear sky; as far as reports of cloud and precipitation are concerned there is no indication of the impending thunderstorms. This illustrates a fundamental difference in the forecasting of frontal and convective rainfall; frontal rainfall is usually present on the synoptic chart when the forecast is made and the forecasting problem is concerned mostly with its movement though also with its change of intensity; on the other hand convective rainfall often develops *in situ* within an air mass and the forecasting problem is concerned with the temperature and humidity and possible changes of lapse rate within the air mass, for which a representative set of tephigrams is necessary.

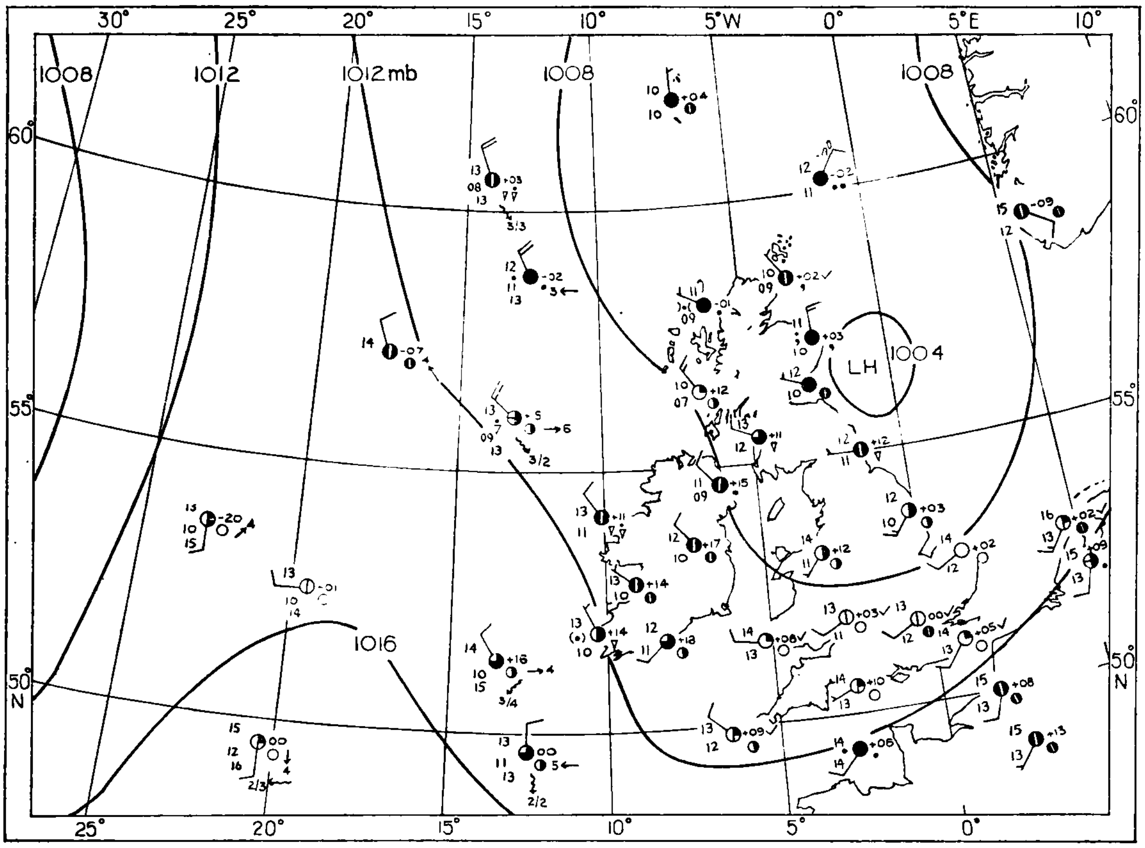
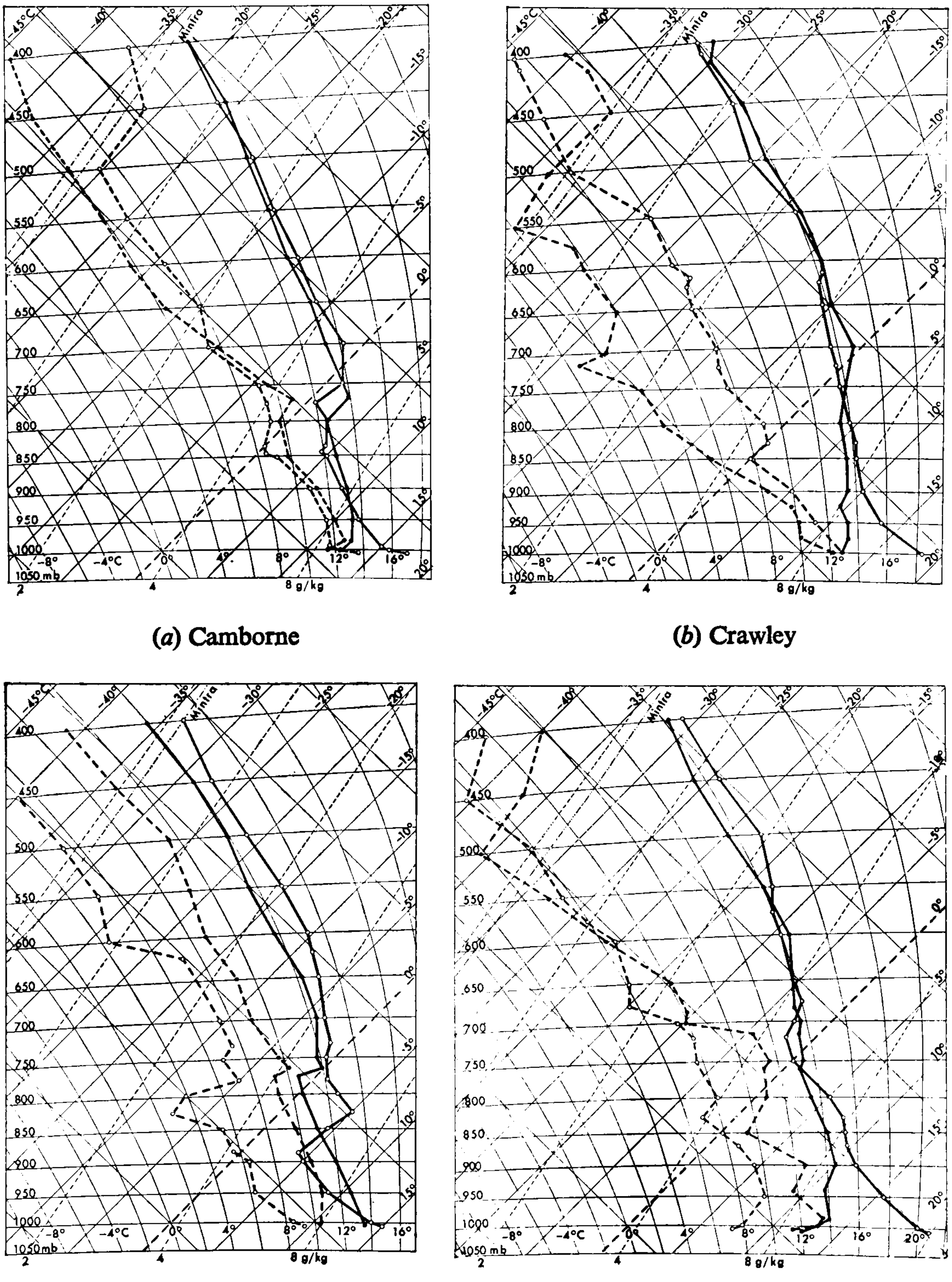


FIG. 128. *Surface chart for 0600 GMT, 18 July 1953*

Tephigrams for the four stations most representative of southern and central England are reproduced in miniature in Fig. 129. By considering first the 0300 ascents (the forecaster's working data in this instance), it is noticed that all four ascents show a lapse rate near to the saturated adiabatic from the surface upwards to at least 400 millibars (about 24 000 feet). The ascent curves for Camborne, Crawley and Hemsby lie roughly along the saturated adiabatic for 13° or 14°C while that for Liverpool (now Aughton) lies nearer the 12°C curve. As insolation is sufficient to produce the normal daily maximum temperature between 19° and 21°C (65° and 70°F), it appears probable that instability with showers or thunderstorms will develop by the afternoon, following the increase of lapse rate caused by heating at the surface. It should also be remembered that the distribution of pressure over the British Isles with its weak cyclonic circulation is a favourable one for thunderstorm development. The humidity in the upper air as shown by Fig. 129 is however not entirely favourable for thunderstorms, for at Hemsby and Camborne there is very dry air in the middle and upper levels and at Crawley there is dry air at all levels except at the surface. The other three stations show fairly



(a) Camborne

(b) Crawley

(c) Liverpool

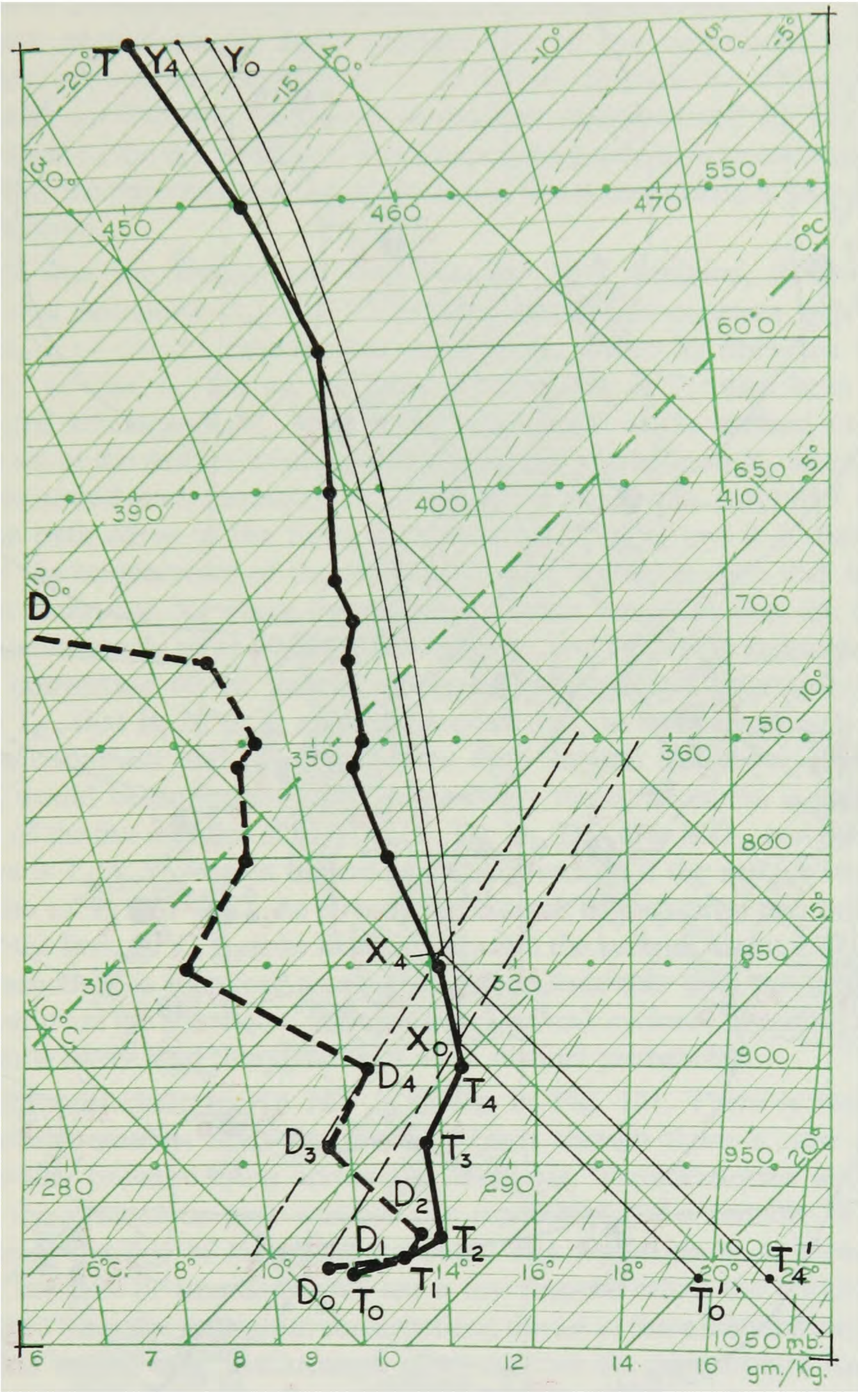
(d) Hemsby

FIG. 129. *Tephigrams for 0300 and 1500 GMT, 18 July 1953*

• ——— • Dry bulb, 0300 GMT   • - - - - • Dew-point, 0300 GMT

o ——— o Dry bulb, 1500 GMT   o - - - o Dew-point, 1500 GMT

moist air from the surface up to about 750 millibars but whether the combination of this moist air with the degree of surface heating to be expected in the afternoon would be suitable for starting strong convection can only be estimated from a closer inspection of the tephigram. For this purpose a larger-scale reproduction of the tephigram for the 0300 Hemsby ascent is shown in Fig. 130. The principles of



Temperature ●——● Dew-point ●– – –●

FIG. 130. Tephigram for Hemsby, 0300 GMT, 18 July 1953

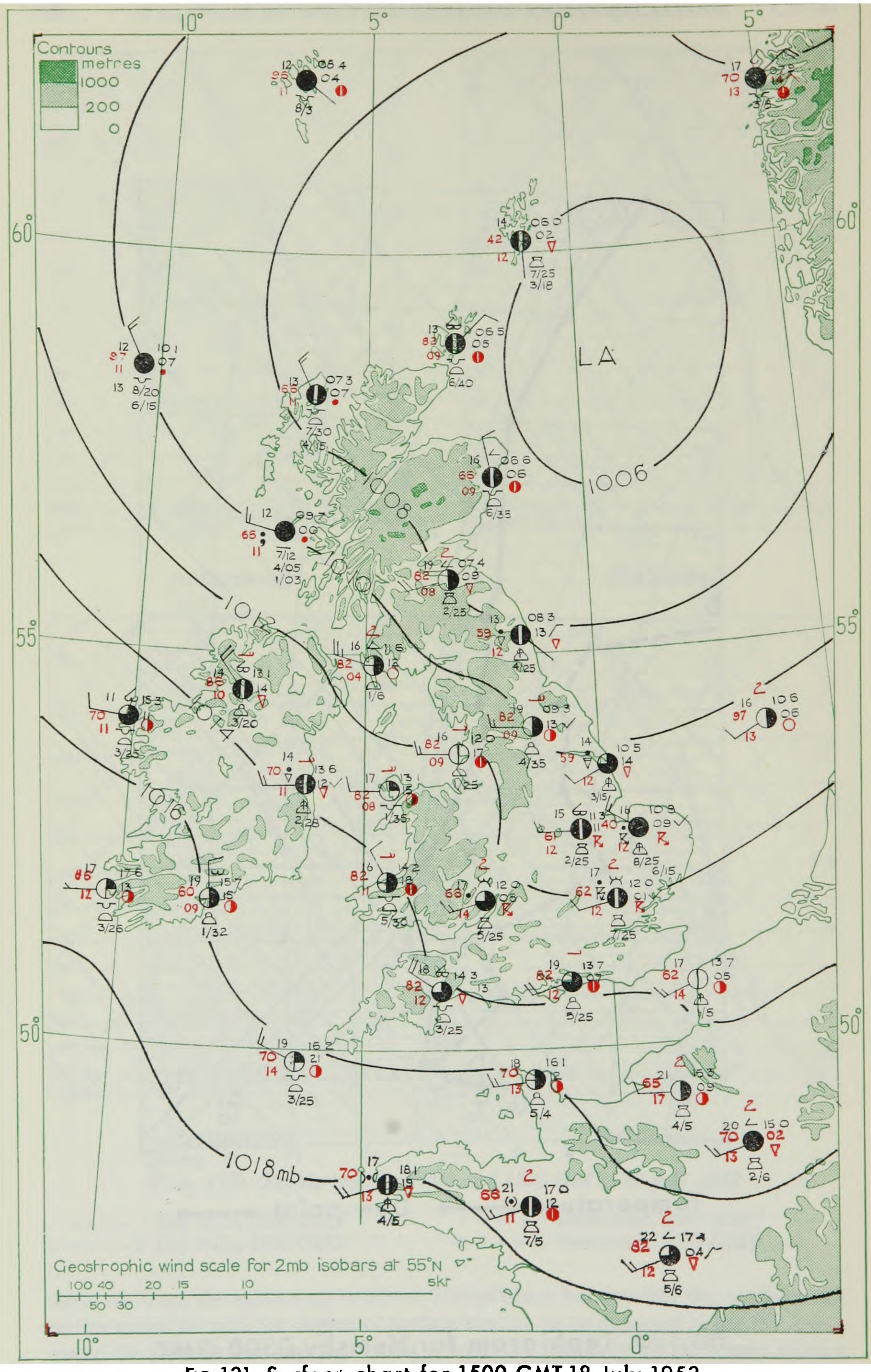


FIG.131. Surface chart for 1500 GMT, 18 July 1953

the method for estimating convection now to be exemplified are explained in Chapters 3 and 6 and it will be assumed that the reader is familiar with them.

In Fig. 130 the dry-bulb temperatures for the Hemsby ascent are marked T_0, T_1, T_2, . . . T from the surface up to the 500-millibar level, being plotted mostly at intervals of 50 millibars, and the points joined by a continuous curve. The dew-points (each plotted on the same pressure line as the corresponding dry-bulb temperature) are marked D_0, D_1, D_2 . . . D. Clearly the existing tephigram represents a stable atmosphere, for the air is nowhere saturated and the lapse rate is nowhere equal to the dry adiabatic. The forecaster's problem is to estimate whether as a result of the day-time heating of the lower layers convection is likely to set in up to high levels.

Consider first the surface air, with pressure 1010 millibars, temperature T_0 (= 12°C) and dew-point D_0(= 11·5°C). Draw the line of constant water vapour content D_0X_0 to cut the environment curve in X_0. From X_0 draw the dry adiabatic X_0T_0' so that T_0' lies on the same pressure line as T_0; also from X_0 draw the saturated adiabatic X_0Y_0. Then if the temperature at the surface were to rise to T_0' the air at the surface would rise along the dry adiabatic to X_0 where it would become saturated and thereafter continue rising along the saturated adiabatic X_0Y_0. The path curve of the rising air would be $T_0'X_0Y_0$ and it is clear that at each level the temperature of the rising air would be higher than that of the environment (except at X_0 where the temperatures are equal) so that the air would ascend spontaneously to at least the 500-millibar level (Y_0). It is deduced therefore that free convection of the surface air would take place if its temperature were to rise to T_0', that is, 20°C. It might be argued that the air thus rising would mix with its environment and thereby become drier so that the above construction, using D_0 would be no longer valid. Accordingly the construction is repeated for the driest of the five samples of air (D_0, D_1, D_2, D_3, D_4); this is D_4, and the resulting path curve is $T_4'X_4Y_4$ – again indicating free convection if the surface temperature were to rise to T_4' (21·6°C). As already mentioned, of the above two critical temperatures the first, 20°C, is quite likely in July and the second, about 22°C, is by no means unusual. It is therefore reasonable to conclude that with the probable rise of temperature in the afternoon, together with the weak cyclonic situation, afternoon convection is likely, and thunderstorms are a distinct probability.

The afternoon chart for 18 July is shown in Fig. 131. There is little change in the isobaric pattern from the morning chart though there is now a distinct trough of low pressure lying across the Midlands. Thunderstorms and showers either in the present or past weather are reported from a number of stations, mostly along the line of the shallow trough. The temperatures at these stations reporting showers or thunderstorms are a few degrees below the critical temperature derived from the tephigram construction; the air at these stations has probably been cooled by the rain, but at some other stations the temperature has reached 19°C. Reports from other stations not plotted on these charts show that the afternoon temperature rose to between 20° and 22°C at a number of places in east and south-east England so that the critical temperatures were in fact reached or exceeded. The afternoon tephigrams (Fig. 129) show on the whole little change from the morning ascents except for the usual rise of temperature and increase of lapse rate to near dry adiabatic in the lowest layers.

That thunderstorms were widespread is evident from Fig. 132 on which a thunderstorm symbol has been marked for each station reporting a thunderstorm between 0900 and 2100 GMT on 18 July.

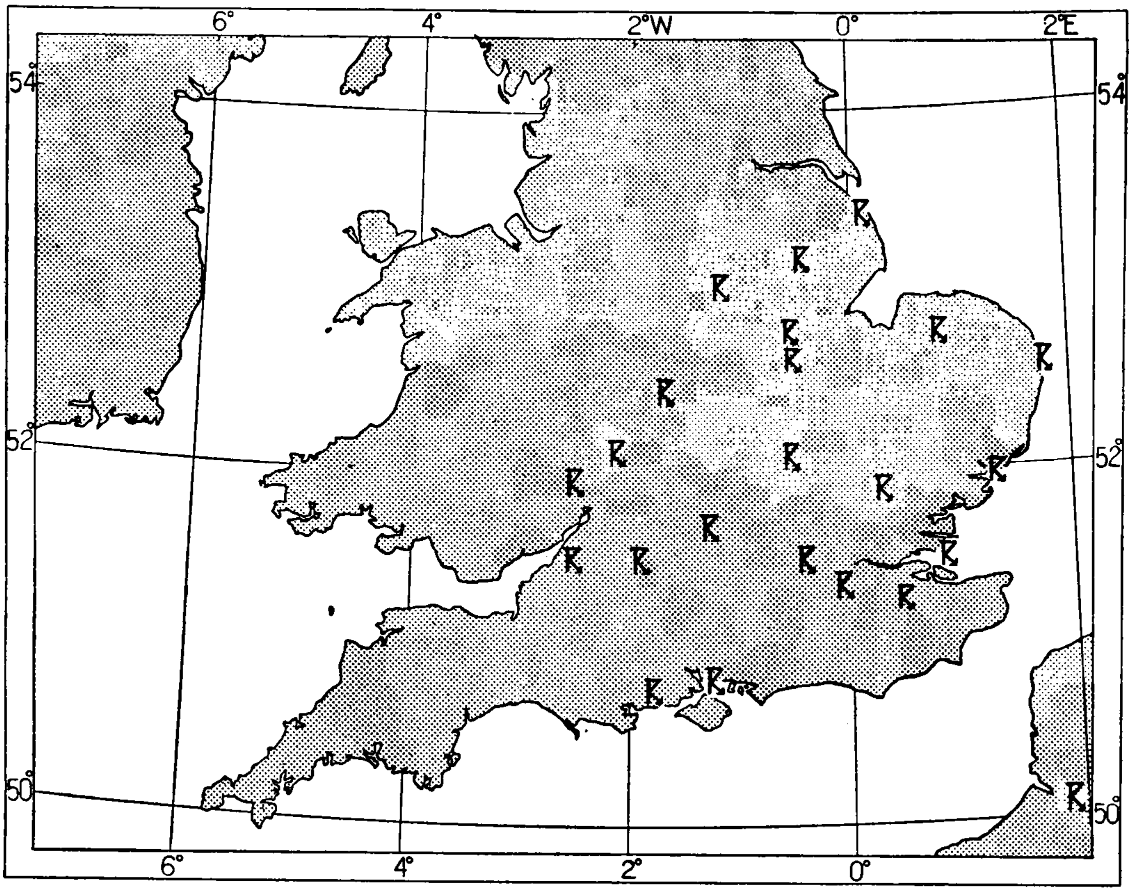


FIG. 132. *Distribution of thunderstorms 0900–2100 GMT, 18 July 1953*

144. PERIOD: 17–18 OCTOBER 1965

This period illustrates an anticyclonic foggy situation over the British Isles.

The general situation at 1200 GMT on 17 October 1965 is shown in Fig. 133. A large anticyclone extends from the Atlantic across the British Isles to north-west Germany with one centre HB south-west of Ireland and another HA over the southern North Sea. A frontal system extends from a depression north of Iceland down the west coast of Norway and across the British Isles between the anticyclonic cells. The cold front is very weak where it passes through the anticyclone and gives no more than layers of cloud, mainly broken, and very occasional showers. Near the centre of the anticyclone the winds are light and variable. There are only small amounts of cloud over most of England and smoke and haze have reduced visibility to poor or moderate.

By 0000 GMT on 18 October (Fig. 134) the anticyclone HB has moved north-eastwards and is centred over north-east Ireland. There is a fairly definite clockwise circulation of winds round the high pressure centre but most stations over England have either calm conditions or very light winds. Skies are clear at most places and fog is already being reported from several inland stations and is dense at Birmingham and Cardington. The weak cold front has drifted south and is giving cloudy conditions with occasional drizzle at Valley and Dublin and broken stratocumulus

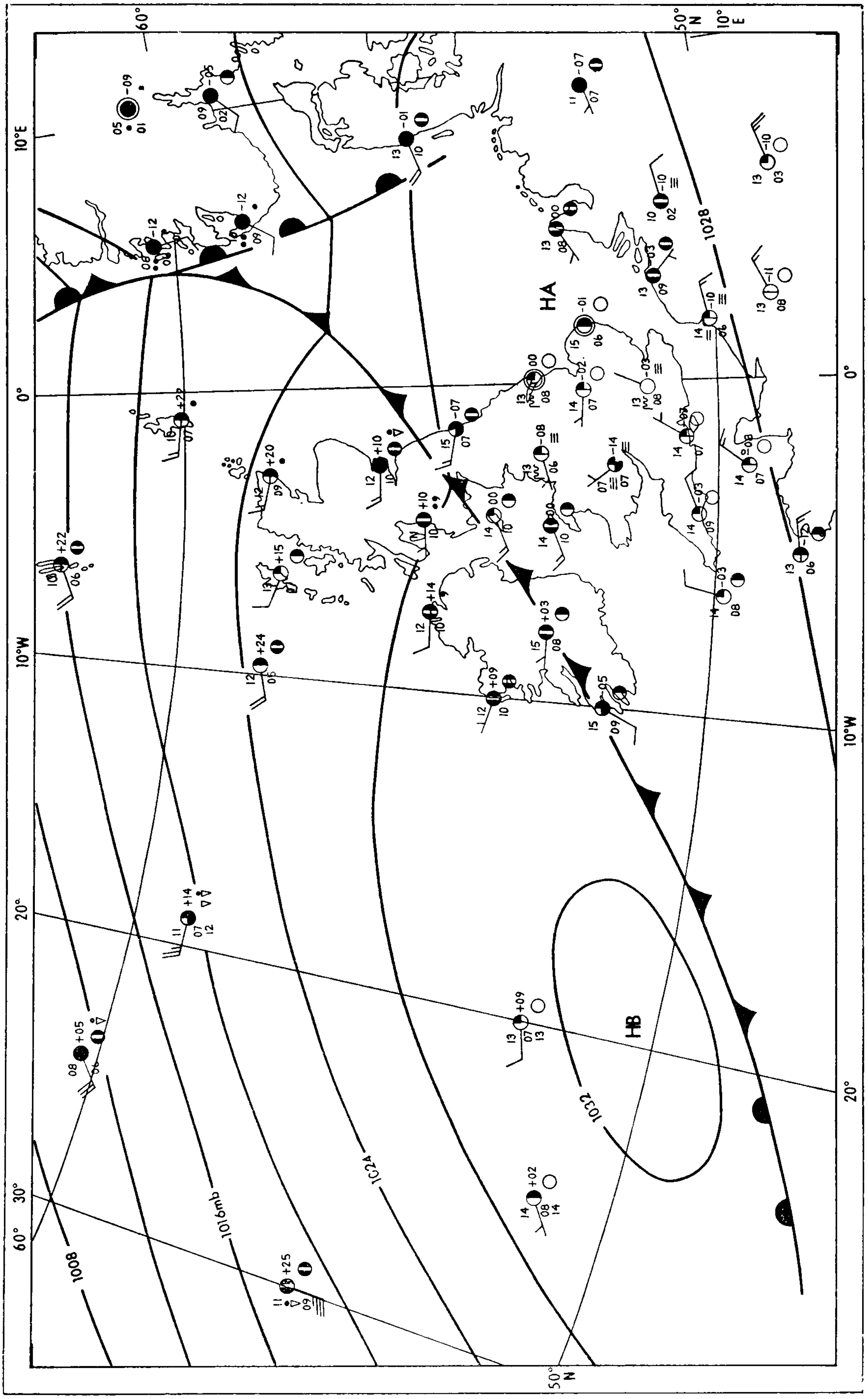


FIG. 133. *Surface chart for 1200 GMT, 17 October 1965*

at Spurn Head and Finningley. The upper air situation for this period is represented by the 500-millibar chart for 0000 GMT on 18 October (Fig. 135). It can be seen that the gradient over the British Isles is very slack with little flow to be expected.

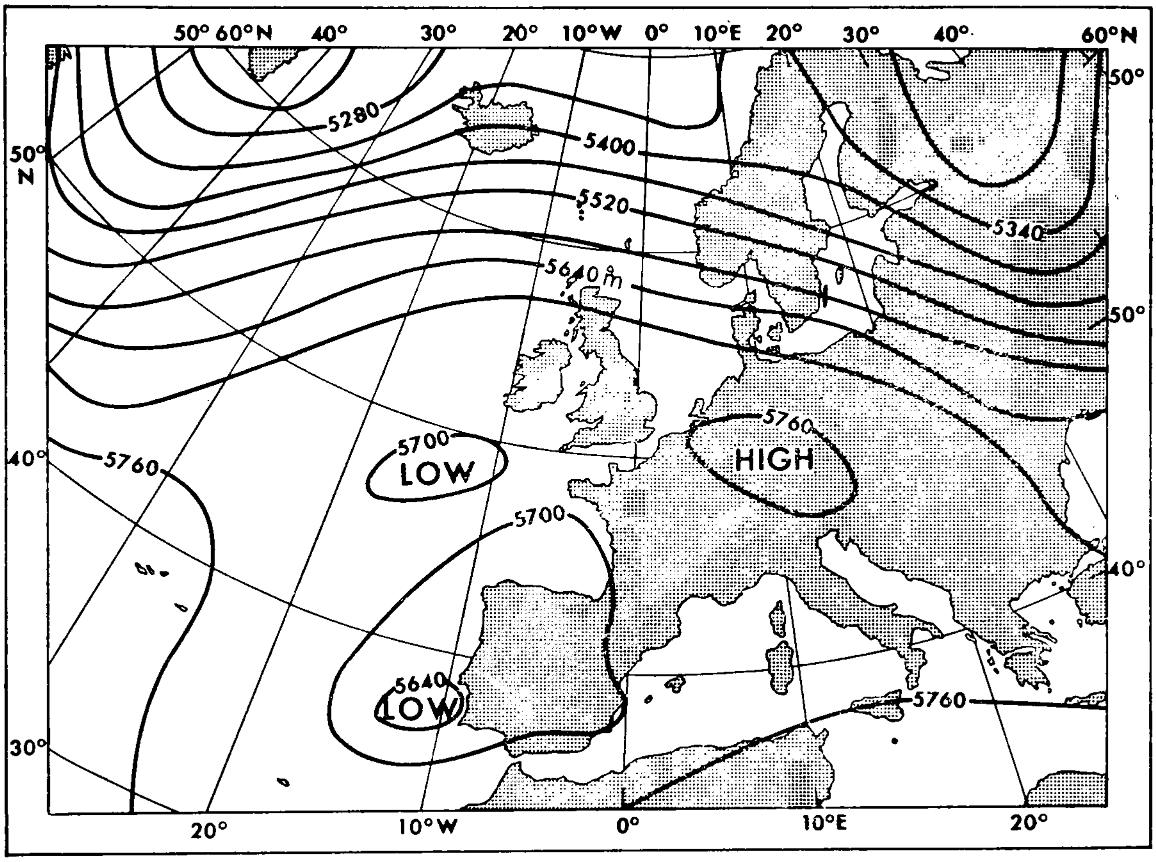


FIG. 135. *Contours for 500 millibars, 0000 GMT, 18 October 1965*

By 0600 GMT on the 18th (Fig. 136) fog, some of it dense, has become widespread over inland areas of south Scotland and much of England. In the foggy areas temperatures have fallen considerably from the midday values of the previous day. Patchy stratocumulus has reduced the fall of temperature at places in eastern England, such as Heathrow, and a light drift from the North Sea has had a similar effect on parts of the east coast. In the extreme south light winds have been maintained by the pressure gradient. This situation exemplifies the difficulty of forecasting for a particular place. A forecast for 0600 GMT on the 18th of fog over much of England, particularly dense in industrial and built-up areas, would have been generally correct, but in the event many places in the south and east retained visibilities of more than 2000 metres because of the complexities mentioned above.

So long as light winds persist, foggy weather may be expected to continue in places where the skies remain clear. The problem of forecasting for areas where there is patchy stratocumulus is a difficult one. On this occasion, despite little change in the upper air pattern, the surface anticyclone HB was centred over the North Sea by 0000 GMT on the 19th. The easterly drift over south-east England became a little more definite and the 0000 observation on the 19th at Heathrow was 772 00408 16100 27509 00900 08605, that is a clear sky, wind 8 knots and visibility 1600 metres. On the other hand, Gatwick for the same time had a clear sky, wind 2 knots and visibility 100 metres.

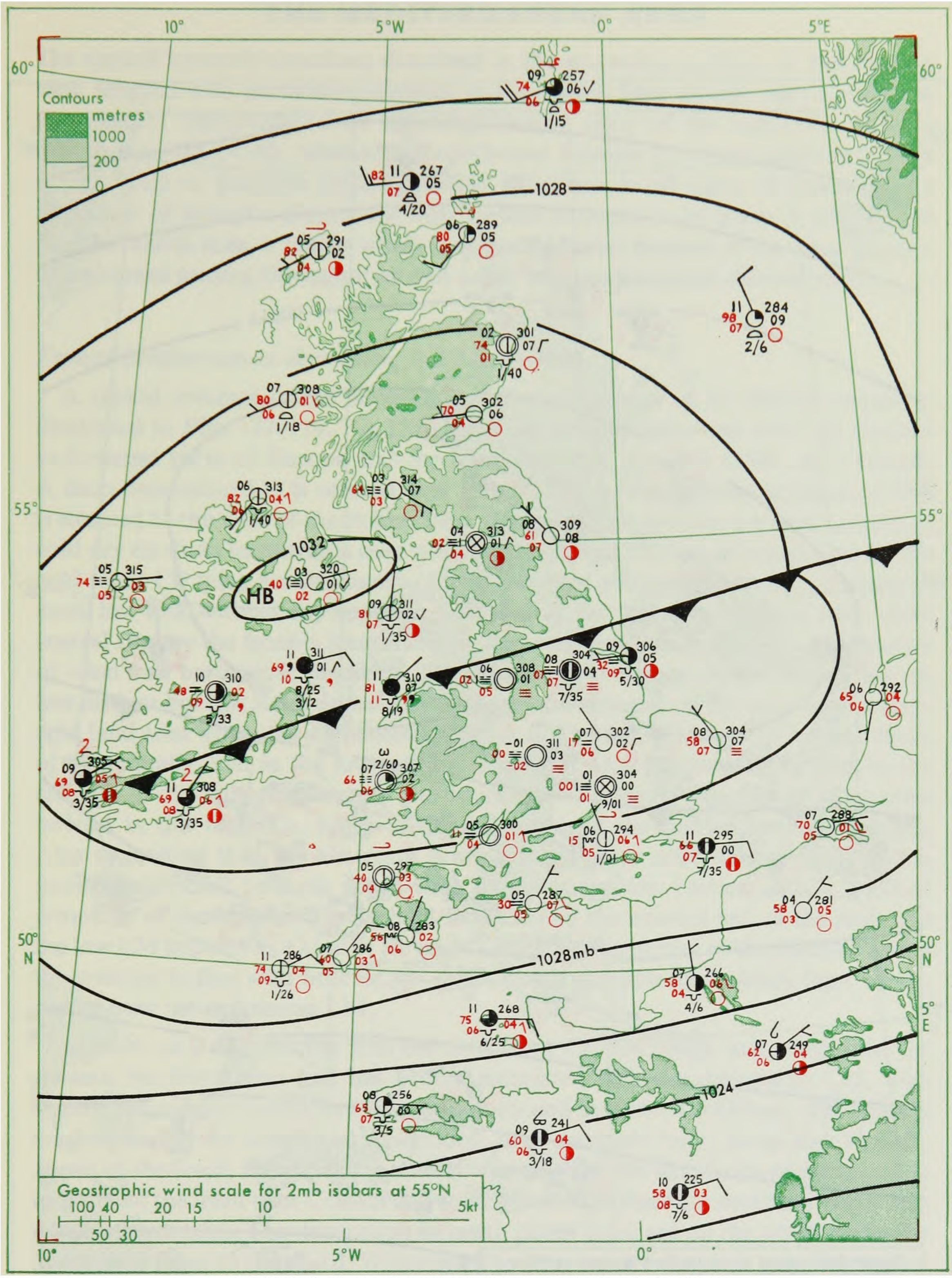


FIG.134. Surface chart for 0000 GMT, 18 October 1965

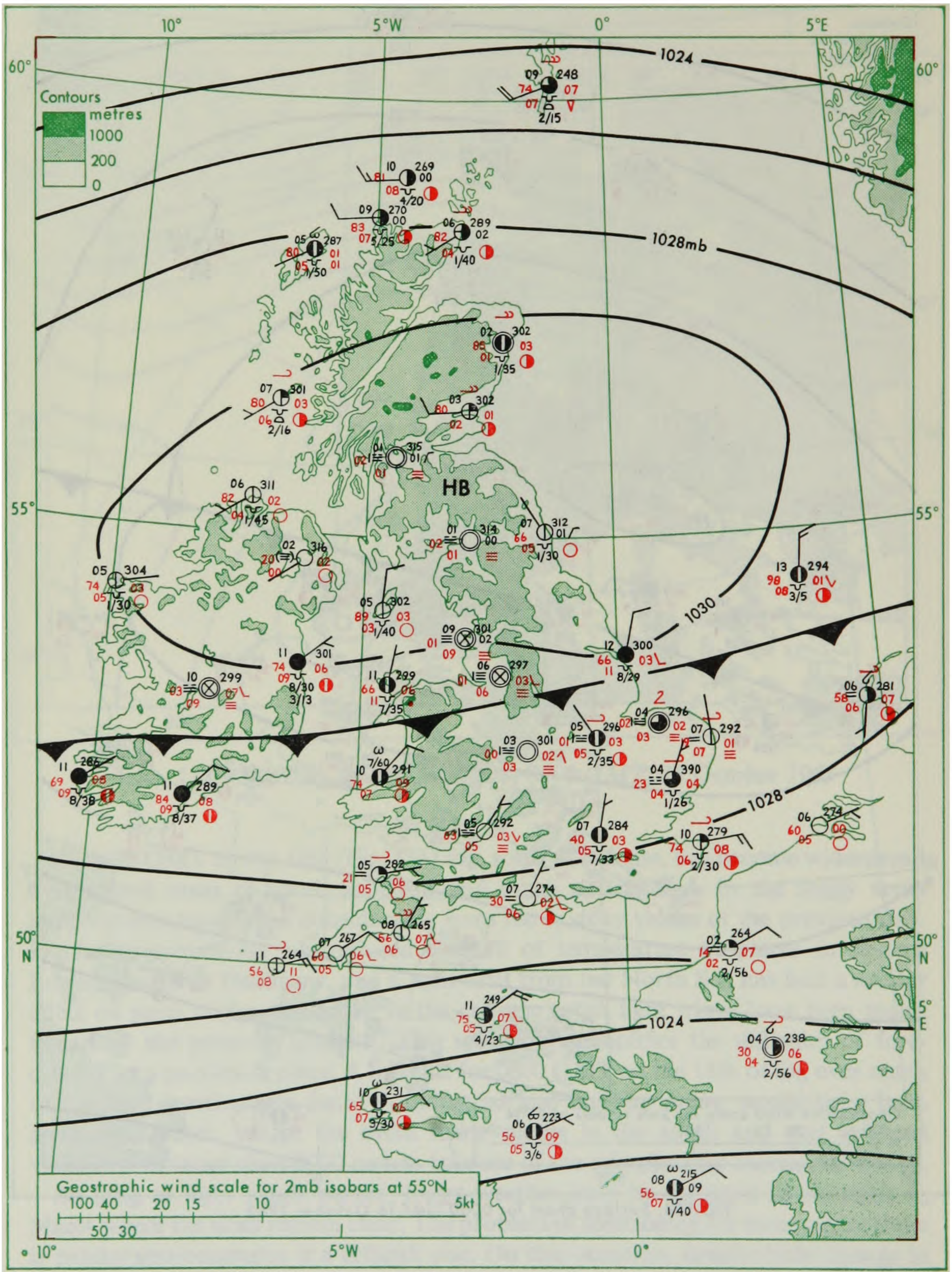


FIG.136. Surface chart for 0600 GMT, 18 October 1965

145. AN EXAMPLE OF SYNOPTIC ANALYSIS FOR THE MEDITERRANEAN AREA

The typical synoptic situations described in the preceding sections of this chapter were selected with particular reference to the British Isles though they are also to some extent representative of the neighbouring parts of the temperate latitude belt, that is, the North Atlantic and north-west Europe. It is impossible in a book of this kind to illustrate typical weather situations in all parts of the world by sequences of synoptic charts; the only further example to be given is one for the Mediterranean area, a region which is of great interest because of the large number of air routes passing through it as well as for its meteorological characteristics.

Typical Mediterranean depression, 1–2 March 1949

A typical instance of the development of a depression in the Mediterranean is illustrated in Figs 137–139. Fig 137 shows the synoptic situation over the western and central parts of Europe and the Mediterranean for 1800 GMT on 1 March. A deep depression LD is centred over Poland and a fairly intense anticyclone HB is situated to the west of the British Isles. Between these systems a strong current of cold dry air flows southwards over western and central Europe. It is preceded by the cold front CE lying across Austria, along the Alps and across southern France. A small low LM, without any apparent frontal structure, lies over northern Italy. Over central Europe the front is identifiable by the usual features, principally the changes of wind and barometric tendency. The changes of temperature and dew-point are less distinct but this is not unusual in such a winter situation in the interior of a large cold land area where the air masses on either side of the front tend to become more or less equally cold in the lower levels. Near the Alps the front is located by the changes in dry-bulb temperature and barometric tendency. The feature of greatest interest in this region is, however, the distortion of the isobars as they cross the Alps suggesting that the airstream flowing southwards over western Germany is strongly deflected towards south-west by the mountain barrier, with a marked crowding of the isobars. The isobars swing round the western end of the Alps and the low LM appears as a kind of lee eddy. Over the greater part of the Mediterranean the weather is fine with little or no cloud, good visibility and mainly light winds, except near the depression LM.

At 0600 on 2 March (Fig 138) the cold front CE has swept well south into the western Mediterranean and has beome associated with the depression LM. This system has moved to southern Italy and deepened by some 4 millibars. The front is marked mainly by winds and tendencies; over southern Italy, Sicily and Tunisia, ahead of the front, the winds are mainly westerly and the tendencies negative, while to north of the front over Corsica and southern France there are northerly winds and positive tendencies. The dew-points to north of the front are on the whole distinctly lower than those to the south. With regard to temperature the main point of interest is the progressive rise of temperature in the cold air mass as this gets farther south over the warm sea: from 3°C in south France to 8°C in Corsica and 10°C in Sardinia. It is evident that the warm sea surface tends to obliterate the frontal contrast of temperature. Precipitation is very scanty near both the depression and the front and is reported mainly as showers; cloud is mostly cumuliform and well broken at a safe height, while visibility is mostly good. (Lack of observations makes it impossible to mark the front with any certainty over the Adriatic and the Balkans.) It is instructive to contrast the weather in this depression with that illustrated at Fig. 119 where

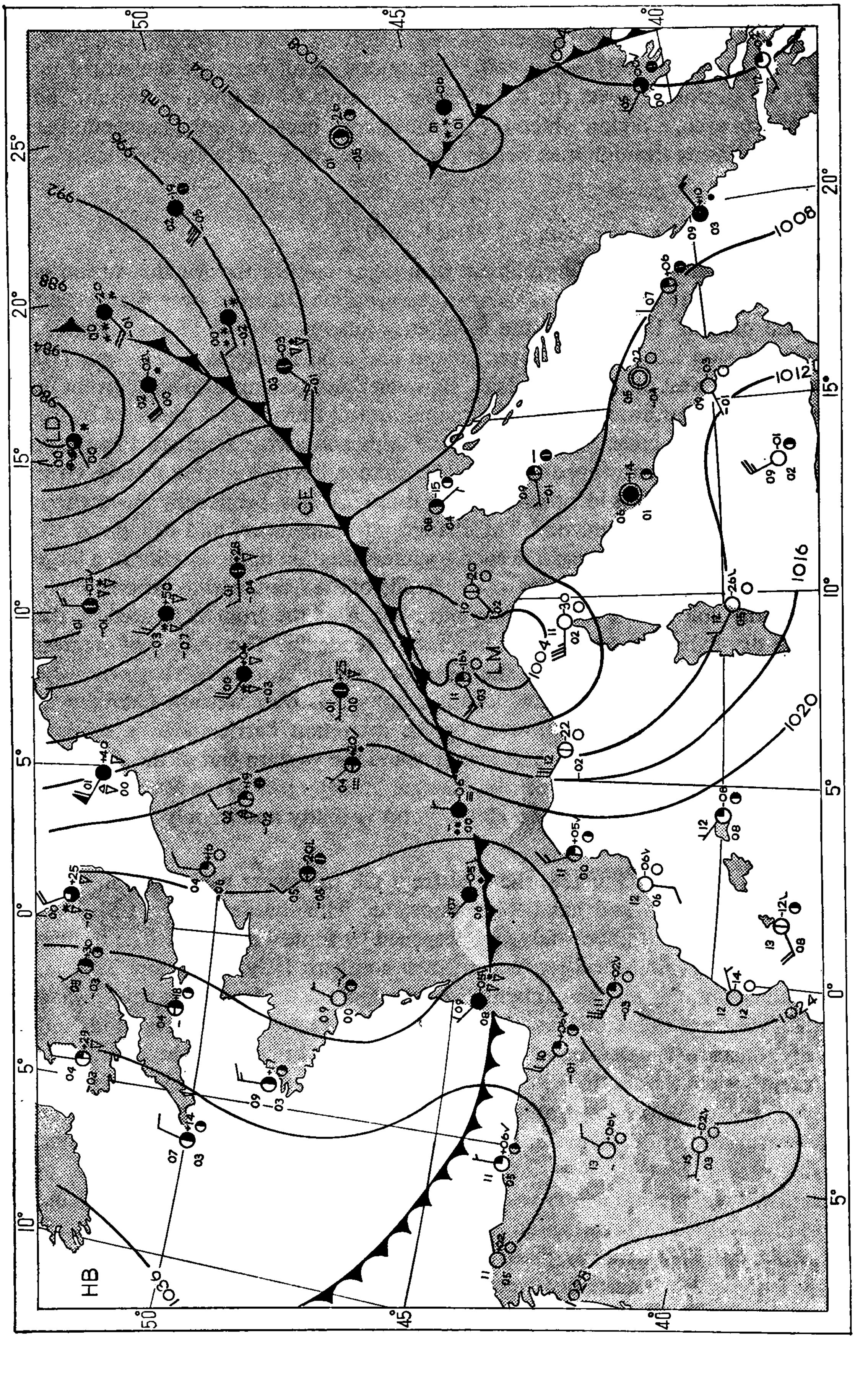


FIG. 137. *Surface chart for 1800 GMT, 1 March 1949*

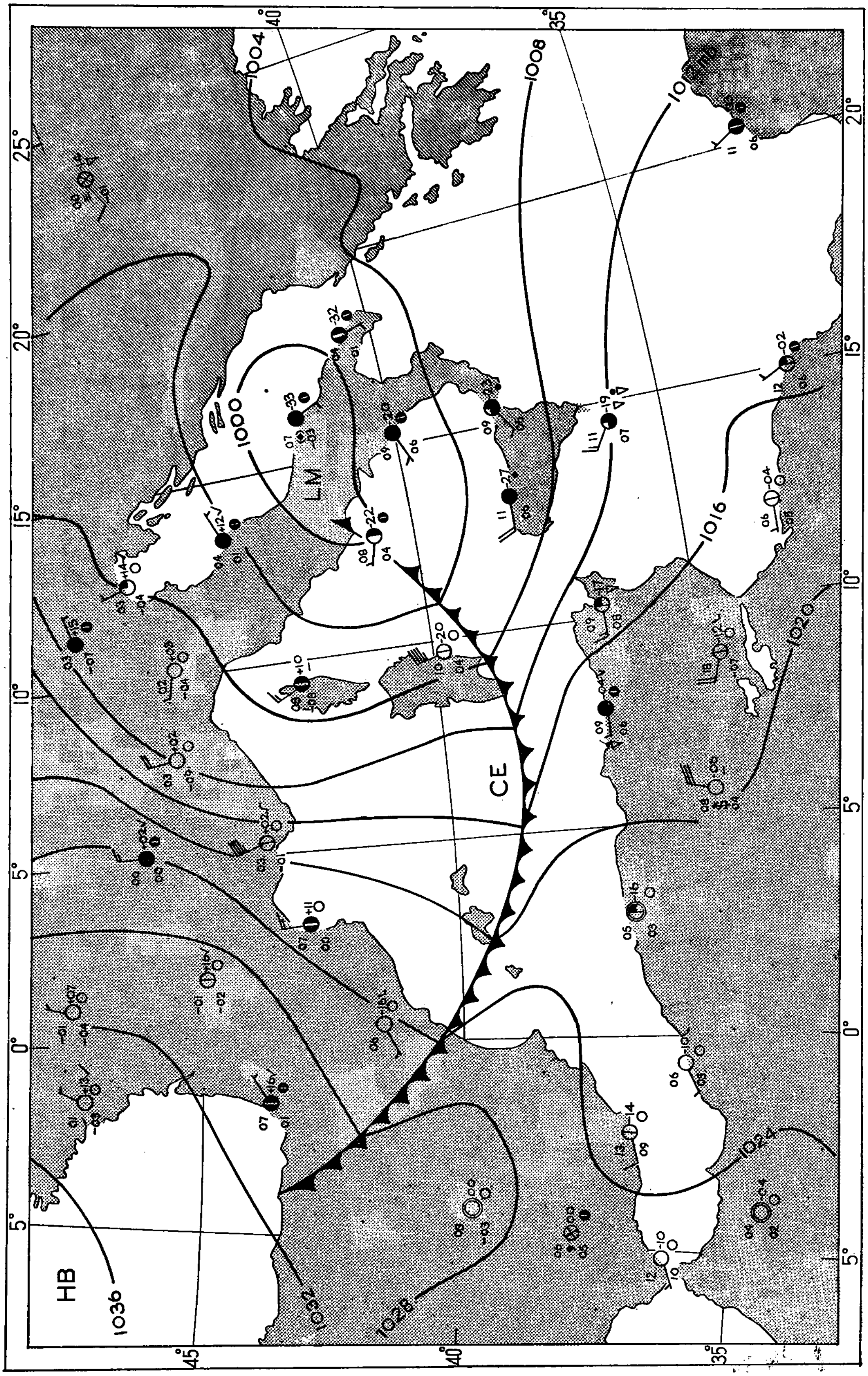


FIG. 138. *Surface chart for 0600 GMT, 2 March 1949*

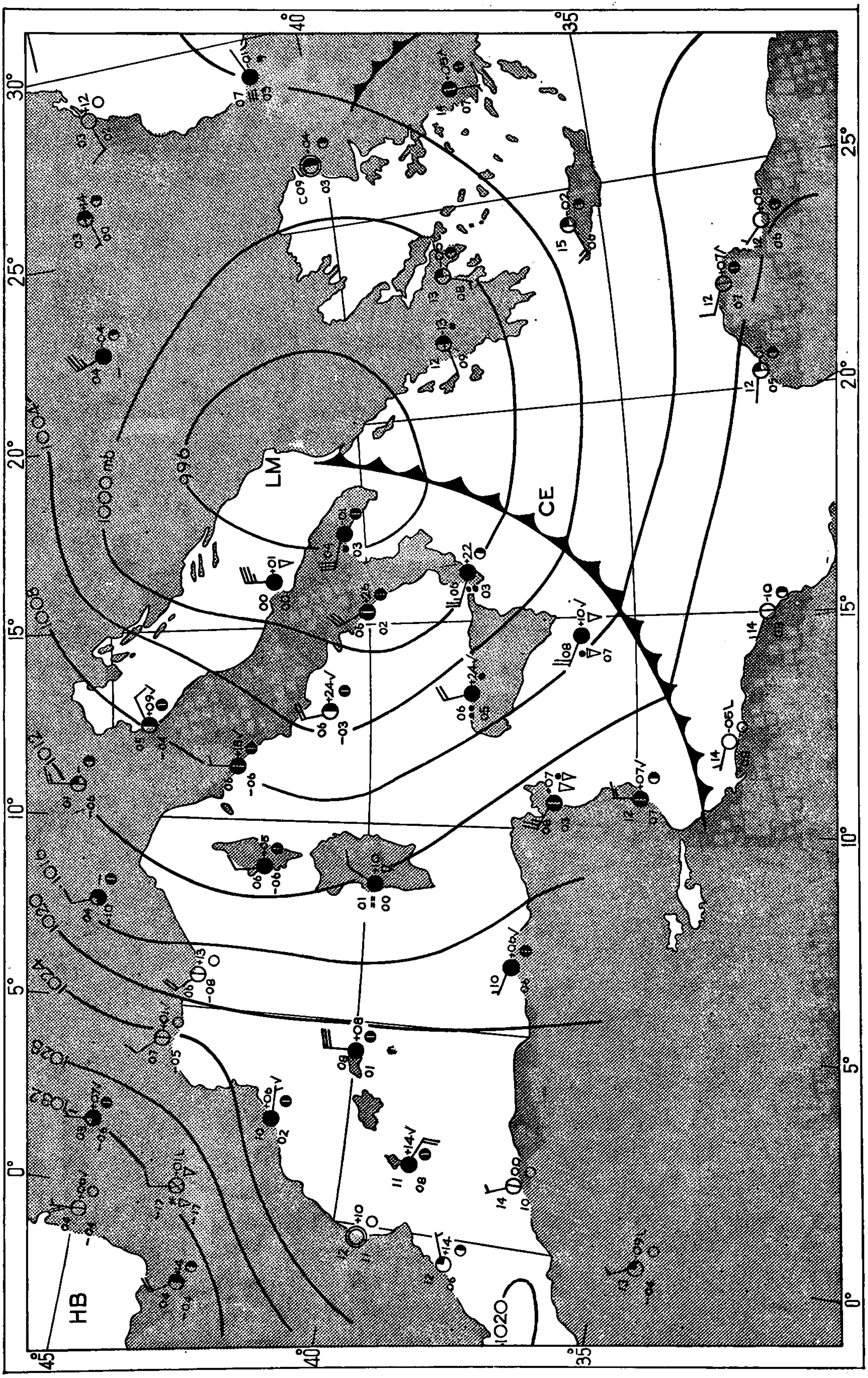


FIG. 139. *Surface chart for 1800 GMT, 2 March 1949*

precipitation, overcast skies and reduced visibility are reported over a considerable area.

At 1800 on 2 March (Fig 139) the depression LM is centred over the southern Adriatic having deepened by over 4 millibars. The front CE has swept round south of the centre and is picked out on the basis of winds and tendencies. Its precise location between Greece and southern Italy is difficult to fix because of the absence of observations. Clearly it has passed Sicily, Malta and the two stations on the east coast of Tunisia but has not reached the stations in Tripolitania or Greece. This chart shows a fairly widespread outbreak of showers or even continuous rain behind the cold front, chiefly over southern Italy and Sicily, with a considerable increase of cloud by comparison with 0600 but the cloud is still mostly cumuliform and its base is not especially low, while visibility continues mainly good.

PART IV

METEOROLOGICAL ORGANIZATION

CHAPTER 21

METEOROLOGICAL SERVICES FOR AVIATION

146. METEOROLOGICAL OFFICES FOR AVIATION

Meteorological offices are maintained at many Royal Air Force formations and aerodromes and at civil aerodromes to provide the meteorological information required by the various consumer interests concerned.

Principal Forecasting Offices, such as at London/Heathrow Airport and Headquarters Strike Command, prepare routine route and area forecasts and charts and analyses for various levels.

Main Meteorological Offices are maintained at Air Traffic Control Centres, at RAF Group Headquarters and at major RAF and civil aerodromes. They normally control a number of subsidiary and observing offices, to whom they give advice based largely on charts and forecasts supplied by their Principal Meteorological Office.

Subsidiary Meteorological Offices are maintained at RAF and civil aerodromes of intermediate importance. The service provided by these offices is not normally on a 24-hour basis but covers various periods during the day and night to meet local requirements. Such offices obtain guidance and advice as necessary from their parent main meteorological office.

Observing Offices are maintained at certain RAF and small civil aerodromes. The service immediately available at these offices is limited to current aerodrome weather reports. The hours of watch may be restricted but are on a 24-hour basis at Master Diversion Aerodromes or if the station forms part of the synoptic network. Forecasts and similar information which may be requested are obtained from a main or subsidiary meteorological office.

147. SUPPLY OF INFORMATION TO METEOROLOGICAL OFFICES

An outline of the general arrangements for broadcasting meteorological information has been given in Section 99. From the Meteorological Communications Centre at Bracknell teleprinter channels radiate to each main office and thence to their subsidiaries. In addition to the teleprinter network of broadcast messages of weather reports and forecasts, analysed charts for surface and upper levels are received on a nation-wide facsimile broadcast from the Central Forecasting Office at Bracknell.

Meteorological offices supplying forecasts and information at major civil airports also receive a broadcast of forecast charts from the Principal Forecasting Office at Heathrow. These include significant weather charts, surface forecast charts, contour forecast charts for upper levels, spot wind charts, and tropopause height and maximum wind charts, all received by a facsimile broadcast at regular intervals on a routine basis. The Principal Forecasting Office at Headquarters Strike Command and at the Main Meteorological Office at Headquarters Air Support Command provide a similar service to RAF stations within their respective commands.

These services ensure that all forecasters at subsidiary offices can be supplied with up-to-date charts, analyses and forecasts from the facsimile network, and the latest observations and landing forecasts from the teleprinter network. This provides an overall picture covering a wide area, and the details of local weather are filled in from the regional network of observations and from the forecaster's experience of local conditions.

Exchanges of forecasts take place between countries and between different regions of the world on a routine basis and the forecasts can be made available to a station requiring them through the general broadcast system.

Broadcast schedules and the information available change frequently and the details of current arrangements can be found in the *U.K. Air Pilot.*

148. INFORMATION AVAILABLE AT METEOROLOGICAL OFFICES

The following charts and data are available for inspection at most meteorological offices serving aviation:

Weather reports and trend forecasts
Synoptic charts and analyses for standard hours showing surface weather over north-west Europe and the eastern Atlantic
Hourly charts of British Isles weather
Surface forecast charts
Forecast upper air charts and plotted tephigrams from the facsimile broadcast from Central Forecasting Office
Upper wind forecast charts for certain areas
Significant weather charts and tropopause height/maximum wind charts from the facsimile broadcast from the Principal Forecasting Office
Forecast QNH
Warnings of gales, squalls, thunderstorms, frost, snow, thaw, fog, and airframe icing.

Special forecasts for flights not on air routes are prepared for issue as required.

149. SUPPLY OF METEOROLOGICAL INFORMATION TO AIRCREW

Before take-off

The supply of information to service and civil aircrews on the ground is arranged according to their particular needs. There is first of all the briefing on the weather situation which is given by the meteorological officer either on an individual basis or on a mass basis when a number of aircraft are taking part in the same operation. At briefing, the features of the weather situation and its expected development are explained to enable those responsible to judge whether the operation can be carried out, how it is to be done, that is, over what route and at what height, which aerodromes will be open in the event of a diversion being necessary, and so on.

Briefing is normally followed by the issue of one or more written documents. For short flights it is sufficient to supply on the appropriate form a forecast for the route and the destination aerodrome and sometimes for alternate aerodromes. For long overseas flights the documentation is rather more elaborate and usually includes charts for the surface and for the standard pressure levels straddling the proposed flight level, significant-weather charts, tropopause height and maximum wind chart, and forecasts for a selection of aerodromes along the route.

If, for certain military operations, it is desirable to supplement the information available to the meteorological officer, the operational authorities may decide that a special weather test flight should be made beforehand. The meteorological officer is consulted on what observations should be obtained on this flight and the results are referred to him for interpretation before any operational decision is taken.

In flight

There are various means by which an aircraft commander can obtain meteorological information while in flight.

(i) *By the interception of* VOLMET *broadcasts.* These consist, in the United Kingdom, of transmissions by radio-telephony from the Air Traffic Control Centres at Prestwick, Preston and West Drayton, and contain weather reports for selected aerodromes. Trend forecasts and runway visual range values may be added. A trend forecast has a validity of two hours following the time of the routine weather report to which it is added.

Some stations overseas make VOLMET broadcasts by radio-telegraphy.

(ii) *By the interception of broadcasts of weather warnings from Air Traffic Control Centres.* These warnings relate to the occurrence or expected occurrence of conditions hazardous to aircraft and are supplied to the Control Centre by the appropriate meteorological officer.

(iii) *By the receipt of a message specially directed to the aircraft by the appropriate control authority.* This usually occurs only for long-distance flights on which 'flight meteorological watch' is kept. This is an arrangement by which the responsible meteorological office maintains a watch on the meteorological conditions for a designated portion of the flight path; any necessary amendments to flight and aerodrome forecasts are transmitted direct to the commander of the aircraft.

(iv) By requesting information from the centre or aerodrome with which the aircraft is in radio contact.

Aircraft reports. The successful operation of any scheme to supply aircrew with meteorological information depends to some extent on their co-operation. Some aircraft make regular or hourly reports, others make reports on request. All aircraft should transmit a report to the appropriate Air Traffic Control Centre when any phenomenon is encountered likely to affect the safety of other aircraft. The Control Centre is then able to notify other aircraft operating in the area, and the meteorological office at the Control Centre is able to arrange distribution of the message to other interested meteorological offices.

After flight

On landing, the flight forecast obtained at departure should be handed in with any comments thereon at the meteorological office, together with a copy of any observations made in flight either in writing or as a pictorial cross-section, and any supplementary information about conditions in flight. Such information can then, if desirable, be disseminated to other meteorological offices and to aircraft in flight.

150. SUPPLY OF METEOROLOGICAL INFORMATION TO AVIATION AUTHORITIES OTHER THAN AIRCREW

The meteorological officer on an aerodrome keeps in close touch with the authorities responsible for the flying programme and for the control of aircraft in the air.

This requires direct consultation with the officer in charge of flying or, in civil operations, with the company's operations officer.

At many aerodromes and establishments it is essential that Air Traffic Control should be kept fully informed on the relevant aspects of weather. Thus at an Air Traffic Control Centre the following information is supplied on a routine basis:

(i) Weather reports for specified aerodromes within and, where necessary, adjacent to the Flight Information Region served by the Control Centre.
(ii) Reports of improvement and deterioration of the weather at certain aerodromes.
(iii) Forecasts of meteorological conditions for the Flight Information Region and for certain aerodromes within it, with amendments as necessary.
(iv) Warnings of reported or expected weather phenomena considered dangerous to aircraft within the Flight Information Region; also any such warnings received from an adjacent Flight Information Region. These messages are prefixed by the word SIGMET.
(v) Forecasts of lowest QNH for altimeter setting regions (Forecast QNH).
(vi) Advice regarding diversions.

At an aerodrome the meteorological officer keeps the air traffic control officer supplied with the same kind of information as that listed above except that it is restricted to the requirements of aerodrome approach and control. Weather reports for the aerodrome are supplied hourly or half-hourly and immediate notification is given when the weather conditions deteriorate or improve through certain limits which may be much more detailed than those which apply under (ii) above.

151. TRANSMISSION OF METEOROLOGICAL INFORMATION TO AND FROM AIRCRAFT IN FLIGHT

To facilitate the transmission of meteorological information from ground stations to aircraft in flight and the recording and reporting of observations made in flight, use is made of special forms, codes and radio broadcasts. Within the United Kingdom all exchanges between ground stations and aircraft are made in plain language on radio-telephony, though codes may be incorporated into the speech for the sake of brevity. These codes are:

(i) The Q code, which is used in flight for requesting information from a ground station; also certain transmitted information may be preceded by the appropriate Q-code group as a means of identification. Each group of the code consists of the letter Q followed by two further letters. Several of the groups refer to meteorological information but it is unnecessary to specify them here. Certain pressure groups have been described in Section 9.
(ii) The word CAVOK, which may be used in a weather report or forecast when the following conditions occur simultaneously:
  (*a*) Visibility: 10 kilometres or more.
  (*b*) Cloud: No cloud below 1500 metres (5000 feet).
  (*c*) Weather: no precipitation and no thunderstorm.

Radio-telegraphy may be used on non-domestic routes and reference is made below to the codes most commonly employed. Full details can be found in *Handbook of weather messages, Part II.*

METAR Used for aviation routine weather reports.

Symbolic form:

METAR GGgg CCCC $\text{dddff}/f_m f_m$

VVVV $(\text{R}V_R V_R V_R V_R/D_R D_R)$ $w'w'$ $(N_s C C h_s h_s)$

or

CAVOK

(Groups in brackets may be repeated.)

TAF Used for aerodrome weather forecasts.

Symbolic form:

TAF CCCC $G_1G_1G_2G_2$ $\text{dddff}/f_m f_m$

VVVV $w'w'$ $(N_s C C h_s h_s)$

or

CAVOK

$(0G_F G_F T_F T_F)$ $(6I_c h_i h_i t_L)$ $(5Bh_B h_B t_L)$ $9i_3nnn$

(Groups in brackets may be repeated.)

A brief decode of the groups in these two codes is given below:

GGgg	= Time of observation	CCCC	= Station indicator group
ddd	= Wind direction in degrees	ff	= Wind speed in knots
$f_m f_m$	= Maximum wind speed in knots		
VVVV	= Visibility in metres	R	= Indicator letter
$V_R V_R V_R V_R$	= Runway visual range in metres	$D_R D_R$	= Number of runway to which $V_R V_R V_R V_R$ refers
$w'w'$	= Present weather	$N_s C C h_s h_s$	= Cloud data
0	= Group indicator	$T_F T_F$	= Forecast temperature at time $G_F G_F$
6	= Group indicator	I_c	= Type of airframe ice accretion
$h_i h_i$	= Height of base of icing level	t_L	= Thickness of icing layer or of turbulent layer
5	= Group indicator	B	= Turbulence
$h_B h_B$	= Height of lowest level of turbulence	$9i_3nnn$	= Supplementary phenomena

AIREP Used for reports from aircraft; meteorological items are contained in the third section.

CODAR Used for the hemispheric exchange of meteorological data from aircraft.

152. GENERAL REMARKS

The success of any system for providing meteorological services to aviation depends to a large extent on co-operation between the meteorologist and those using the information provided. Thus any request for meteorological information should state specifically what is wanted. The difference between a weather report (conditions observed at a particular time) and a weather forecast (conditions expected over some future period) should be remembered. If a forecast is requested for a particular flight, the route should be clearly stated and time of take-off, height, and duration of flight should be given. As the preparation of a route forecast takes time, particularly if documentation is required, inquirers should ensure that they make their requests as early as possible, otherwise delays may occur. If a forecast has been received but there has been a delay in take-off, the aircraft commander should make sure that his forecast is still valid and, if necessary, obtain an amended forecast. Attention to such details will help to maintain a high standard of service.

This chapter has of necessity to be brief. More details concerning procedures and the codes used for transmission of data, etc. will be found in the relevant International Civil Aviation Organization publications, in the *U.K. Air Pilot* and in the *Handbook of weather messages.*

PART V

CLIMATOLOGY

CHAPTER 22

GENERAL CIRCULATION AND WORLD CLIMATE

153. INTRODUCTION

It is no part of the aim of the following chapters to present a detailed discussion of the climatic conditions all over the world. Although the qualified navigator should be in a position to undertake a flight to any part of the world, all that can be expected is that he should have sufficient knowledge to appreciate in a general way the sort of conditions likely to be encountered should a flight need to be undertaken at short notice. The airline pilot will of necessity be required to make a special study of particular routes and as a necessary professional qualification he must have a good general groundwork of climatological knowledge upon which to build. The present treatment is confined to this groundwork.

From the theoretical aspect the study of world climatology and the general circulation of the atmosphere is required as a background against which temporary and local phenomena, such as depressions and anticyclones, may be considered.

154. IDEALIZED GENERAL CIRCULATION

Were the surface of the earth everywhere level and uniform and were it not subjected to the variable heating effect of the sun, it might be supposed that the atmosphere would everywhere be in a state of complete stagnation. Under the influence of differential solar heating, gradients of temperature and pressure are, however, created and the whole atmosphere assumes a complicated circulatory motion. An examination of the charts of average winds over the earth's surface (Figs 149, 150) shows clearly a number of well-defined major air currents extending over large areas of the globe which may be regarded as reasonably permanent features although they may be liable to fluctuations and perturbations due to passing disturbances. Together with the associated upper winds and vertical currents, these major currents define what is known as the general circulation of the atmosphere. The physical explanation of the observed circulation is by no means a simple matter, even apart from the complications caused by the different conditions of temperature over land and sea and by variations in topography. It is therefore helpful to put on record what is believed would be the nature of the circulation if the rotating earth were entirely covered with a uniform ocean. It may appear unnecessarily artificial, but in the present discussion the idealized circulation forms a useful background against which observed facts may be considered, so bringing out more clearly the part played by the contrasts between land and sea. The system is based largely on the observed circulation over the southern hemisphere which is mainly covered by large stretches of unobstructed ocean; it is however not without theoretical foundation.

If the earth were a uniform globe, the average temperature would vary only with latitude, the lowest values appearing at the poles and the highest near the equator, corresponding to the differences in the amount of heat received from the sun. From what has been said in Section 7, it then follows that the pressure would be greater at any given height over the equator than at the same height over the poles, so that the upper air would tend to drift from the equator towards the poles. This in turn would lead to a rise of surface pressure in high latitudes and a fall in low latitudes with the consequence that, near the surface, air would begin to drift towards the equator, thus completing the circulation. The effect of the earth's rotation, however, in deflecting these currents to the right in the northern hemisphere and to the left in the southern hemisphere means that the circulation would take on a largely zonal character and would probably present the complicated structure presented diagrammatically in Fig 140.

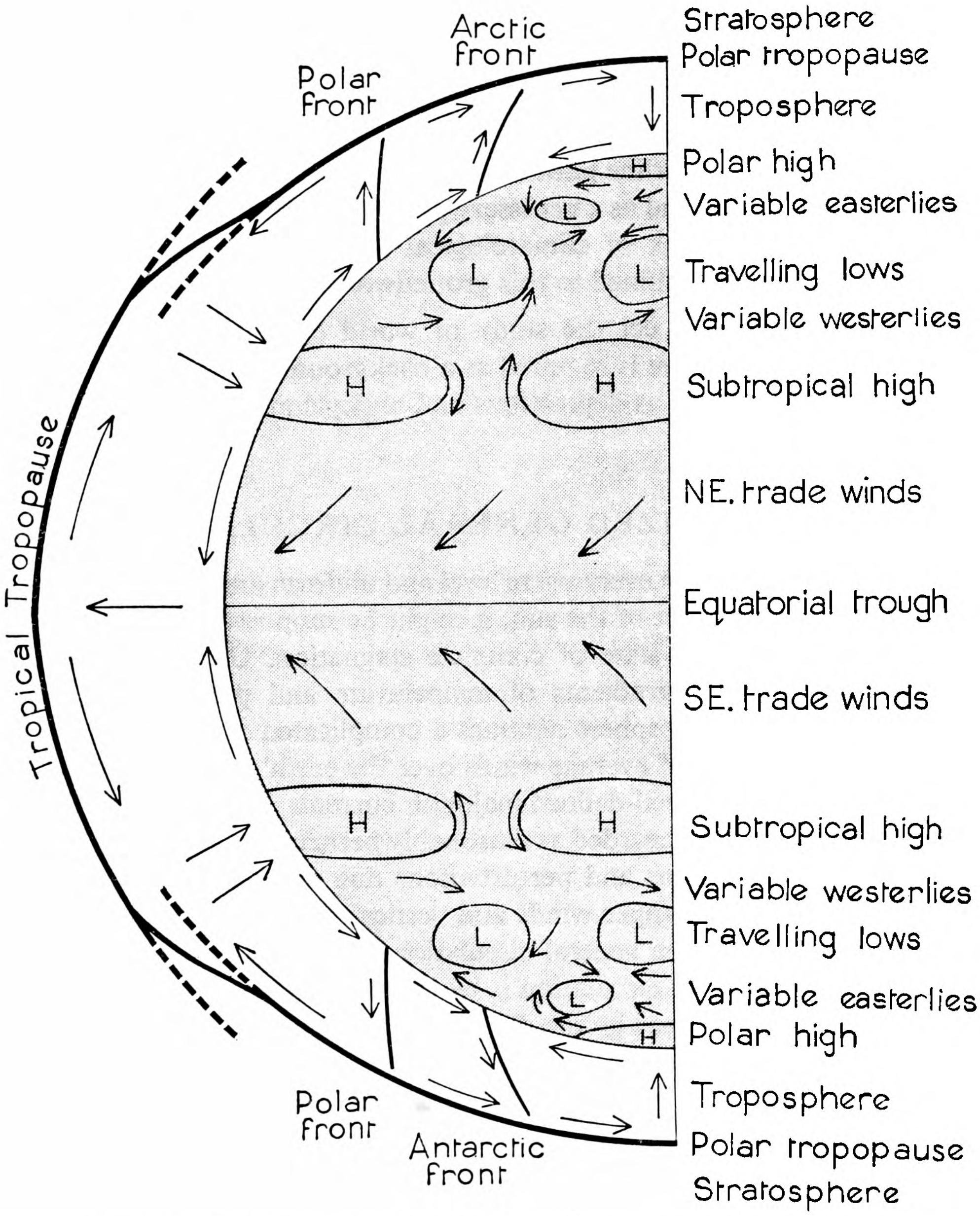


FIG. 140. *Idealized distribution of surface pressure over the earth*

Also shown is a section through the troposphere with idealized circulation and fronts in the vertical. The vertical scale is about 150 times the horizontal.

Moreover it may be noted that a circulation composed merely of zones of easterly and westerly winds, even on a uniform earth, is also impossible. One reason for this has been explained in Section 28 where it was seen that there is a net gain of radiant heat in low latitudes and a net loss in high latitudes. The maintenance of the existing distribution of mean temperature over the earth requires that heat be conveyed from the equator polewards by large-scale movements of air which together with the return currents imply exchange of air between different latitudes. There are also arguments of a dynamical nature which show that there can be no permanent unbroken belts of wind or pressure right round the earth. There can therefore be no simple zonal circulation; northerly and southerly winds must be present at times in all latitudes and their existence necessitates a complicated circulation with transient cyclones and anticyclones which accordingly constitute an essential feature of the general circulation.

Most of the features of the zonal circulation can however be recognized on the actual earth and the names of the various zones of pressure and wind are entered in Fig 140. These include the equatorial trough which is a zone of rising air fed by the trade winds from either side, which in turn consist of air which has subsided in the subtropical highs. In accordance with the preceding discussion, these highs cannot consist of permanent belts of high pressure encircling the globe but must occasionally break down in places to permit a transverse flow of air across them – this too accords with the facts. Also near the poles the inferred high pressure is in reality by no means a permanent feature. In middle latitudes, streams of warm air moving north-eastwards from the borders of the subtropical highs come into contact with outbursts of cold air from the polar regions and result in the formation of the familiar travelling frontal depressions in which the warm air ascends above the cold air.

To make the zonal scheme a still closer approximation to reality, allowance must be made for the seasonal variation in the apparent position of the sun. On account of the inclination of the equatorial plane to that of the earth's orbit round the sun (the ecliptic), the 'thermal equator' (the zone of highest temperature) moves away from the geographical equator into the summer hemisphere. As would be expected, the belts of tropical low pressure, subtropical highs and temperate westerlies oscillate similarly. Consequently certain transitional regions are affected by one wind system at one season and by a neighbouring system at another. In practice the range of oscillation over the oceans (excluding the exceptionally large range over the Indian Ocean) amounts to about 10 degrees of latitude for the subtropical highs and to about 15 degrees for the tropical low-pressure belt. The movements lag behind that of the sun by about two months.

155. BASIC CLIMATIC ZONES

The general characteristics of weather in the idealized zones of pressure and wind can be readily inferred. On account of dynamic cooling of the air by ascent, abundant cloud and precipitation forms in the temperate belts of low pressure and in the equatorial region, while predominantly arid conditions are maintained by subsidence at the poles and in the subtropical belts of high pressure. The seasonal movement of these systems produces transitional regions in those zones which come under the influence of one belt or another according to the season. Thus near the solstices the tropical rains are displaced into the summer hemisphere and we are led to speak of tropical transitional regions between the equatorial low and the subtropical highs, regions which are subject to tropical rain in summer and to dry trade-wind weather in winter. Nearer the equator there are two rainfall maxima at about the time of the equinoxes and two rainfall minima at about the time of the solstices, that is when the sun has respectively its greatest and least declination. The disturbances of the temperate zone extend furthest towards the equator in winter, so that regions on the poleward fringe of the subtropical high have winter cyclonic rains and a dry summer.

These elementary climatic zones have been inferred from the consideration of a uniform earth but they are nevertheless found to correspond widely although by no means entirely with actual conditions. The characteristics of the zones as found on the actual earth can now be set out in greater detail.

Polar climate

In the idealized scheme, the pressure distribution in the polar regions is anticyclonic and descending air gives dry quiet weather. In practice the anticyclone is not permanent but is frequently displaced by travelling depressions which give periods of unsettled weather and snow. The annual precipitation (in terms of equivalent rainfall) is mostly less than 250 millimetres. Where the mean temperature is below freezing-point, the surface is continuously covered with snow or ice. Where the mean temperature of the warmest months is above freezing-point, the 'tundra' type of climate prevails, characterized over land by the existence of vegetation consisting largely of mosses, lichens and grasses, while the subsoil remains frozen throughout the year.

Disturbed temperate climate

In this zone the weather is controlled by travelling frontal depressions and by the less frequent high-pressure systems. Winds are variable but mainly from a westerly point and gales are frequent. There is much precipitation of a cylconic type throughout the year, often augmented orographically. In winter it may fall as sleet or snow and there is no dry season.

Warm temperate (transitional) climate

This consists of disturbed temperate conditions in winter but arid conditions in summer on account of the poleward shift of the subtropical anticyclone. Thus the winters are cool and unsettled with cyclonic rain while the summers are warm and dry. This type is also known as 'Mediterranean' being commonly experienced in that region.

Arid subtropical climate

Under the continuous influence of subtropical anticyclones, skies are clear and the climate is warm and practically rainless. These conditions produce the typical desert climate, with the result that most of the great deserts of the earth lie within this zone.

Bordering the desert regions is found the semi-arid 'steppe' type of climate in which rainfall occurs in a short season in either winter or summer depending on whether the area lies north or south of the neighbouring desert. This makes possible a restricted growth of vegetation which often takes the form of grassy, treeless plains.

Tropical transitional or savannah climate

The weather of this type in winter consists of dry trade-wind conditions but in summer it is determined by the belt of equatorial rains. The duration of the wet period decreases as the latitude increases, until the type merges into the semi-arid or steppe type.

Equatorial climate

This comprises two main wet seasons associated with the passage of the sun north and south across the equator, but there is no really dry season. There is much convective cloud, and rain falls in heavy showers with frequent thunderstorms. Both temperature and humidity are high and almost uniform throughout the year.

The above somewhat idealized description of climatic zones will now be used as a background to facilitate further description of the observed conditions. Divergencies from the purely zonal arrangement are to be attributed to the disturbing influence of the distribution of land and sea and of topography generally. The predominating disturbance arises from the effect of the different thermal properties of ocean and land on the temperature distribution. We consequently begin with a description of the actual distribution of temperature over the earth and then proceed to discuss other elements in turn.

156. SURFACE AIR TEMPERATURE

The broad features of the distribution of air temperature over the earth's surface may be considered as a function of latitude with temperatures decreasing gradually from equatorial to polar regions but appreciably modified by the distribution of land and sea, the whole being subject to variation according to the season of the year. The mean isotherms for January and July, given in Figs 141 and 142, may be taken as expressing the extremes of winter and summer and it is sufficient for a preliminary study to confine attention to these months; the distribution at other seasons is in general intermediate between the two.

The highest mean temperatures occur mainly within the tropics. The northern summer is somewhat hotter than the southern summer. In July the 30°C isotherm includes a substantial area of northern Africa, southern Asia and the southern portion of the North American continent; the value rises to over 35°C over the centre of the Sahara desert. In January the 30°C isotherm encloses only small areas over southern Africa and western Australia.

The temperature decreases towards the poles but the isotherms run sensibly parallel to the lines of latitude only over the large oceans of the southern hemisphere. Elsewhere they are much distorted in passing from land to sea, the general behaviour being in accordance with the rule that the land is warmer than the sea in the same latitude in summer and cooler in winter. The effect is particularly noticeable in the northern hemisphere where the lowest temperatures of winter are displaced from the pole to the land areas of northern Siberia, Greenland and northern Canada. It is in these same regions that the annual variation shown in Fig. 143 (the difference between the mean temperatures of the hottest and coldest months) is largest, exceeding 60 degC over northern Siberia where the summer temperatures are comparable with those of the British Isles, while the average winter temperatures are down to −45°C. In contrast, the climate of equatorial regions is extremely equable, with an annual range generally less than 3 degC.

The mean daily range of temperature, that is, the average difference between the day maximum and the night minimum, is shown in Fig. 144. It illustrates some important climatic differences between land and sea and between coasts and interiors of the continents.

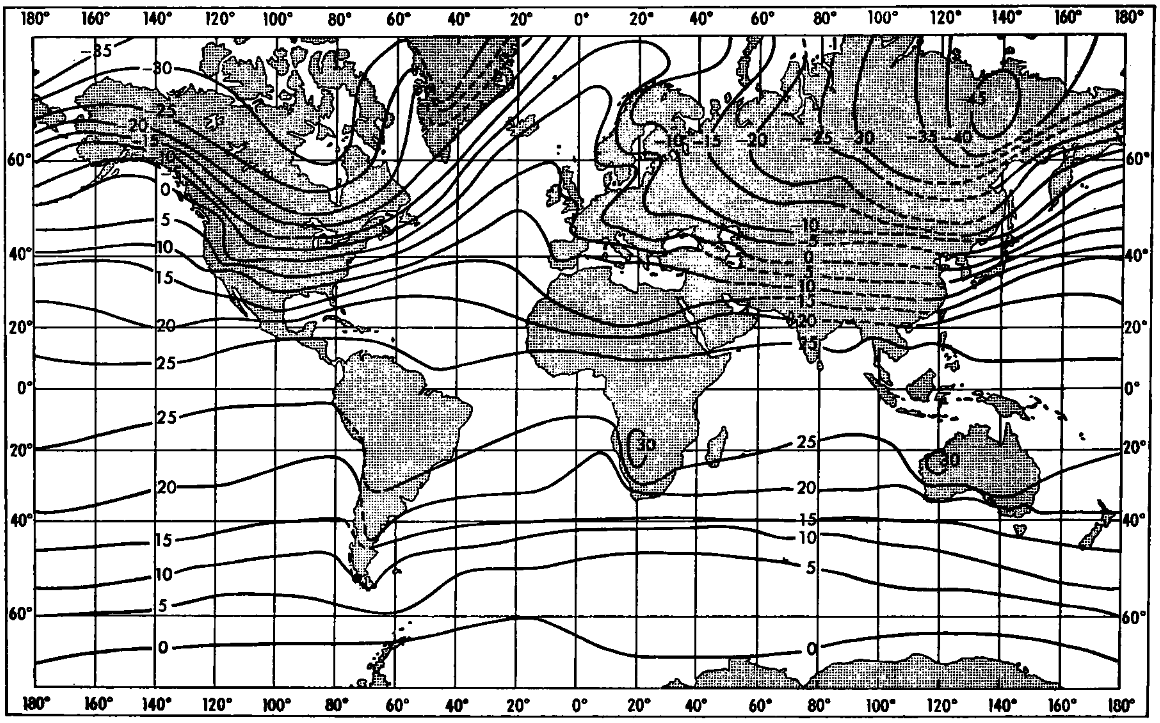


FIG. 141. *Average temperature (°C) at mean sea level in January*

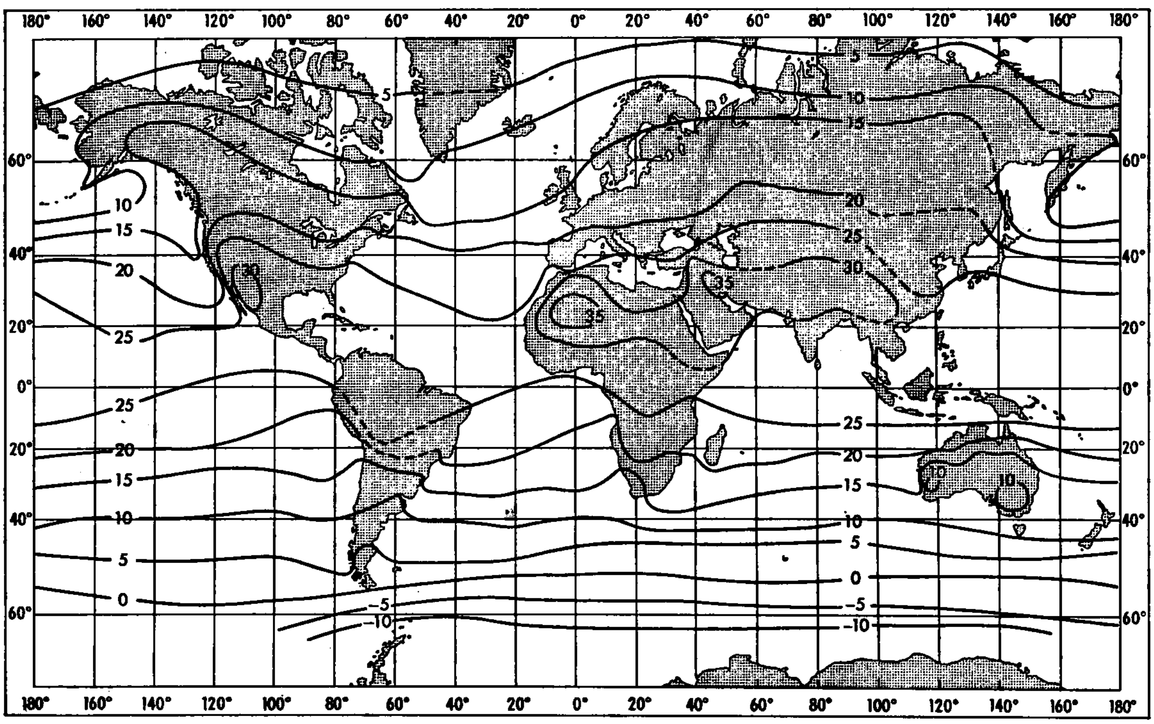


FIG. 142. *Average temperature (°C) at mean sea level in July*

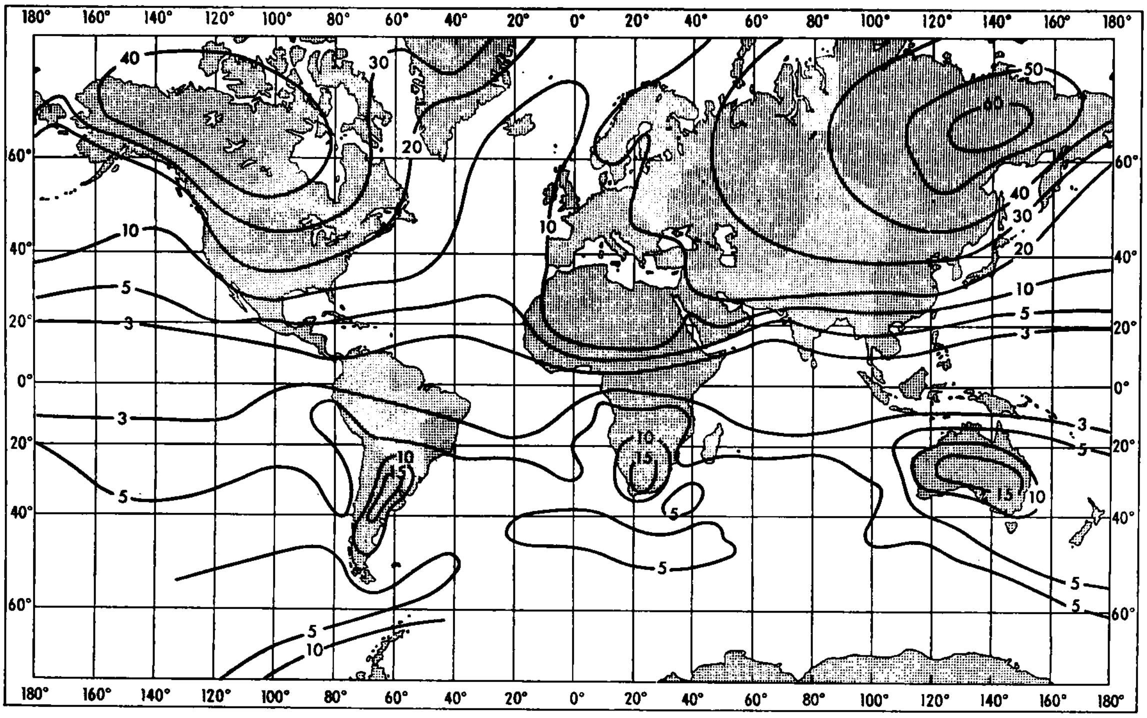


FIG. 143. *Annual range of mean monthly temperature (°C)*

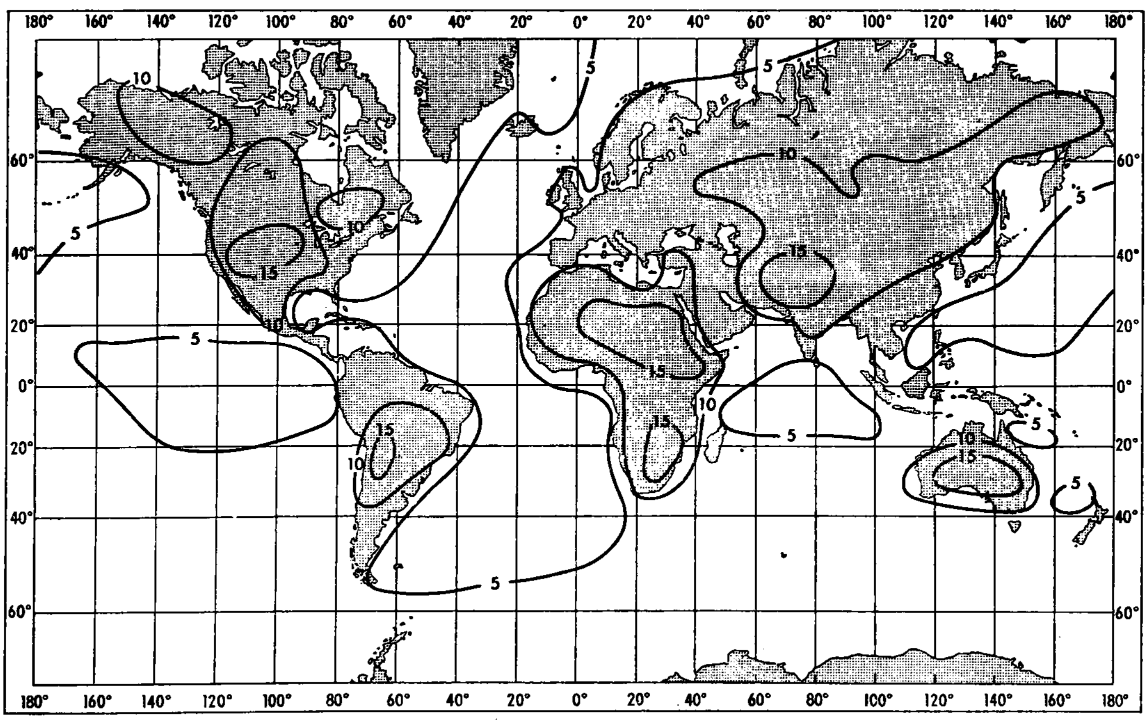


FIG. 144. *Mean daily range of temperature (°C)*

Many other lessons are to be learnt from the charts of mean temperature. Of significance is the crowding together of the isotherms (indicating a steep gradient of temperature) which occurs in the winter on the eastern side of the continents particularly in the northern hemisphere. The greater frequency of depressions and their greater intensity in winter than in summer are largely due to these contrasts. Also worth noting is the striking difference between the run of the isotherms across North America and across Europe and Asia in summer. From the Pacific Ocean to the American continent there is a rapid rise in mean temperature and the hottest part of the continent is in the west. From the Atlantic across Europe to Asia the change is very gradual and the highest summer temperatures are in the east. The explanation may readily be provided when consideration is given to the prevailing winds and to the barrier which the Rocky Mountains present to the cool maritime winds from the west, a barrier which has no counterpart in the Old World. The tempering influence of the Pacific is thus confined to the coastal regions of western North America while that of the Atlantic extends far into the European continent. Similarly in winter the Himalayas effectively confine the cold continental air to the north and prevent any really cold weather over India. North America, having no transverse range, has a completely different régime. The warm moist air from the Gulf of Mexico is able to penetrate northwards into Canada, and cold outbreaks of Arctic air can sweep unobstructed across the entire continent in the opposite direction.

157. UPPER AIR TEMPERATURE

The average decrease of temperature with height has been seen to be 2 degC per 1000 feet throughout the troposphere while in the lower stratosphere the temperature increases or changes little with height. Of particular importance for aviation are the heights of the isotherms of 0°C and —40°C (these being the limits within which nearly all cases of airframe icing occur) and the height of the tropopause, for the reasons summarized in Section 72.

Height of the 0°C isotherm

This height is important in relation to the risk of airframe icing. The distribution of the mean height over the world is shown in Figs 145 and 146 for January and July respectively. In January the height is about 16 000 feet near the equator while in July it exceeds 18 000 feet in parts of southern Asia. Polewards the height decreases and in the northern hemisphere it first coincides with the earth's surface in latitudes which differ widely between ocean and continent in a manner reflecting the large variations of surface temperature. The contrast with the more uniform lines of the southern hemisphere is noteworthy.

Height of the —40°C isotherm

This is the height above which airframe icing becomes rare. In January the mean height increases from about 16 000 feet in the Arctic to some 35 000 feet at the equator; in July the range extends from about 25 000 feet in the Arctic to 38 000 feet over north-west India.

Height of the tropopause

Some figures of the mean height of the polar and tropical tropopauses are given in Table 12. The two tropopauses are usually discontinuous in the neighbourhood

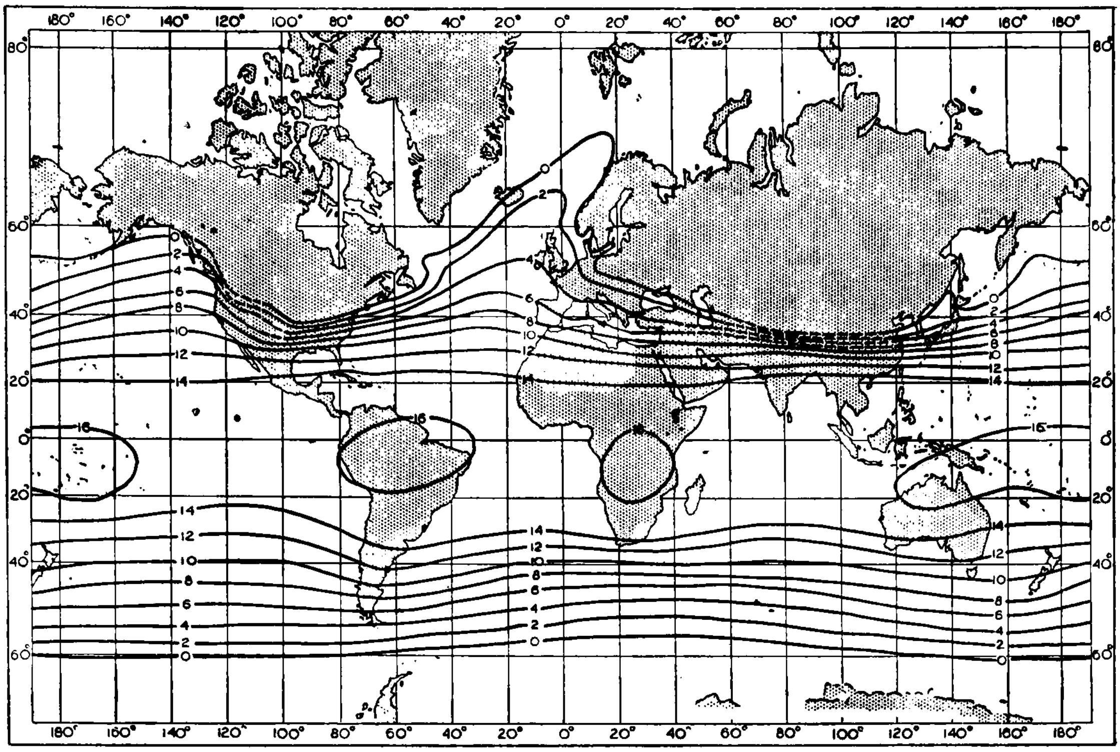


FIG. 145. *Height (in thousands of feet) of 0°C isotherm in January*

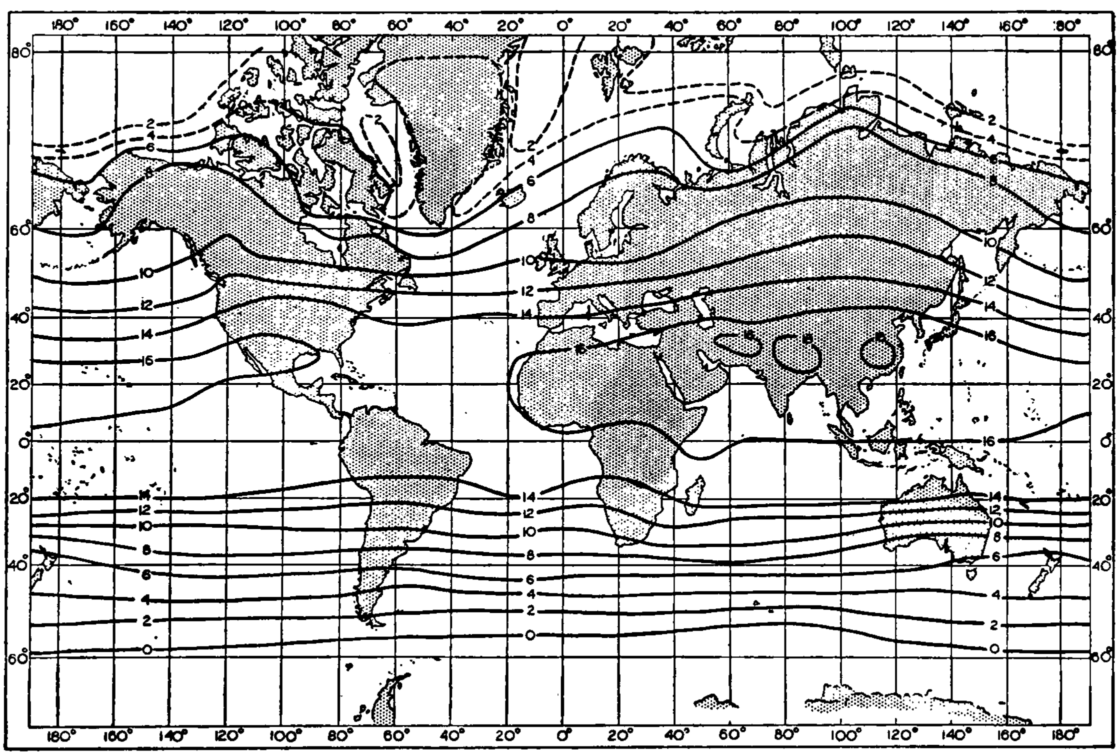


FIG. 146. *Height (in thousands of feet) of 0°C isotherm in July*

of latitude 40° and overlapping often occurs as illustrated in Fig. 140. In addition to the variations with longitude shown in this table, mention should be made of the low January value (about 28 000 feet) of the polar tropopause near Sakhalin (50° N 140° E), while in July the values are high for the latitude over both Siberia and Alaska.

TABLE 12. *Mean height of polar and tropical tropopause*

	90°W		0°		90°E		180°	
Lat.	Polar	Tropical	Polar	Tropical	Polar	Tropical	Polar	Tropical
				feet				
(*a*) January								
°*N*								
70	27 000		31 000		29 000		29 000	
60	27 000		32 000		31 000		30 000	
40	35 000	57 000	37 000	54 000	35 000	51 000	35 000	54 000
20		56 000		56 000		57 000		56 000
0		56 000		57 000		57 000		56 000
°*S*								
20		54 000		55 000		55 000		54 000
40	38 000	52 000	38 000	52 000	38 000	51 000	38 000	51 000
60	30 000		29 000		29 000		31 000	
(*b*) July								
°*N*								
70	32 000		33 000		36 000		34 000	
60	35 000		35 000		39 000		35 000	
40	43 000	51 000	43 000	52 000	43 000	51 000	42 000	51 000
20		53 000		56 000		55 000		53 000
0		54 000		55 000		54 000		53 000
°*S*								
20		54 000		54 000		53 000		53 000
40	34 000	49 000	34 000	49 000	34 000	49 000	34 000	49 000
60	31 000		31 000		31 000		31 000	

158. PRESSURE

Corresponding to the temperature charts of Figs 141 and 142, charts of average sea-level pressure for the two months January and July are given in Figs 147 and 148. In examining these charts it is helpful to note to what extent they agree with the pressure distribution of the idealized circulation of Fig. 140. The equatorial zone of relatively low pressure extends through all longitudes in January; it lies rather to the south of the equator with the lowest pressures over Africa, Australia and South America. In the northern summer the zone shifts well to the north of the equator (corresponding to the seasonal oscillation in the relative position of the sun) but could hardly be described as a continuous belt. It is evident enough over the Atlantic and Pacific Oceans and also across South America, but across east Africa and the Indian Ocean the low-pressure centre over south-west Asia completely displaces the idealized system.

At all seasons the southern hemisphere manifests the ideal features quite closely. The subtropical high-pressure zone appears as a well-defined belt in July, centred in about latitude 25–30° S, but with some variation in intensity. In January the belt is some 5° or 10° further south and is more definitely broken by the tendency for lower pressures over the heated continents. From this high-pressure belt there is a steady and continuous fall of pressure to approximately 60° S, after which pressure would be expected to rise again slowly in the Antarctic continent. Observations are insufficient to define the mean isobars accurately on the continent, the average height of which is about 7000 feet, but there is little doubt that average sea-level

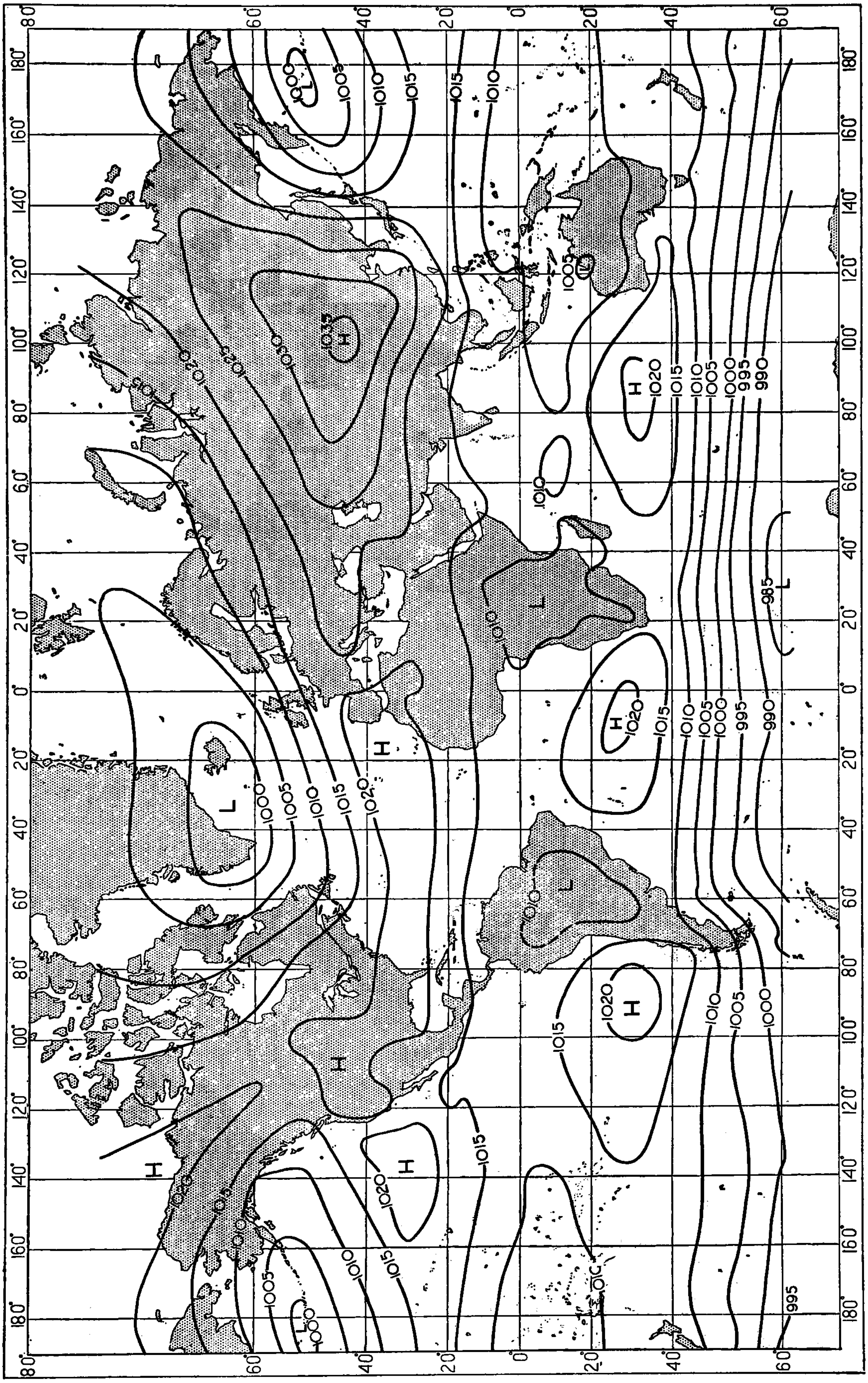


FIG. 147. *Average pressure (millibars) at mean sea level in January*

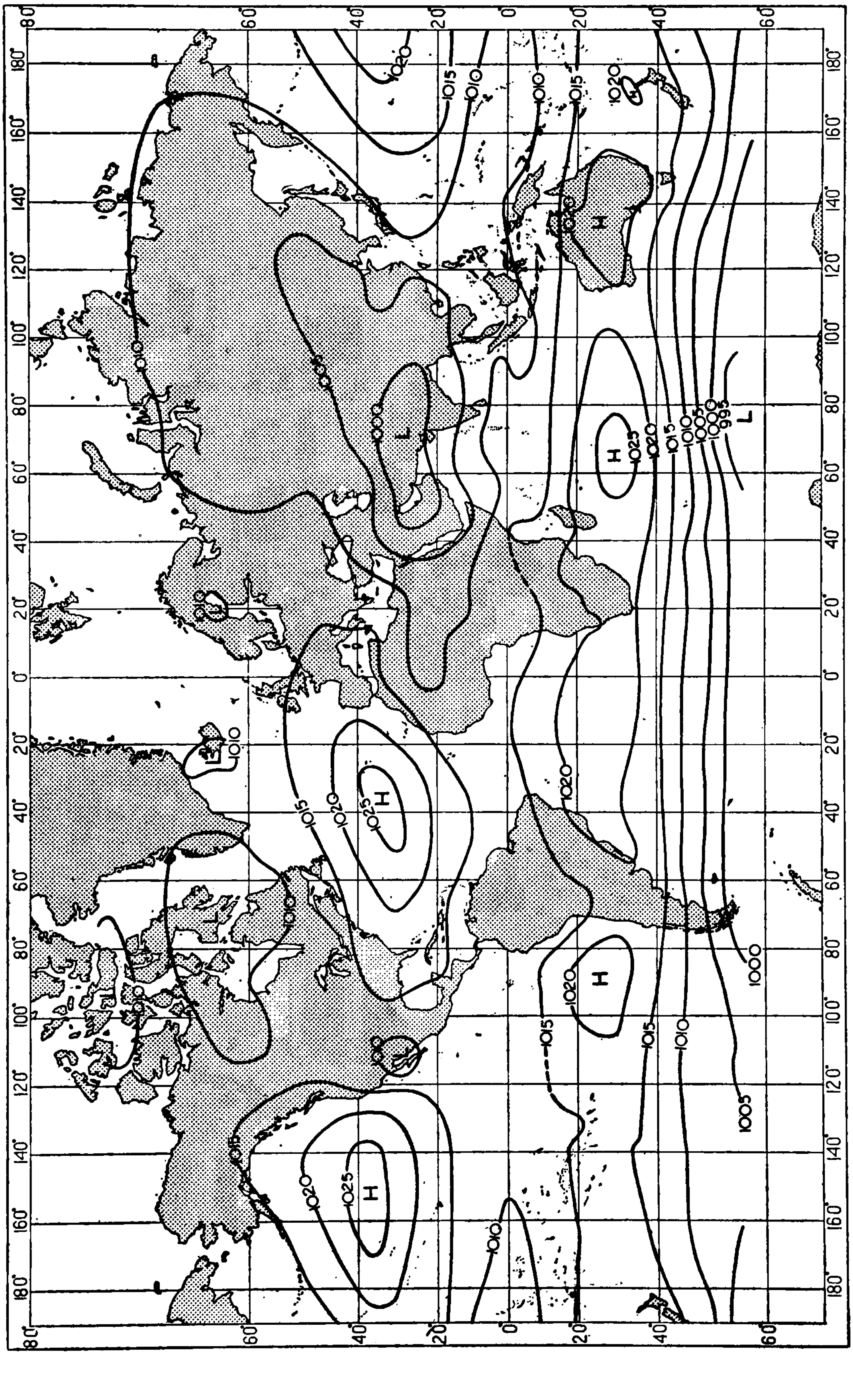


FIG. 148. *Average pressure (millibars) at mean sea level in July*

pressure there is relatively high, as required by the ideal circulation. In the belt of low pressure centred at about 60° S there are frontal depressions which move from west to east at all times of the year.

In the northern hemisphere there are large departures from the ideal circulation. In January the high-pressure areas are found in subtropical latitudes over the Atlantic and the Pacific, but they are centred in higher latitudes over the two great land masses. The highest pressures are to be found in the centre of Asia and again over central North America at about 45° N. On the other hand the low-pressure belt, which in the southern hemisphere occurs near the 60th parallel, degenerates in the northern hemisphere into two well-defined centres located over the northern Atlantic (with a tongue extending eastwards into the polar seas) and over the northern Pacific. These two dominating features over the relatively warm oceans, known respectively as the 'Icelandic low' and the 'Aleutian low', are all that is left of the low-pressure belt after the breaks caused by the cold continental anticyclones. The polar region shows no very definite pressure characteristic but is covered by a col between the continental highs and the oceanic lows — the theoretical polar anticyclone fails to appear as a permanent feature.

In July in the northern hemisphere the subtropical high-pressure belt reduces to two well-defined centres in the Atlantic and Pacific Oceans centred about 35°N. The same latitudes over the continents are covered by the monsoon depression of south-west Asia and the less deep but still significant seasonal low over the west of the U.S.A. Pressure remains, on the whole, low over the northern North Atlantic with recognizable centres near the Arctic Circle, but the pressure distribution in high latitudes shows no dominating features at all comparable with the subtropical oceanic highs and the monsoon low.

It is important to remember that the field of pressure represented on the charts is only an average distribution expressing the statistical result of day-to-day variations. In tropical regions these variations are slight and the seasonal chart closely represents the daily chart except on the relatively rare occasions when, in certain localities, tropical cyclones develop. In temperate and polar latitudes, however, the average pressure distribution often bears no relation to that appearing on a synoptic chart. Temporary anticyclones appear where the pressure is on the average low, while depressions may invade the high-pressure regions. The Asiatic anticyclone is the most stable feature of the northern winter, although even that is not immune from temporary disturbances. The corresponding high over North America is much less persistent and this area is very frequently invaded by depressions or troughs of low pressure moving from west to east. Even the subtropical high-pressure systems, in both hemispheres, experience numerous perturbations and, although the pressure variations are not as pronounced as in higher latitudes, they are very significant for daily weather and must be carefully analysed for the purposes of forecasting. These high-pressure belts usually consist, at any one time, of from four to six individual anticyclonic cells separated by cols in which, however, the pressure is still relatively high.

159. SURFACE WIND SYSTEMS AND ASSOCIATED WEATHER

The prevailing winds for January and July are shown in Figs 149 and 150. On comparing these with the pressure charts it is evident that the two are related, as would be expected, in accordance with the form of Buys Ballot's law appropriate

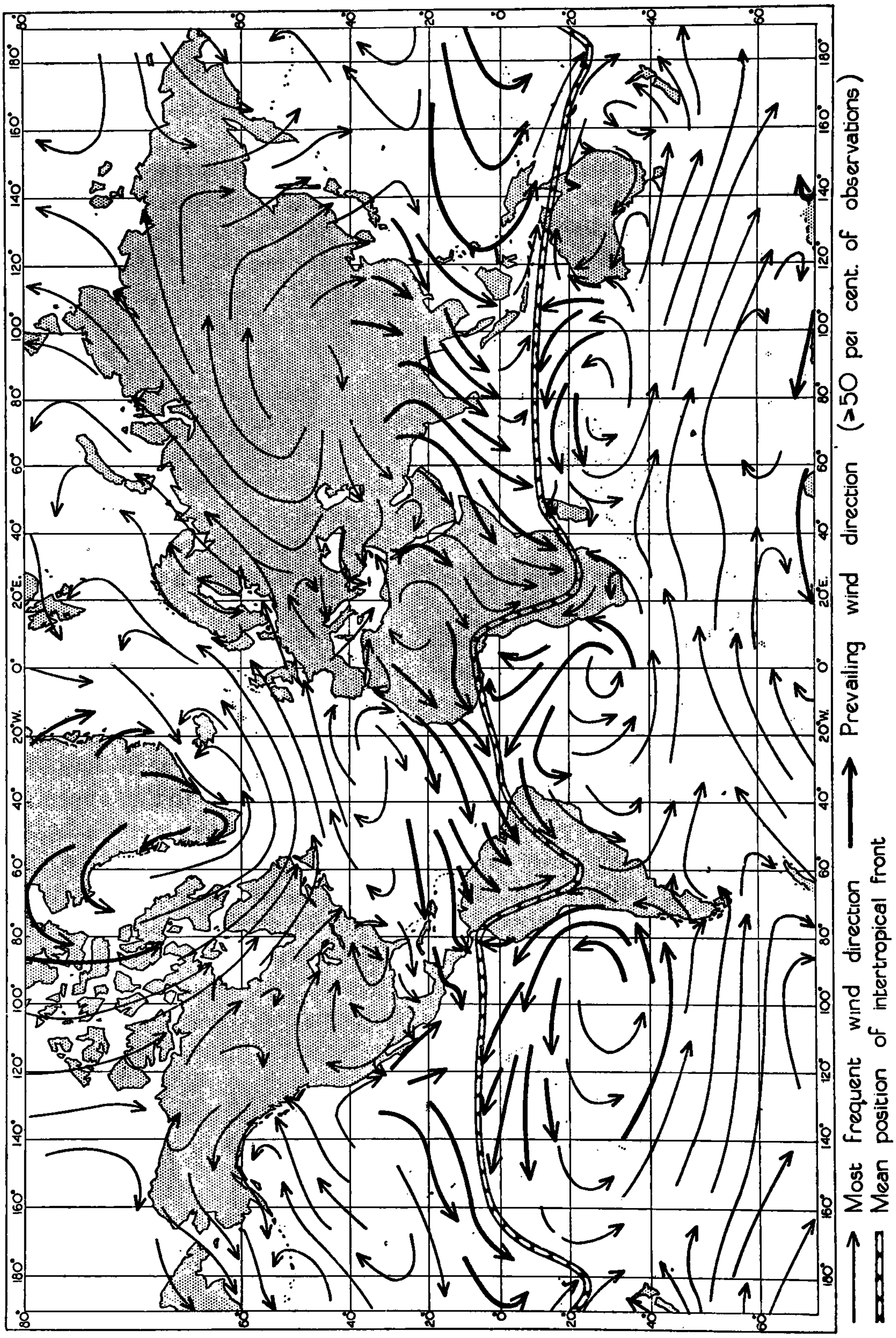
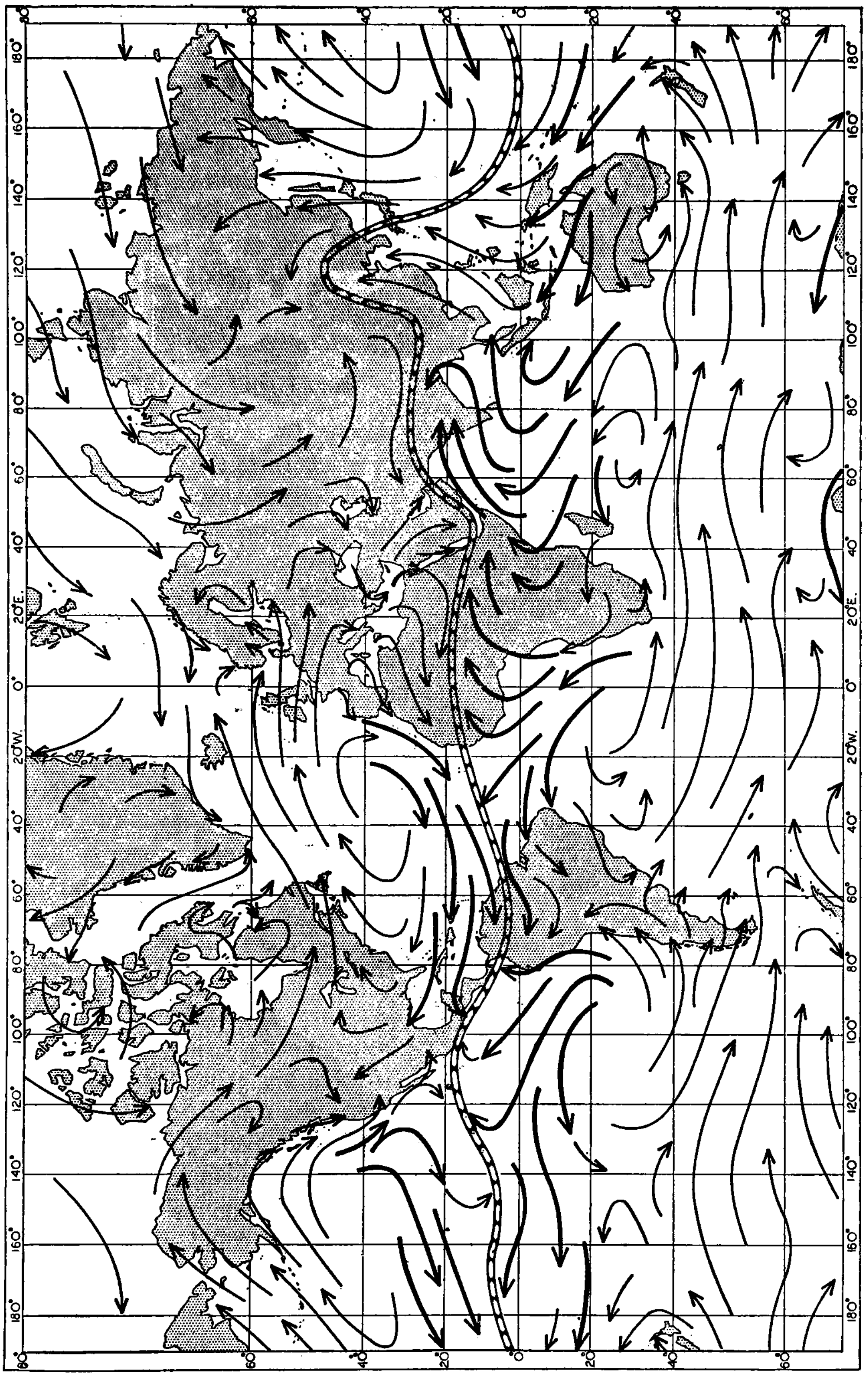


FIG. 149. *Prevailing surface winds in January*



→ Most frequent wind direction ⟶ Prevailing wind direction (>50 per cent. of observations)

Mean position of intertropical front

FIG. 150. *Prevailing surface winds in July*

to the hemisphere. The agreement is however not very close in low latitudes, and very near the equator the relationship fails completely. This again would be expected for we have seen (Section 24) that the balance between the pressure gradient and the deviating force of the earth's rotation cannot become effective in very low latitudes and that the geostrophic wind is then no longer a useful indication of the true wind.

The main wind currents are the trades, the monsoons and the prevailing westerlies of temperate latitudes. Outside these broad streams the winds are much more variable in direction, mainly light with dry weather in the anticyclonic areas, often squally with disturbed weather in the low-pressure areas – although there are in most parts interruptions in the prevailing type. In the southern hemisphere the polar easterlies of the idealized circulation are found near the coasts of Antarctica on most of the few occasions when observations are available. In north polar regions the winds depart very considerably from the zonal scheme and persistent easterlies are not typical.

Equatorial trough

In the idealized circulation of Fig. 140 the equatorial region is marked by a zone of relatively low pressure in which the trade winds from the two hemispheres approach one another, there to ascend and return polewards at higher levels. This scheme agrees well with the average conditions over parts of the oceans, but less so over the continents where the trades are affected by the monsoons. On the mean charts (Figs 149–150) the zone may be identified in July in about latitudes 5° N to 15° N over the Pacific and Atlantic oceans; in January it is found nearer the equator over the Atlantic and eastern Pacific, but south of the equator elsewhere. Since the equatorial trough is a zone of confluence, zones of horizontal velocity convergence may occur, and if the airstreams have different densities, conditions may be favourable for the formation of a front. Over the continents a temperature or humidity discontinuity can often be found, but over the oceans such differences between the trade winds of the two hemispheres are small and generally no discontinuity can be observed. Thus the zone of confluence between these two wind systems cannot strictly be represented as a continuous air-mass boundary, even a diffuse one. The term 'intertropical front' (ITF) can be used to describe this zone, but it is also known as the 'intertropical convergence zone' (ITCZ).

The so-called intertropical front differs in many ways from the fronts of temperate latitudes. Experience has confirmed the existence of organized lines of cumulus or cumulonimbus and belts of cirrostratus which resemble the cloud systems found with the fronts of higher latitudes, but other features are generally lacking. For example, there is little evidence that over the oceans the intertropical front has a clearly defined frontal surface, as with many fronts in temperate latitudes; over the continents differences of temperature may be found in the upper air which are related to the different lapse rates in the air on either side of the intertropical front, but the behaviour of the 'front' in space and time makes it difficult to assign to it the same role that extratropical fronts play in a high-latitude depression. In the tropics, air masses with a long sea track are very moist in the lower layers from evaporation, but they may be much drier aloft if they originate as the subsiding air of the subtropical anticyclones. When such air masses are lifted, convective instability develops (Section 35) and large cumulus and cumulonimbus clouds form, giving rise to showers, line-squalls and thunderstorms. Thus although it cannot be definitely asserted that there is an intertropical 'front', it is tempting to assume the existence of a front in those regions of upward motion of which the intensity and horizontal extent vary with time and place according to the circumstances of the convergence. Such convergence may have one or more of several causes – dynamic, orographic, or simply diurnal heating. The tops of the cumulus and cumulonimbus spread out at levels where the lapse rate is more stable and form extensive sheets of altostratus and cirrostratus, while other formations may develop independently from instability in the middle and upper levels. Several such layers of middle and high cloud may be present at the same time and rain may fall from the altostratus over wide areas. The width of the belt of disturbed weather varies according to the scale of the convergence; observations have shown that the width may range from about 25 to 300 miles. Outside the clouds, visibility is good except in heavy rain where it may be reduced to a few yards. The cloud base is usually 1000 feet or more above the sea but it may descend practically to the surface in heavy rain. The surface wind in the vicinity of the line of disturbance is squally, often reaching force 4 or 5 while squalls of force 8 or 9 are occasionally reported.

On occasions of vigorous development, some large rain areas occur and thick masses of nimbostratus and altostratus are found with violent turbulence within the clouds. Usually, however, conditions are somewhat patchy and there are areas of lighter rain and broken cloud. In low-level flight, if no clear lane can be found it is usually considered best to cross the line of disturbance at the lowest practicable level within 1000 feet of the surface of the sea. At such levels, violent turbulence is then much less likely to be encountered.

The height of the cloud tops in the vicinity of the lines of disturbance varies considerably according to circumstances. On many occasions the belt of cloud may be flown over without difficulty at 20 000 feet but on occasions of active development cloud should be anticipated well beyond that level, in extreme cases even up to the tropopause, that is, in the neighbourhood of 55 000 feet. The spreading out of the cloud tops into extensive sheets of cirrus occasionally conceals the cumulonimbus turrets with their severe turbulence; at other times the cirrus sheets may persist long after the convective cloud has dispersed. Ice accretion in active areas is liable to be serious in convective cloud at temperatures from 0° to —40°C or even lower. The corresponding heights at which these temperatures are found in the equatorial regions are about 16 000 feet and 35 000 feet respectively but are greater by about 2000 feet over southern Asia during the northern summer. Day-to-day variations may also cause fluctuations of a few thousand feet. It is clear that the clouds should, if practicable, be avoided even at high levels, but that if they are entered then suitable precautions should be taken against turbulence and icing.

Doldrums

During periods of feeble development, the zone of confluence of the trades from the two hemispheres degenerates into an area of detached cumulus clouds and scattered showers. Often there is a broad area of very light winds and mainly dry weather, and the term 'doldrums' is sometimes used in a special sense to describe these areas, although in the past it has been used in a wider sense to include the whole zone of confluence of the trades.

Tropical revolving storms

The characteristics of these violent cyclonic disturbances were described in Section 118. Most of these storms originate in the zone of confluence of the trade winds; the main localities affected are indicated in Fig. 151 where some representative tracks are also shown. Occurrences outside these regions are not unknown particularly in the northern Pacific where they have been observed in most longitudes. They are known as cyclones in the Bay of Bengal and in the Arabian Sea, hurricanes in the West Indies and south Indian Ocean, typhoons in the China Seas and willy-willies in the vicinity of western Australia. The South Atlantic is the only tropical ocean in which no storms have been recorded – perhaps because the intertropical front is not found there.

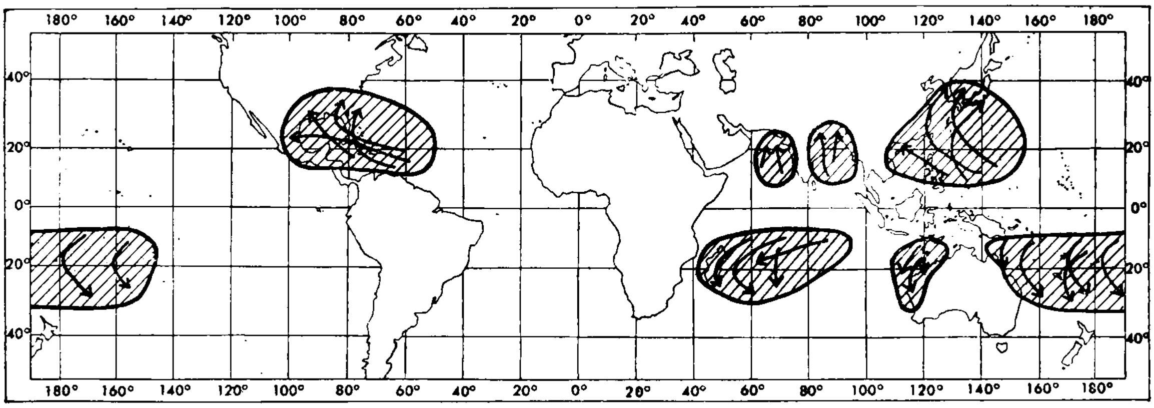


FIG. 151. *Main regions affected by tropical revolving storms, with some representative tracks*

Over the open ocean the storms originate within 5–15° of the equator and move slowly westwards with a velocity generally less than 15 knots. Sooner or later they usually curve away from the equator and, after crossing the tropic, follow an easterly track. In other words they skirt round the subtropical oceanic anticyclones, but irregular tracks are not uncommon. The longer the life of the storm the larger becomes the affected area until after reaching temperate latitudes the disturbance either fills up or continues as an ordinary depression.

Frequency tables of the occurrences in different parts of the world are available but in some areas the figures depend so much upon whether individual storms have chanced to cross a shipping route or an important island that they do not afford a basis for reliable comparison. In the China Seas they occur perhaps 20 times a year over the area as a whole, near the West Indies, in the south Indian Ocean and in the Bay of Bengal perhaps 10 times, while elsewhere they are only occasional.

Tropical storms are principally a phenomenon of the summer and autumn months – from June to October in the northern hemisphere and from December to April in the southern hemisphere. They usually reach their maximum frequency in August and September or January and February according to the hemisphere, although in this respect the Bay of Bengal and the Arabian Sea are anomalous, the storms being there more common with the advancing and retreating monsoon than at its height. Tropical storms are almost or entirely unknown in late winter and early spring.

Trade winds

By the trade winds are understood the steady winds which blow on the equatorial side of the subtropical high-pressure regions in both hemispheres. In the southern hemisphere the trade wind is well defined at all seasons. It is for the most part a south-easterly current but divergences from this direction are evident in some localities in agreement with the distribution of pressure. Thus on the west African coast the wind is deflected to a south or south-westerly direction while in places on the opposite side of the Atlantic the flow becomes north-easterly; there are similar effects in the other oceans.

The trade wind of the northern hemisphere – the north-east trade – covers a wide stretch of the Pacific. In the northern winter it crosses the equator in the vicinity of the East Indies and extends, as the north-westerly summer monsoon, to the coast of Australia. In the Atlantic this trade wind extends from the coast of Morocco nearly to the equator in January but lies some 10° further north in summer. The monsoonal variation near the Gulf of Mexico turns the trade wind into a south-easterly current in summer, an effect similar to that over the western Pacific. In the Indian Ocean there is no north-east trade wind proper but only the winter monsoon wind from that direction.

The trade winds are remarkably steady and persistent, hence their name which comes from the expression 'to blow trade' meaning to blow constantly from one direction. Since they consist of air which has subsided in the subtropical high, an inversion with its base between about 3000 feet and 8000 feet usually limits the cloud type to fair-weather cumulus and stratocumulus. There are however occasional disturbances in the form of weak belts of cloud and showers; moreover, cloudiness increases as the equatorial trough is approached and as the length of the sea track increases – thus the western oceans are more cloudy than the eastern in the trade-wind belt. Over mountainous country orographic ascent may result in copious rain. Thus the south-east trade brings rain all the year round to the eastern coasts of Madagascar and in summer to South Africa. The north-east trade is a rain-bearing wind over the mountains of the West Indies and again over the Hawaiian Islands which, at 20° N, are in the heart of the trade-wind belt. Generally speaking, trade-wind rain, when it occurs, is caused by a combination of orographic ascent and vertical instability set up by surface heating over the land. It is of a local showery character, often thundery, and falls mainly during the day.

Monsoons

The tendency in middle latitudes for high pressure to develop over the continents in winter and for low pressure in summer is reflected in persistent seasonal winds known as monsoons, a name which is derived from the Arabic word for season. These winds are most developed over southern and eastern Asia.

In winter the outflow from the Siberian anticyclone appears as a north-westerly wind across the Pacific coasts, north-easterly or northerly across southern China, Burma and India, and continues across the Indian Ocean to meet the south-east trade south of the equator. Although it is a remarkably constant wind it is not entirely free from disturbances. Both central China and north-west India are affected by shallow winter depressions which cause breaks in the monsoon. The Siberian anticyclone begins to give way in March when continental temperatures are rising rapidly, but the south-west monsoon of India does not arrive until June. Over the western Pacific the direction of the summer monsoon is southerly or south-easterly.

The winter monsoon of Asia is a cool or cold wind with dry weather predominating except where the wind has a long sea track. During the summer, moist equatorial air moves north and reaches India as the south-west monsoon. The onset and retreat of the monsoon is shown in Fig. 152. The reason for the sudden onset of the monsoon and the change in pattern of the upper winds is not fully known. Large changes occur in the high-level wind flow associated with the onset of the south-west monsoon. The axis of the subtropical westerly jet stream shifts quite suddenly from south to north of the Himalayas in early June and the high-level easterly winds spread north across India during May and June (compare Figs 156 and 161 which show streamlines at 200 millibars in January and July).

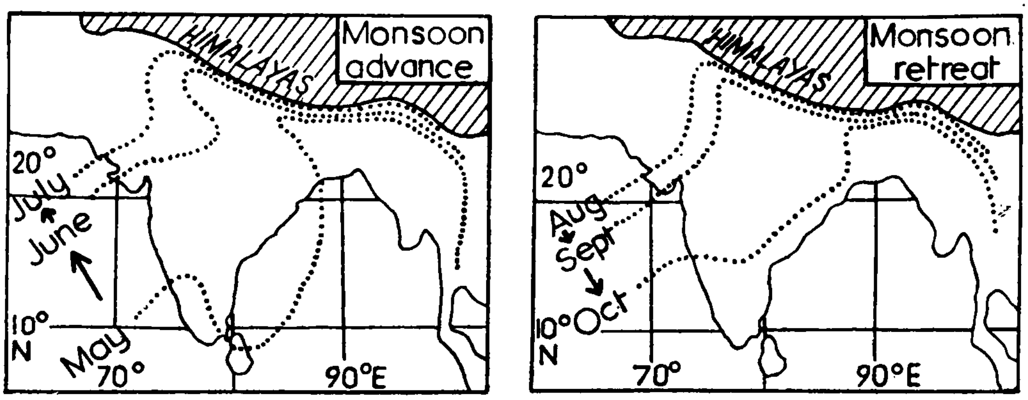


FIG. 152. *The advance and retreat of the summer monsoon*

Brief mention follows of important examples of monsoonal winds prevailing in other parts of the world. Over north-east Australia, the winter south-easterly trade wind gives way to a north-easterly in summer. Over Africa the south-east trade wind of the South Atlantic extends far inland north of the equator as the south-west monsoon of the northern summer; on the tropical east coast there is a change from north-easterly to south-easterly or southerly winds between the winter and summer of the northern hemisphere – this is partly the effect of pressure changes over Asia. Over North America the monsoonal change is not as complete as over Asia but there is a marked change in the prevailing winds particularly to the north of the Gulf of Mexico. Across the Atlantic seaboard the seasonal change is from north-westerlies in winter to south-westerlies in summer although neither current is a steady one, being much disturbed by travelling depressions.

Temperate westerlies

Between the subtropical high and the low-pressure systems of higher latitudes, westerly winds prevail. In the southern hemisphere they extend in a belt around the earth between latitudes about 40°S and 60°S, extending rather further equatorwards in winter than in summer. The winds vary in direction as depressions move along from west to east but they are mainly from a westerly point and frequently of gale force, giving the name to the 'roaring forties'. In the northern hemisphere the belt of westerlies is broken in winter by the continental anticyclones and it is therefore found mainly over the oceans on the southern side of the Icelandic and Aleutian low-pressure systems. Weather conditions are very disturbed and gales are common; the winds are more variable than in the southern hemisphere. In summer the westerlies of the northern hemisphere are on average much less strong than in winter and have little constancy, but they nevertheless reach gale force at times under the influence of depressions.

160. UPPER WINDS

From the accumulated observations of wind and pressure in the free atmosphere, charts showing the mean distribution of upper winds over the earth have been compiled. These mean winds are to a considerable extent derived from the pressure distribution according to the geostrophic rule, supplemented by radar observations which have become available in recent years and which are particularly valuable at high levels and also in the tropics where the geostrophic rule does not apply.

A selection from these charts is given for January and July in Figs 153 - 162 in which the direction of the mean vector wind for the month is shown by streamlines and the speed by lines of equal speed or 'isotachs'. It should be remembered that these charts give no indication of the variability of wind in both speed and direction which is usual from one day to another or even from hour to hour. The variability in the temperate zones has already been noticed; at all levels it is greater there than in other parts of the world. Because of the variability in direction, the mean vector speed is always less than the mean speed irrespective of direction; where the changes in direction are large, this difference may be considerable, but where the winds blow persistently from one direction, as for example in the subtropical jet streams, then the differences become almost negligible.

Since the temperature in the troposphere decreases on the average from the tropics towards the poles, the average thermal wind (Section 26) is westerly and for this reason the mean circulation in the upper air might be expected to consist of generally westerly winds circulating round centres of low pressure above the poles. In the southern hemisphere, the circulation outside the tropics appears to be mainly of this simple pattern and the prevailing westerlies of temperate latitudes increase with height with but little change in direction.

In the northern hemisphere the same tendencies are manifest; westerlies, increasing with height, prevail in the upper levels of the troposphere outside the tropics in spite of the large divergence from the zonal scheme of winds at the surface. The well-defined pressure systems such as the Aleutian and Icelandic lows give closed circulations only in the lower levels and become reduced in the upper levels merely to troughs in the circumpolar whirl. The Asiatic high of winter has entirely disappeared by 10 000 feet.

The general character of the variation of the westerly component with height in both troposphere and stratosphere is described in Sections 26 and 72. It was seen that in the northern hemisphere the greatest speed in the upper troposphere is attained at about 25–40° N in January and 40–45° N in July, while the mean speed in midwinter exceeds 100 knots in places. Exceptionally high speeds occur in the jet streams of temperate latitudes but these are transient phenomena and do not appear on the mean charts; on the other hand the subtropical jet stream forms an outstanding feature of these charts.

In high latitudes in winter the prevailing westerlies continue increasing with height above 50 000 or 60 000 feet (Fig. 157) in accordance with the temperature distribution, and westerlies there are prevalent up to at least 100 000 feet. At these high levels the winds increase as westerlies in a polar vortex around the winter pole but decrease in spring and become easterly during the summer.

Within the tropics conditions are more complicated and the geostrophic rule, upon which the theory of thermal winds is based, cannot be assumed. On the border of the subtropical high, the trade wind is a shallow current; above it the circumpolar westerlies are reached at a height of some 5000–10 000 feet. Nearer the equator the easterly flow may extend to considerable heights, perhaps into the stratosphere, although westerly winds are found in certain conditions even at low levels. On the whole, however, there is a solid easterly current which extends over some 10–20° of latitude and moves north and south with the sun in step with the movement of the low-pressure belt at the surface; the greatest speed of these easterlies in summer is found near the tropopause in latitude 15°N where it averages as much as 70 knots in places. Monsoonal effects modify the easterly flow at low levels in certain longitudes. Thus the south-west monsoon of India extends as a westerly

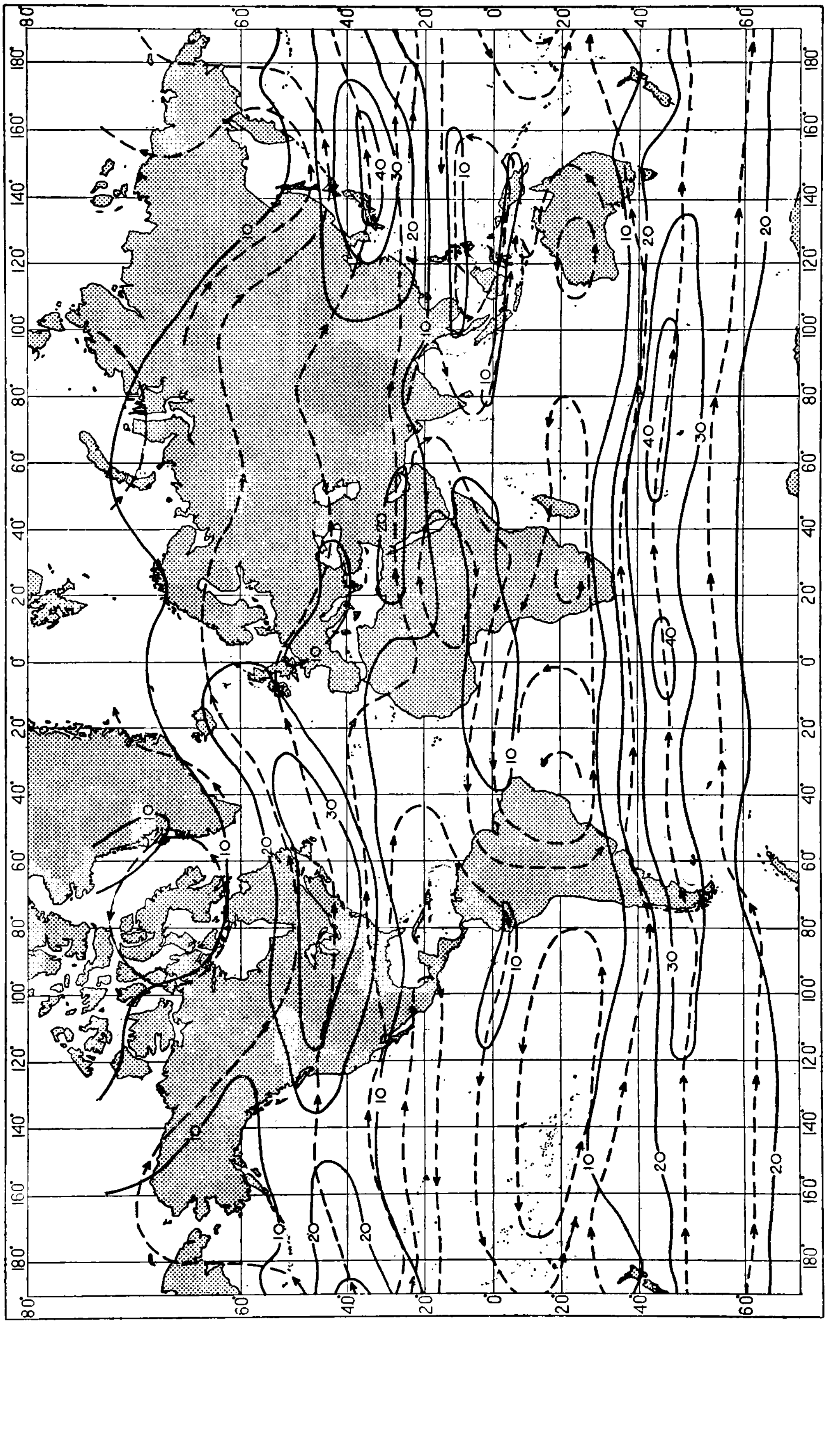


FIG. 153. *Streamlines and isotachs (knots) of the vector wind distribution. January - 700 millibars*

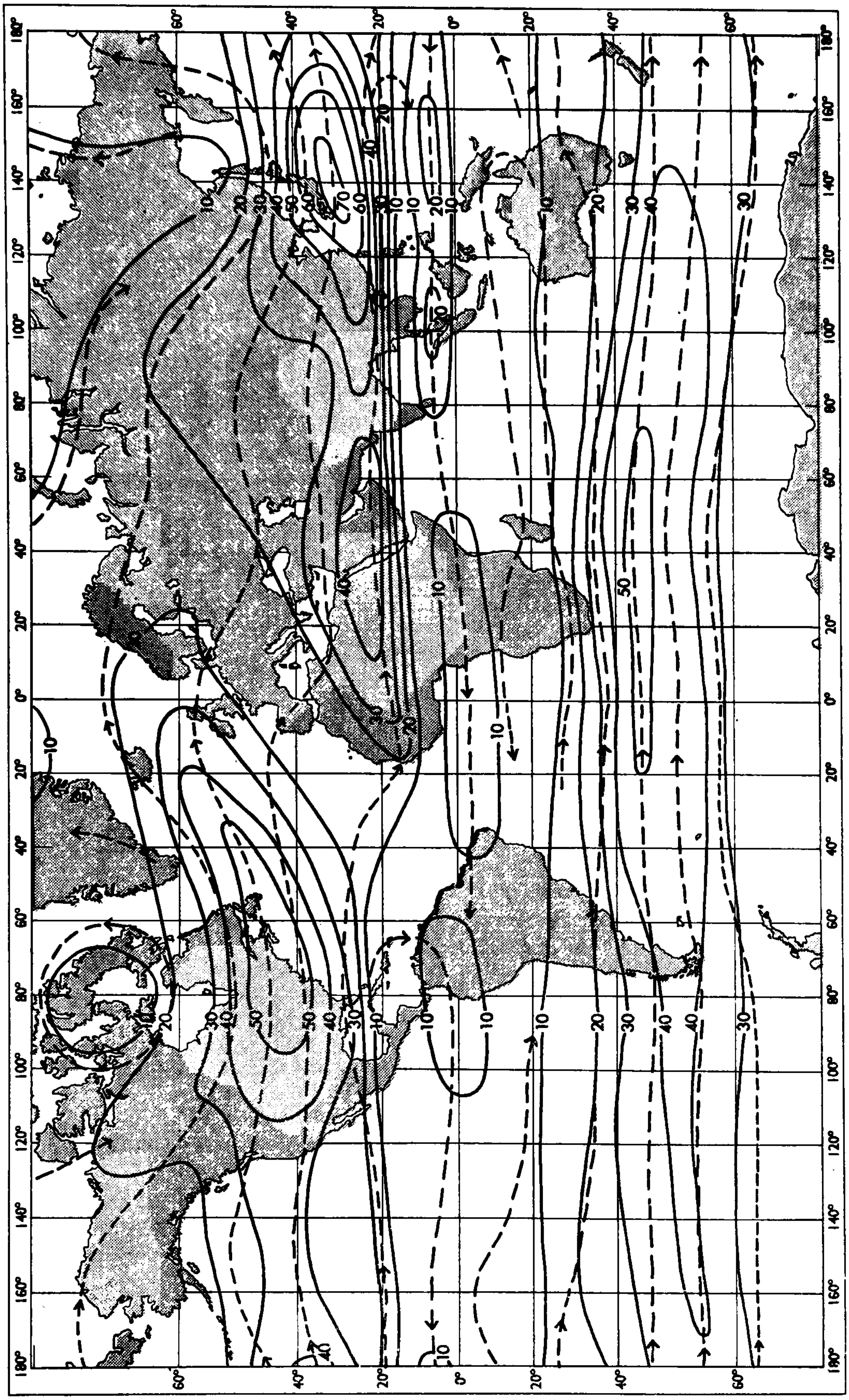


FIG. 154. *Streamlines and isotachs (knots) of the vector wind distribution. January – 500 millibars.*

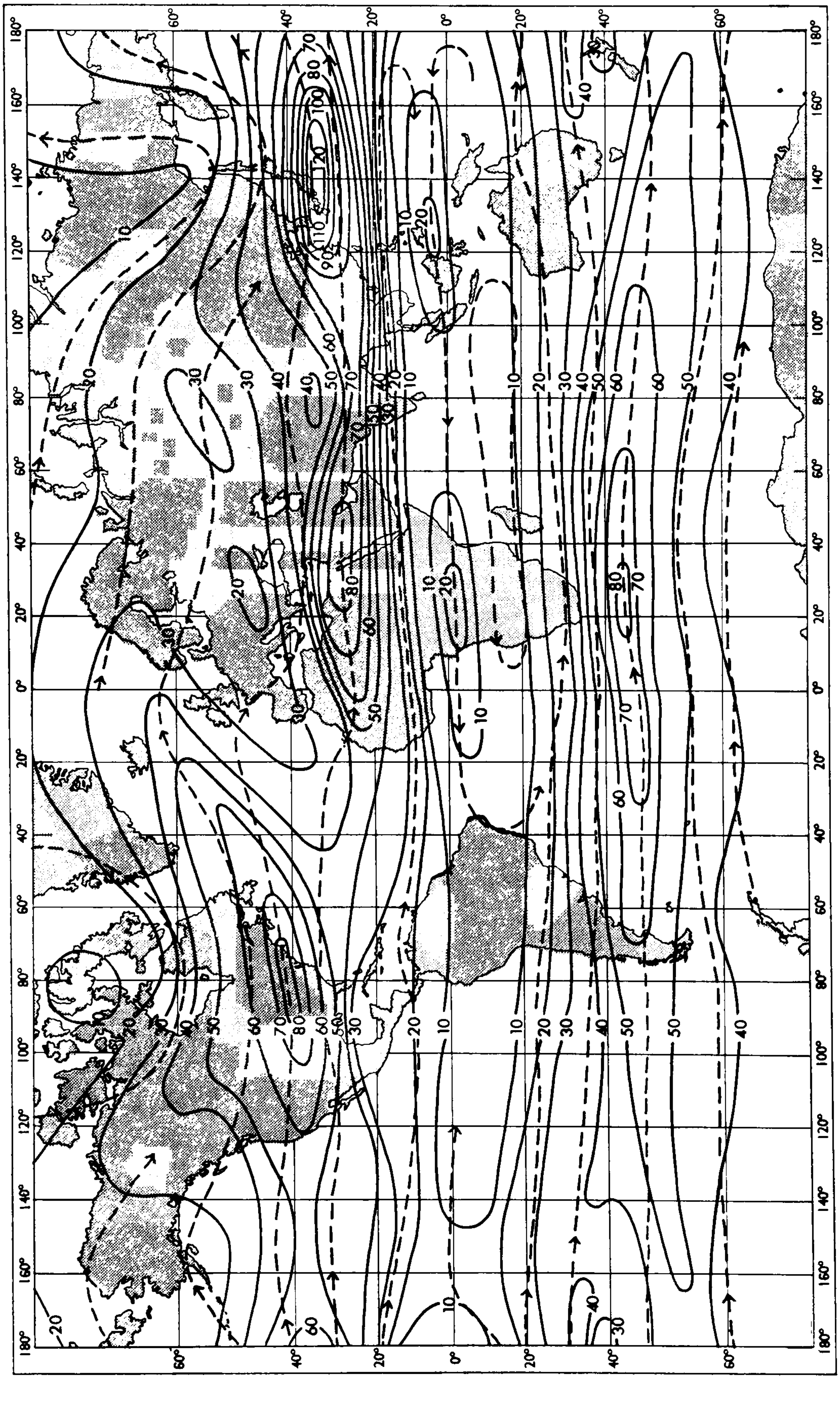


FIG. 155. *Streamlines and isotachs (knots) of the vector wind distribution. January – 300 millibars*

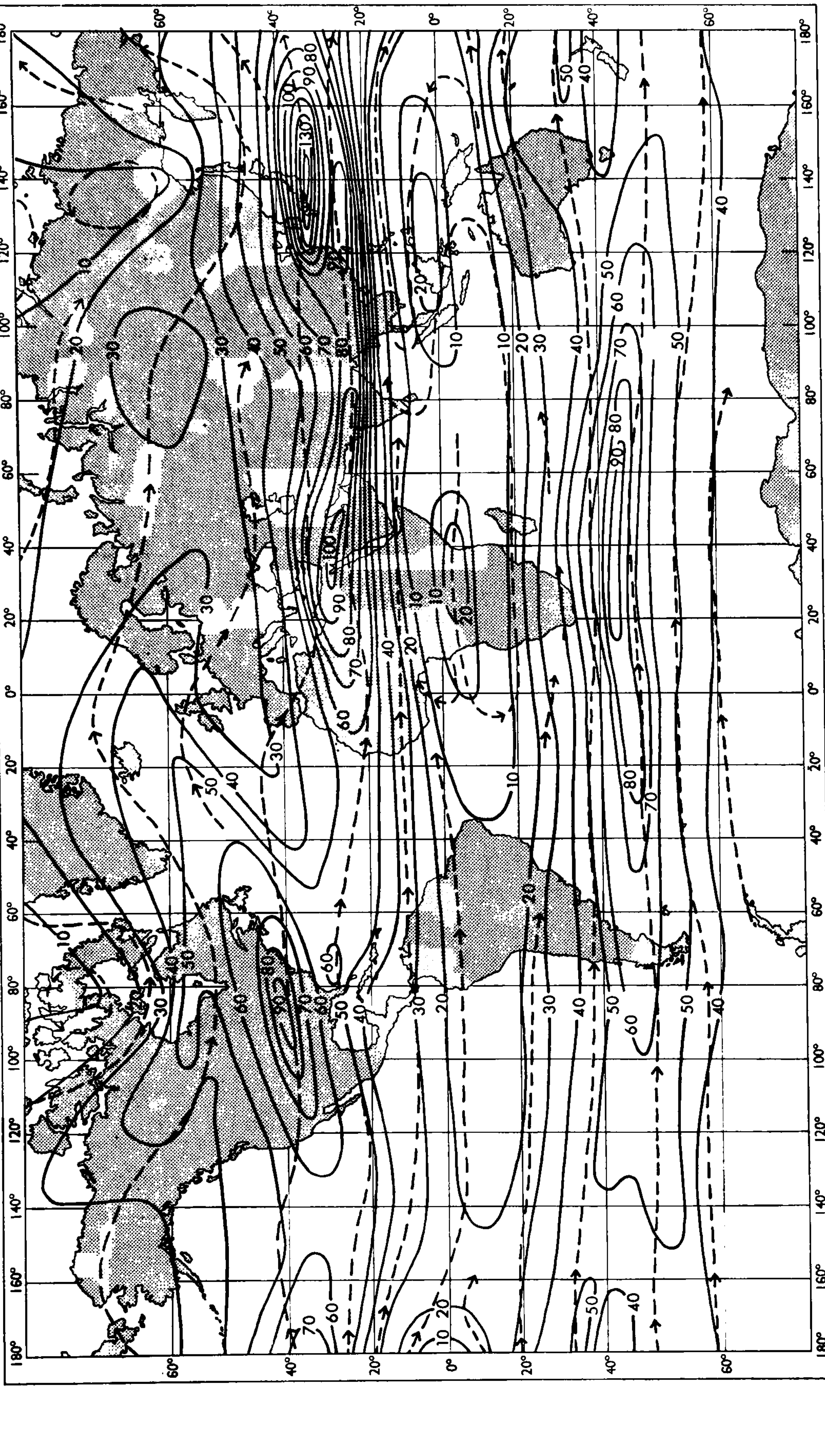


FIG. 156. *Streamlines and isotachs (knots) of the vector wind distribution. January – 200 millibars*

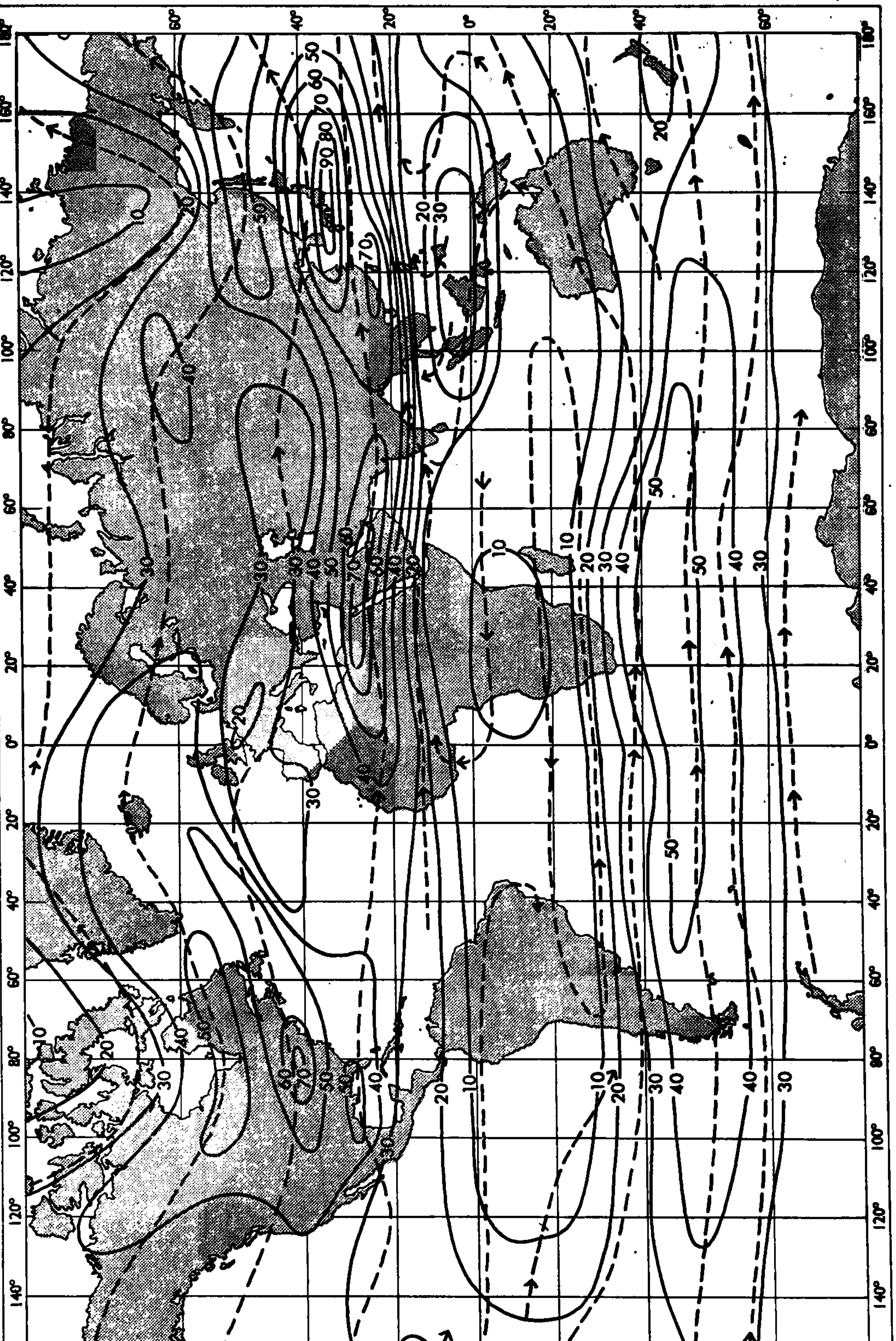


FIG. 157. *Streamlines and isotachs (knots) of the vector wind distribution. January – 100 millibars*

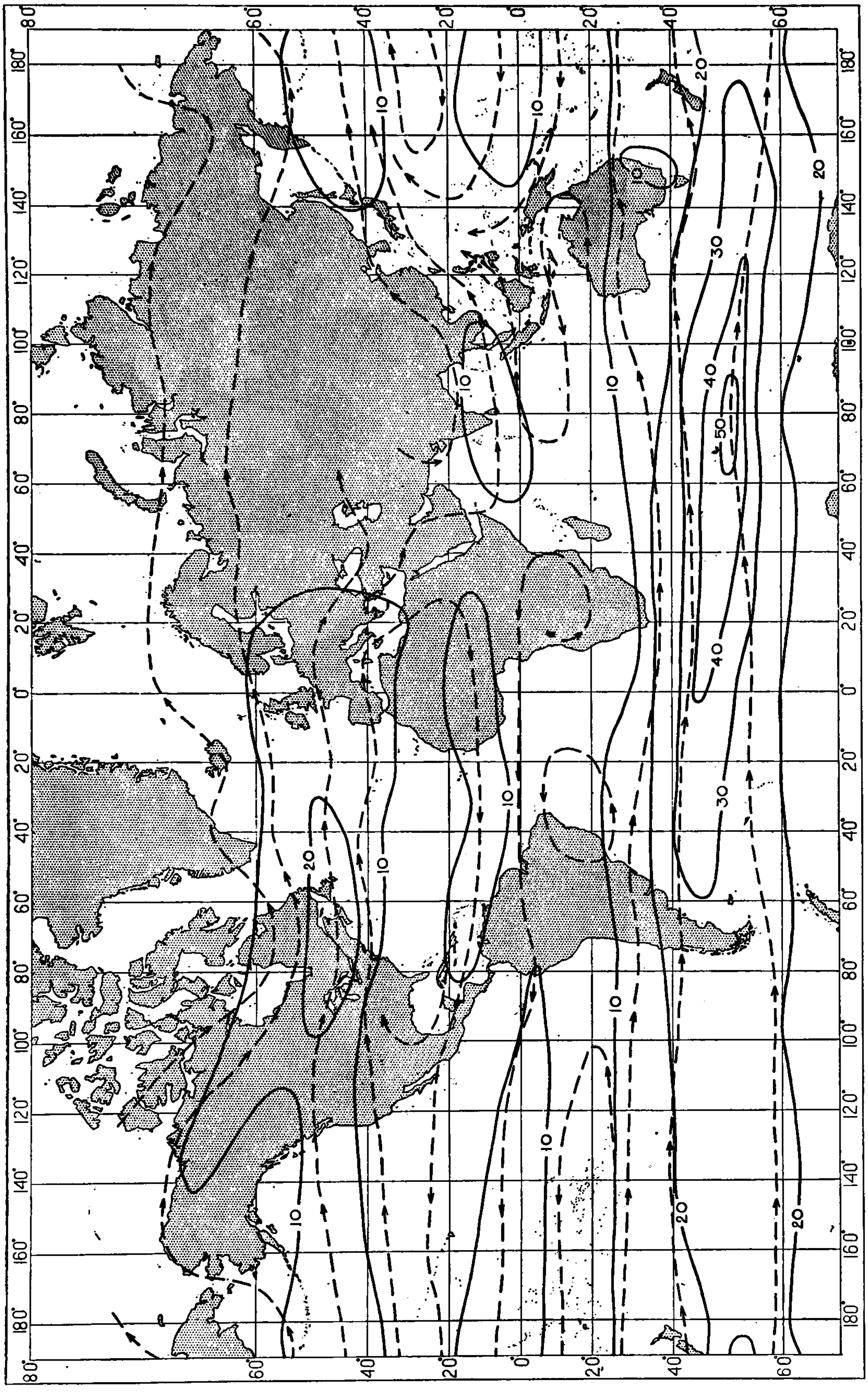


FIG. 158. *Streamlines and isotachs (knots) of the vector wind distribution. July – 700 millibars*

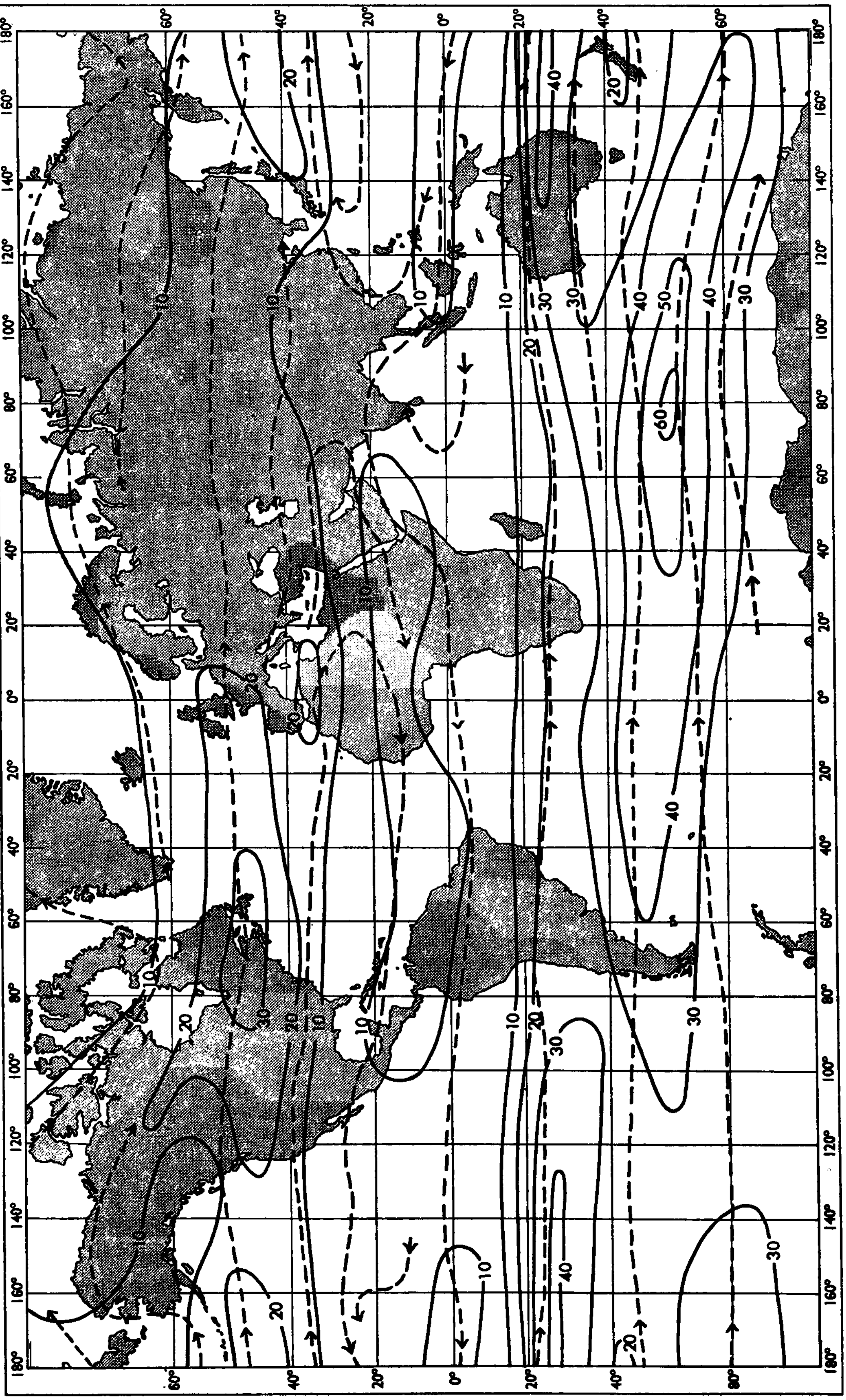


FIG. 159. *Streamlines and isotachs (knots) of the vector wind distribution. July - 500 millibars*

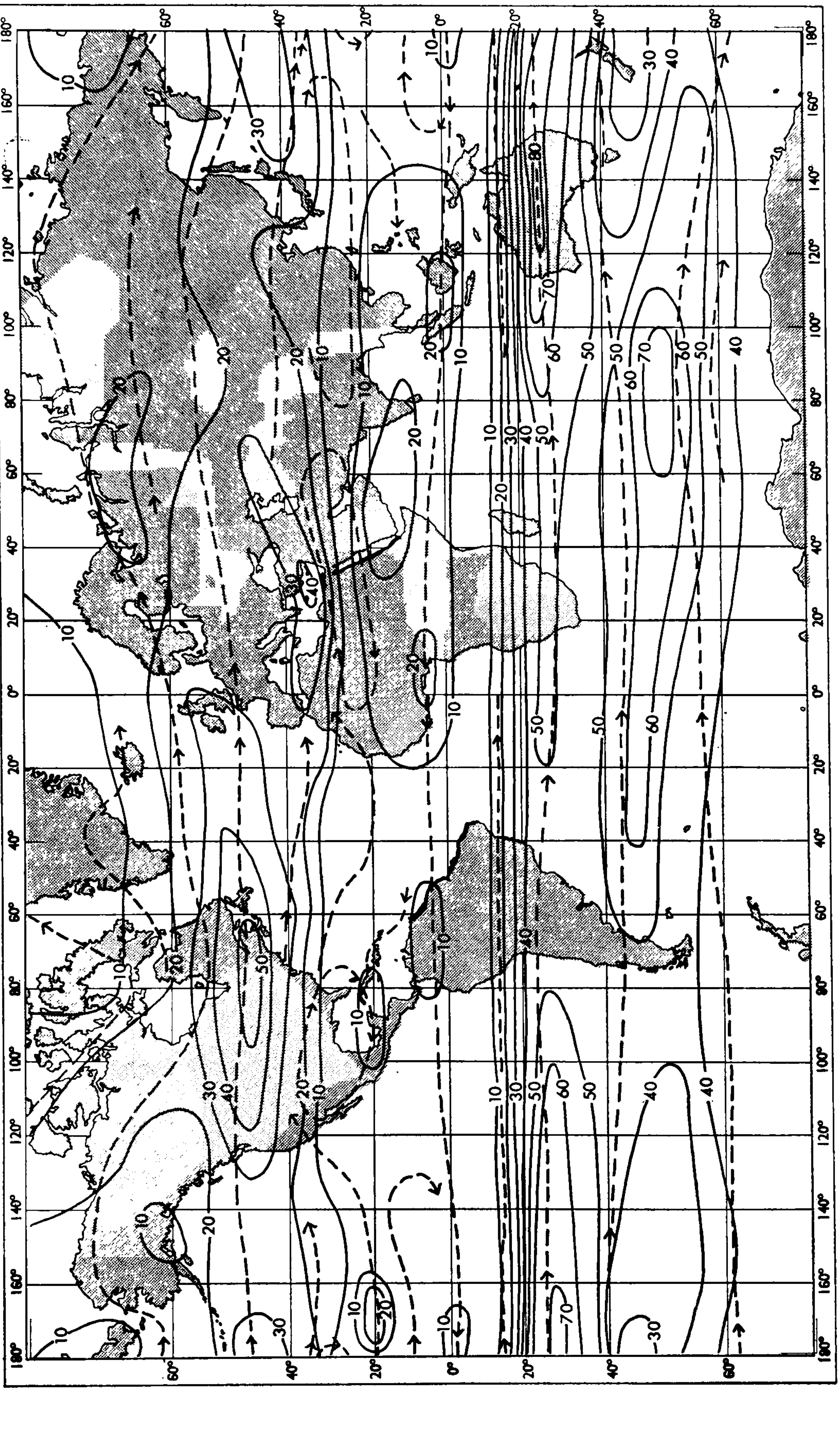


Fig. 160. *Streamlines and isotachs (knots) of the vector wind distribution. July - 300 millibars*

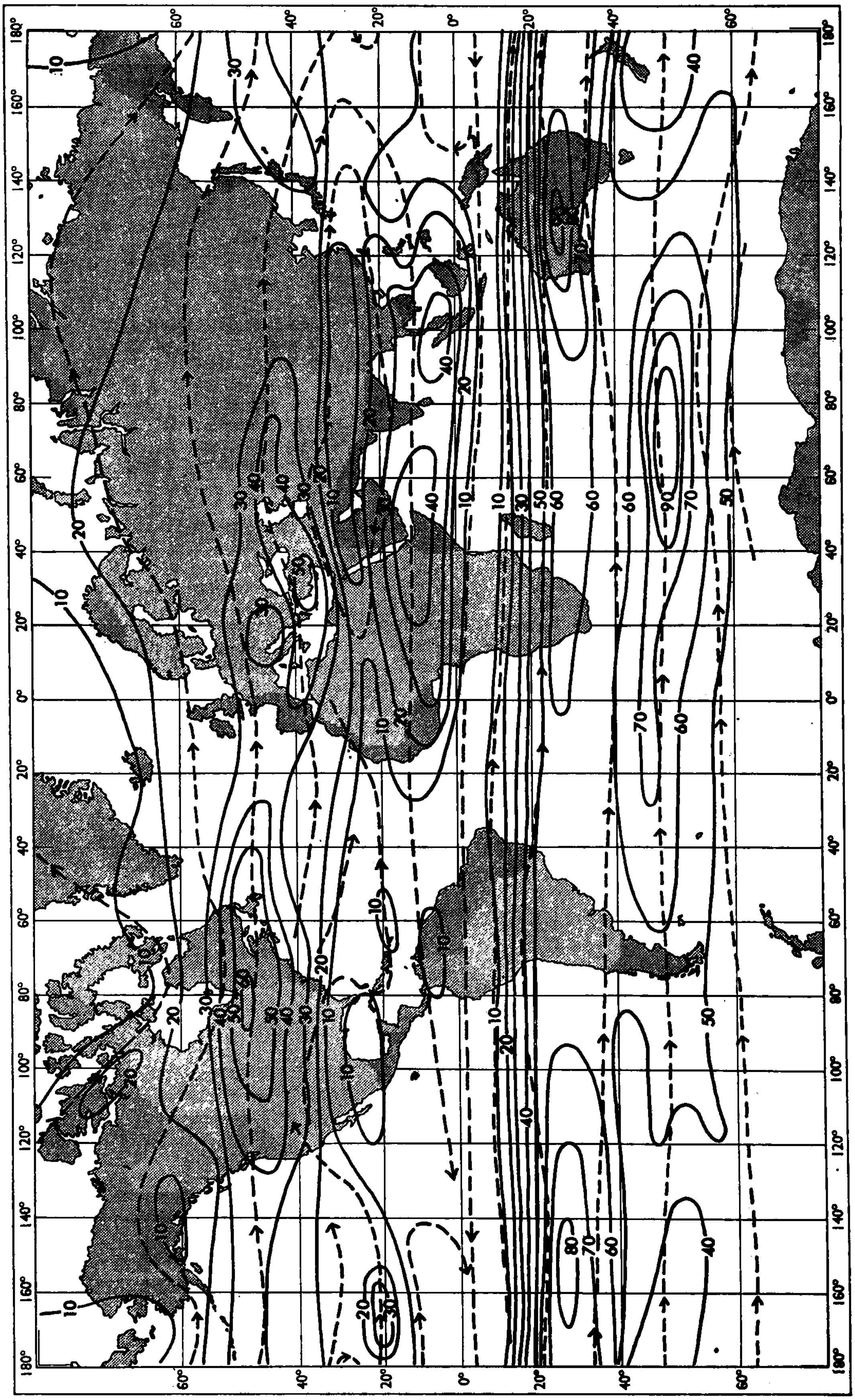


FIG. 161. *Streamlines and isotachs (knots) of the vector wind distribution. July – 200 millibars*

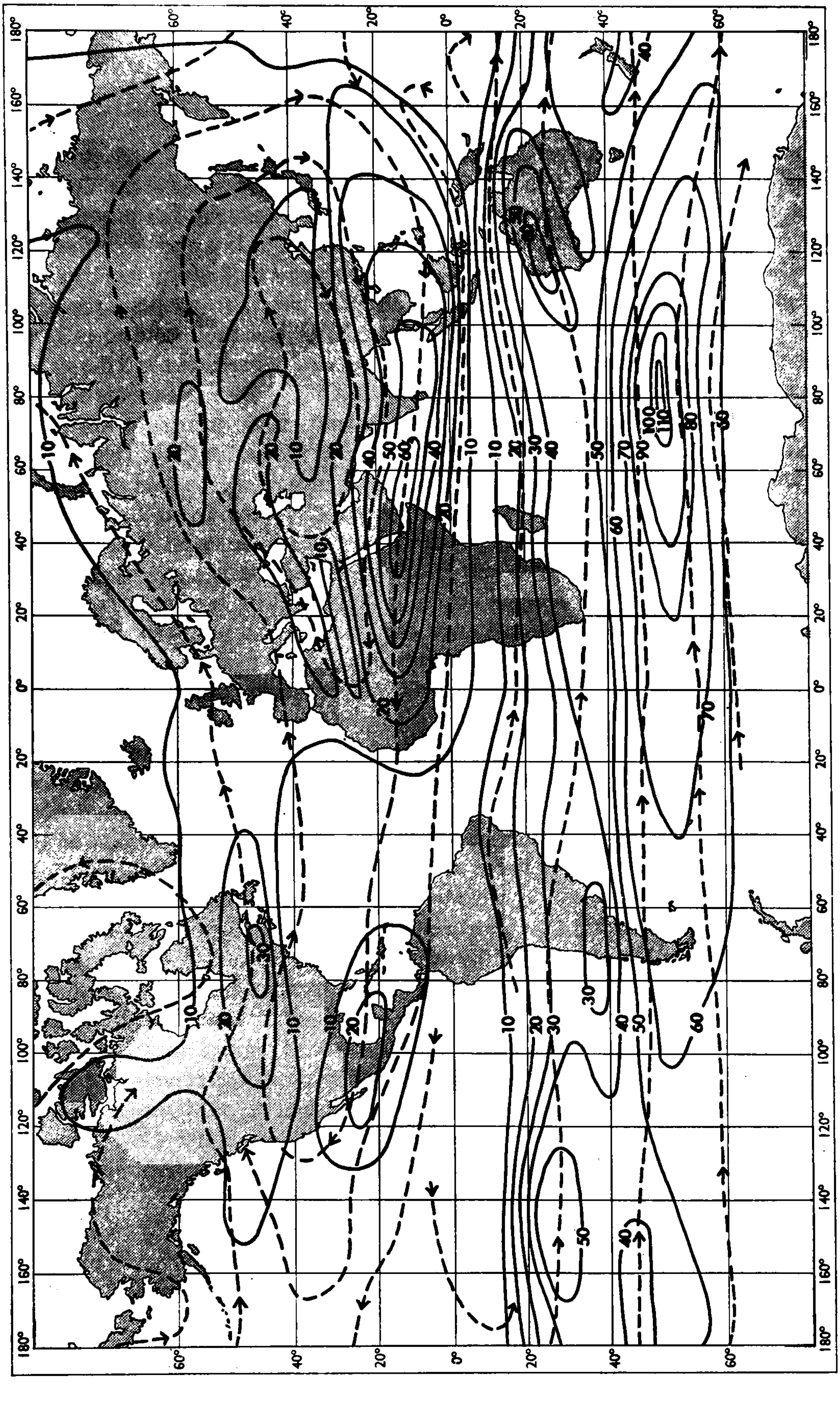


FIG. 162. *Streamlines and isotachs (knots) of the vector wind distribution. July – 100 millibars*

current to about 15 000 feet or more before being replaced by easterlies. The south-west monsoon of west Africa is a shallow current, the easterlies being firmly established by 10 000 feet.

161. CLOUDINESS

The total amount of cloud is of interest in relation to visual navigation, astro-navigation and to the risk of ice accretion, although in each of these cases the distribution of cloud with height is also relevant.

On the whole the cloud cover is least in the subtropical belts of high pressure, especially over the continents, and is often negligible over the desert areas. Over the oceans in these zones the average cover is about 4 oktas (eighths). In the equatorial belts of low pressure the cloudiness averages around 5 oktas but over land, because of diurnal variation, the value is normally greater than this by day and less by night. In the temperate belts the average cover is about 6 oktas. In all belts there are variations depending mainly on orographic effects. As has been noticed in other respects, the Asiatic monsoons bring about pronounced departures from the simple zonal distribution. Cloud cover over much of India and Burma exceeds 6 oktas during the south-west monsoon but skies are practically clear during the north-east monsoon, in which season the cover is only 2 oktas or less in the Gobi Desert also. In the temperate zone of the northern hemisphere in winter, the average reaches 6 oktas or more in a belt which extends eastwards from the Great Lakes to the British Isles and then broadens to embrace most of north and west Europe as far as the Ural Mountains. In the north-west Pacific there is similarly a very cloudy area, but shelter by the Rocky Mountains results in lower values over much of North America. In summer, cloud amounts over the oceans remain almost as great as in winter in spite of reduced cyclonic activity; over land there is generally some decrease although here again the effects of diurnal variation should be considered. In high latitudes the cloudiness increases generally and mostly exceeds 6 oktas north of the Arctic Circle except over Greenland where it is reduced to about 4 oktas.

In regard to the effect of cloud on landing conditions, the frequency of cloud base below 1000 feet above ground gives a useful guide. Although cloud occasionally falls below 1000 feet over the sea or low ground during periods of bad weather, aerodromes particularly liable to very low cloud are usually those subject to orographic effects. These effects have been discussed in Chapter 6; often cloud is formed on the surface, giving fog at the aerodrome or at a height of a few hundred feet. Much depends on the situation of the aerodrome relative to the local topography and to the direction of approach of moist air currents; thus elevated coastal aerodromes with on-shore winds are particularly liable to be 'closed' by low cloud. It would however serve no useful purpose to attempt to generalize on this point. Where particular information is required on the frequency of occurrence of height of base of low cloud, this can be obtained for established aerodromes through the appropriate meteorological service or through the International Civil Aviation Organization. The same holds also regarding frequencies of low visibility. Some figures are available in *Meteorology of airfields* by C. S. Durst.

162. PRECIPITATION

The distribution of average annual rainfall is given (very much generalized) in Fig. 163; the seasonal variations are discussed for certain areas in Chapter 23.

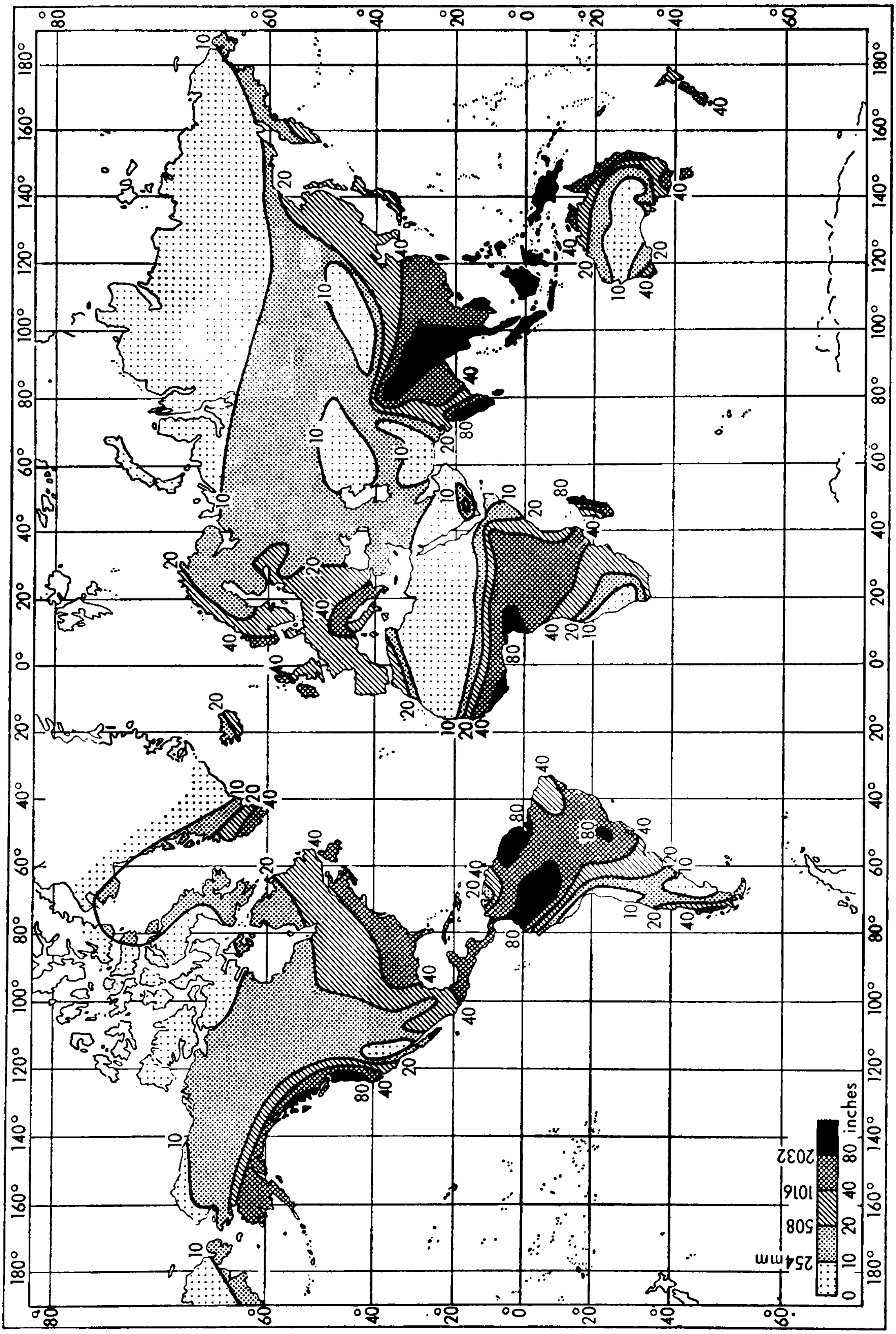


FIG. 163. *Mean annual precipitation – very much generalized*

The rainiest regions are found in the equatorial belt and in south-east Asia while the arid regions (rainfall less than 250 millimetres) are mainly in subtropical areas and at high latitudes. In the subtropics the dryness is caused mainly by subsidence in the subtropical highs, and over the continents it results in the true desert regions. At high latitudes the small amounts of precipitation is a result partly of subsidence in high pressure and partly of low temperature; snow or rain occur at times but the cold air has little capacity for moisture and the amounts of precipitation are small.

In broad terms, the rainfall of low latitudes is of a convective and orographic nature; in temperate and polar latitudes cyclonic rainfall, again modified by orography, predominates. There are however exceptions; tropical cyclones are responsible for heavy local rain in certain areas, while convective rain is common over the temperate land masses in summer. Snow is the normal form of precipitation where the average surface temperature is below 0°C and it occurs at high levels even in the tropics; it will often be encountered in flight in localities where it is quite unknown at the ground. The distribution of hail is much the same as that of thunderstorms (see below) except that it is rare near sea level in the tropics. The most severe hailstorms are reported from the plateaux of South Africa and from the interior of the United States in summer.

Some important effects of orographic features upon rainfall distribution may be noted. Most significant is the difference between the windward and leeward sides of the land masses, particularly when these rise to a considerable altitude. The windward side with respect to the prevailing rain-bearing wind normally receives much the greater amount of precipitation; in extreme cases the leeward side may be arid. A good example is to be found in Australia where the summer monsoon rains affect only the east and north leaving arid conditions in the south-west. Similar desert regions are found in South Africa (the Kalahari), northern Chile (the Atacama) and the extreme south-west of the United States. The relative coldness of the neighbouring ocean current on these western coastal regions is a factor contributing to the aridity.

In temperate latitudes, where the rain is mostly cyclonic with prevailing westerly winds, the west coasts and mountain slopes are the cloudiest and receive the greatest rainfall. Illustrations of these tendencies are to be found in New Zealand and in temperate South America. Again, the Pacific coast ranges of North America confine the influence of the moist oceanic air to a narrow coastal strip of Canada and the northern United States, where in winter the rainfall is very heavy. Although depressions may cross the mountains their rain-producing capacity is largely removed in the process, unless eventually they draw in further humid air from another source such as the Gulf of Mexico or the Atlantic. The absence of corresponding mountains in Europe affords a striking contrast, for moisture of Atlantic origin may be precipitated far over central Europe and Russia. Another illustrations is found in Asia where the summer monsoon rains are exhausted over the mountains of India and south-east Asia, leaving large areas of the interior quite arid.

163. THUNDERSTORMS

The parts of the world most affected by thunderstorms are the tropical rain areas. In several of these areas the average frequency approaches or exceeds 30 per cent of days (110 days per year), namely southern Mexico, Panama, central Brazil, central Africa, Madagascar and Java. The last, with a frequency of about 220 days, is probably the most thundery region of the world. Thunderstorms too are generally much more frequent over the land than over the sea. Within the subtropical belts of high pressure there are several localities where thunder is rare although, except for the deserts of north Africa and Arabia, the areas lie almost entirely over the sea and western coasts of the adjacent continents. Thus the thunder-free coasts include those of northern California, Chile, northwest and south-west Africa, in fact just those localities where sea fog is common. Since sea fog results from surface cooling and thunderstorms mostly from surface heating, this is not surprising. The regions with less than 1 per cent of days with thunder (4 days per year) include the whole of the Arctic and Antarctic where, indeed, thunder is almost or entirely unknown.

There is often a very marked seasonal variation in thunderstorm frequency. In tropical areas this is associated with the seasonal variations in the positions of the main rain areas. In temperate latitudes surface heating in summer gives rise to frequent storms over the central and eastern United States, and to a lesser degree, over Europe.

164. FOG

Fog – as distinct from cloud on high ground, which is common in all but the driest parts of the world – is unusual in the tropics except in certain localities where fog or mist forms frequently in low-lying parts at night but clears quickly after sunrise. Within the trade-wind zones it occurs only as sea fog near the western coasts of the continents where the sea surface temperature is low (caused largely by the up-welling of ocean waters). Such coasts are found in south-west Africa, Chile, Morocco and California. It is interesting to note that in all these areas the hinterland is arid and the moisture, which is confined to the surface levels (with a temperature inversion above), is inadequate to provide more than cloud or perhaps drizzle.

Fog is predominantly a feature of cool temperate and high latitudes. Over the land it is common in autumn and winter in quiet conditions whenever the air is of maritime origin. It is also particularly prevalent in summer in north polar latitudes whenever milder air arrives from the south. Over the sea maritime air of subtropical origin is likely to bring fog at any season of the year, although spring and early summer are the most likely seasons. There are certain notoriously foggy areas where the sea surface temperature is very low for the latitude. The Newfoundland Banks, affected by the cold Labrador ocean current, and the corresponding coast of Asia affected by the cold Kamchatka current are the most important.

165. DUSTSTORMS

In all the arid regions of the continents duststorms and dust haze are the main hazards to aviation. The areas with annual rainfall less than 250 millimetres (see Fig. 163), with the exception of the tundra and polar regions, may be taken as broadly defining the areas subject to dust. The most important are central and south-west Asia, north Africa (the Sahara and the Sudan), south-west Africa (the Kalahari desert), South America (northern Chile and central and south-eastern parts of the Argentine), the central and south-west United States, and the interior and west of Australia.

166. GALES

Persistent wind of gale force is associated with both tropical and extratropical cyclones. Tropical cyclones (see Section 159) are exceptional phenomena which affect only certain areas of the tropical oceans and the neighbouring coasts. Extratropical cyclones or depressions are a normal feature of temperate latitudes and gales are in consequence common in the zones of prevailing westerlies in both hemispheres, being strongest over the sea and in the winter season.

Apart from large-scale cyclonic disturbances, gales may occur as local phenomena of orographic origin, as temporary squalls of a line-squall or thunderstorm character, or in association with tornadoes (Section 119). They occur too widely to permit of brief summarizing.

CHAPTER 23

AVIATION CLIMATOLOGY OF AIR ROUTES

167. INTRODUCTION

Planning, whether for a particular flight or for the inauguration of regular operations, usually requires considerable advance preparation necessitating the fullest information concerning meteorological conditions to be expected on the route. Knowledge is required not only of average conditions but also of the extremes, or more particularly of the frequency with which conditions exceed (or fall short of) certain specified limits. This information is essential in order that items such as fuel-load and pay-load may be assessed in relation to the characteristics of the aircraft to be used. On some routes the extremes of adverse conditions such as headwinds, or fog at terminal airfields, are at times too severe to allow of a flight then being made; it is an advantage, therefore, to know beforehand the frequency and likely duration of conditions making a flight impracticable. If the frequency is appreciable, an airline operator may decide to plan for an expected regularity of service of, say, 85 per cent, meaning that cancellations are not expected to exceed 15 per cent of all occasions over some particular period. Hence detailed information of a climatological nature is required in order that the economic potentialities of a regular service can be estimated.

It is desirable also that those immediately concerned with the flights should have a general knowledge of conditions prevailing on the route. A prior study of the climatology of the area traversed by the route prepares the pilot and navigator for the general nature of the weather conditions to be expected according to the time of year and provides a background against which the route forecast can be the better appreciated both before and during flight.

In a full account of the aviation climatology of an air route, attention should be given to each of the following items in each month or season described. To facilitate description, it may be convenient to divide a route into two or more sectors.

(i) The general meteorological characteristics of the route, including the prevailing distribution of pressure and wind, the climatic type, the positions of main frontal zones, etc.

(ii) The average direction and speed of the wind at points on the route, both at the surface and at selected upper levels; frequency of gales at the surface; variability of wind; equivalent headwinds.

(iii) Visibility, in particular the frequency of bad visibility due to fog, dust, etc.

(iv) The variation of temperature at the surface and in the upper air; the height of the 0°C isotherm and of the tropopause.

(v) The average cloudiness at points on the route; the diurnal variation and vertical extent of clouds.

(vi) The frequency and intensity of precipitation in its various forms, including thunderstorms.

(vii) The frequency and severity of conditions favourable to airframe icing.

(viii) Conditions at terminal aerodromes, including alternates.

(ix) Any other meteorological conditions of importance to aircraft on the route.

Detailed climatological information for all air routes with notes on terminal conditions would require more space than is available in a book of this kind. The average conditions along any route in January and July can be derived from the charts and text of the previous chapter, but even in a brief account such information requires supplementing by some indication of the variability of the different elements and of the extremes attained, of the day-to-day synoptic variations and also of the topographical and other local peculiarities affecting the routes. The routes crossing the North Atlantic are deliberately treated at some length by way of an example. The treatment of subsequent routes is briefer, but further information can be obtained from Chapter 22 and from sources listed in Appendix III and possibly the reader will draw on his own experience of flying on air routes.

168. NORTH ATLANTIC TRUNK ROUTES

The regular routes across the North Atlantic cover an area which extends in the north to Iceland and southern Greenland and in the south to the Azores and Bermuda. The climatic zones most often affecting this area are therefore those of the disturbed westerlies and of the subtropical or Azores anticyclone.

Winter

General features. The outstanding feature of the weather of the northern routes is the prevalence of cyclonic activity. The region affected in this way extends further south on the western than on the eastern side of the ocean. This is because the land mass of North America, most of which is cold in winter, allows cold air from Canada to penetrate to quite low latitudes without much modification before being brought into contact with warm moist air from the subtropical seas. Often this contact takes place between Florida and Bermuda. A sharp front forms in this area, depressions originate there, develop and move away north-eastwards; meanwhile the lows occlude and the centres drift to the north of the polar front, the mean position of which in midwinter extends from Florida towards south-west England. While these frontal lows on the whole move towards the north-east, individual paths are often erratic although movement towards a westerly point is uncommon. The rate of movement is often about 30 knots but may have any value up to about 80 knots.

In addition to the lows which form on the polar front, other low-pressure systems develop in the polar air masses over the sea to the north and still others form along secondary fronts over North America. Since many of the tracks of the deeper depressions pass through or near the area between Iceland and southern Greenland, this is an area of low mean pressure – the so-called Icelandic low; this however is not a constant feature, and on synoptic charts the centre or centres of low pressure will often be found elsewhere. Southwards the mean pressure increases and reaches a maximum in the subtropical high at about 30° N. This high is itself variable in position and intensity and is at times displaced by cyclonic developments. Occasionally it links up with the winter highs over North America, or across Europe with the Asiatic high. Although depressions with their stormy weather predominate, anticyclones and ridges of high pressure occasionally bring dry quiet weather to parts of the area. Generally, however, weather is very changeable, both at any one place and along the track of any particular flight. The type of situation illustrated in Fig. 107 is typical of many days in winter and the same general features – highs, lows, fronts and extensive air masses – appear on nearly all synoptic charts of this region. The details change from day to day, and in very disturbed situations frequent changes of wind and weather are encountered by an aircraft in flight or at a terminal aerodrome.

Wind. The average distribution of wind both at the surface and at higher levels may be seen from the charts of Chapter 22 but on any particular occasion the winds differ widely from the mean. The more northerly routes often pass to the north of the low-pressure centres where winds are easterly, at least in the lower levels. At the surface gales are often widespread while rapid changes of wind occur at fronts. With increasing altitude the winds become modified, the general effect being an increase of the westerly component throughout the troposphere. From the charts it will be seen that the mean vector wind is roughly westerly or north-westerly over the British Isles, becoming west-south-westerly over the ocean north of 40° N in association with an upper trough over eastern North America. There are however minor seasonal variations. In jet streams, which are quite frequent in the upper half of the troposphere over this area, the wind velocity commonly exceeds 100 knots, and even 200 knots on occasions; jet streams with an easterly component are comparatively rare. Although much variation of wind is normally encountered on any one transatlantic crossing, the equivalent headwind on a westbound flight (or tailwind when eastbound) is usually considerable. For example, on the great-circle route from London to New York in winter at 30 000 feet, the mean equivalent headwind for the whole route averages about 50 knots and may reach to about 100 knots. Still higher extreme values may be attained on individual sectors, for example from Great Britain to Iceland, on which it is possible for a jet stream to lie along the whole track.

Because of the great variability of wind and weather, the most appropriate track and flying height must be chosen on the basis of the forecast conditions on each occasion. In extreme cases, strong headwinds may necessitate a postponement of the flight in slow aircraft.

Strong winds near the surface are very turbulent in the vicinity of high ground, but generally over the open sea turbulence is marked only near fronts and in the cold air to the rear of a depression, especially when the air has newly arrived from high latitudes or from the cold continent of North America. The turbulence may be severe in convective clouds at cold fronts and in unstable polar air masses. Clear-air turbulence at high levels should be anticipated in the neighbourhood of jet streams (Section 69).

Cloud and precipitation. The total cloud cover averages about 6 oktas over most of the area. The frontal cloud systems are usually extensive both vertically and horizontally; in these the base of the lowest cloud may fall almost to the surface and it is often impossible to avoid the clouds while in flight at levels below about 20 000 feet or at times even 30 000 feet. In contrast, the convective clouds of the polar air masses are commonly isolated although occasionally they become more or less continuous, as when convergence is taking place over a wide area and a low is developing in the polar air. Widespread sheets of stratocumulus with tops not above about 4000 feet are common in the tropical air masses and occasionally in polar maritime air. Anticylones are usually devoid of cloud in the middle and upper levels but over the sea, as over the land in winter, they are subject to low stratocumulus, the formation of which is favoured by the subsidence inversion which is usually present.

Precipitation similarly is often widespread and continuous near the warm fronts but usually of a showery type elsewhere. To the north and north-west of the North Atlantic there is often snow right down to the surface; elsewhere frontal precipitation, which starts in the upper levels as snow and as such may be encountered in flight, melts before reaching the surface.

Ice accretion. Since the mean height of the 0°C isotherm is low (Fig. 145), even down to the surface in the north and west of the area, it follows that a great depth and extent of frontal cloud systems are included within the temperature range favourable for airframe icing which is therefore likely to be severe at times. The convective clouds are also favourable for icing, the more so when they are massed together. Rain ice in precipitation below warm fronts and occlusions is unusual except when the cold air is of recent continental origin.

Visibility. Over the sea, fog and mist form as a result of advection, usually in a mass of tropical air as it becomes cooled and stabilized on moving into higher latitudes; one example is provided by the warm sector of a depression. Fog is accordingly often widespread in such situations but if the wind speed is more than moderate, very low stratus is more likely than fog although visibility may still be reduced by drizzle. Frontal precipitation too is often the cause of reduced visibility over a wide area. In fresh polar air, the visibility is good except in showers but it may become less good subsequently, especially if the air mass returns northwards and so becomes subject to surface cooling.

Advection fog is notably frequent, even in winter, off Newfoundland where air from the warm Gulf Stream regions passes over the cold waters of the Labrador Current which are maintained at a low temperature by melting ice. The neighbouring coastal regions, cold in winter, are also liable to this type of fog when the wind is on-shore. Further, the seaboards on both sides of the Atlantic are subject to rather frequent radiation fog in winter in association with high-pressure systems and this may be aggravated by smoke from industrial areas.

Above the surface levels, visibility is generally good except in cloud but smoke haze may be encountered over and near the continents.

Summer

In this season the mean pressure chart (Fig. 148) shows that instead of the pronounced Icelandic low of winter, there is merely a weak low centred over north-east Canada; also the subtropical or Azores high has intensified and its centre moved some 5° further north; moreover, the winter continental highs have disappeared, The mean position of the polar front is further north and in July extends from Newfoundland to the north of Scotland. However, the change of conditions from the winter is one of degree rather than of kind and for that reason the summer régime need not be described at the same length as that of winter.Thus in summer the lows are on the whole less frequent and their tracks mostly lie further north than in winter, while anticyclonic conditions prevail for longer spells. Even so, conditions vary much from year to year, some summers being marked by an almost unbroken succession of lows. Generally the lows of summer are less vigorous than those of winter, gales are less frequent, upper winds weaker and prolonged stormy conditions unusual. The contrast of temperature between polar and tropical air streams are reduced and the fronts are less marked in consequence. Also the polar air is less cold relative to the sea surface, and the associated convective cloud, showers and squalls are less frequent than in winter. On the other hand, much stratocumulus cloud forms in the south-westerly air streams and total cloud amount averages about the same as in winter. The level of the 0°C isotherm is, of course, higher than in winter but ice accretion, although less frequent, remains an important problem.

In respect of visibility over the oceans, the summer conditions are worse than those in winter. Spring and early summer are the seasons when advection fog is most frequent. It forms in the south-westerly tropical air and often becomes widespread over the ocean. The western approaches to the British Isles are affected and also the coasts where the wind is on-shore. The same seasons also produce the maximum frequency of fog off Newfoundland.

Terminal conditions

Under this heading are included brief notes on weather at terminal airports. Almost any airport is likely to become unusable at times; the most frequent cause is fog or very low cloud, while other conditions such as heavy snow and gales may also be prohibitive. The limiting conditions for the safe operation of aircraft depend on many factors including the type of aircraft and the landing aids available, but all that can be done here is to indicate in very broad terms the nature and frequency of the meteorological conditions likely to make an airport unusable.

London/Heathrow Airport. This terminal is subject to a rather high frequency of fog, mainly of radiation type but aggravated by smoke when the wind is from between north-east and east-south-east. Fog is most frequent from October to March and can be persistent; November is the worst month, the duration of fog then averaging 20 per cent of the total time, though the steady decrease in atmospheric pollution as the Clean Air Act takes effect must cause a gradual reduction in this figure. Very low cloud occurs at times during periods of bad weather in any season but especially in winter and in association with fronts approaching from east or south.

Prestwick Airport (Scotland). On account of the sheltered situation, fog and very low cloud are rare, although with moist air approaching from west-south-west the cloud base may be near the surface.

Keflavík (Iceland). The airfield is exposed to the sea from east-south-east through south to north-west, and in all months the weather is for the most part controlled by the movement of Atlantic lows. Actual fog is rare but low stratus from the sea is frequent, with the base often below flying minima. The principal cause of low visibility at the surface is precipitation – fine rain or drizzle which accompanies the stratus in summer, and snow in winter. Drifting snow is also a hazard. Gales occur throughout the year, being most frequent in winter when they occur on an average of 6 or 7 days per month.

Narsarssuak (Greenland). This airfield is located 50 miles from the south-west coast at the head of a steep-sided fiord, so that winds are almost invariably from south-west or north-east. North-easterly flow is most typical of winter when lows pass to the south of Greenland; it is then of föhn type and although the weather is clear, the wind often reaches gale force locally. A north-easterly flow also occurs when pressure is high over the ice cap, a condition which is common in spring; the wind may then be maintained at about 20 knots for some hours, so delaying take-offs which can be made only towards south-west, that is, downwind in these conditions. Summer conditions are characterized by stable air masses, and any flow from a westerly or southerly direction produces the sea stratus so prevalent over these waters. In this way the mouth (at least) of the fiord is filled with cloud, so blocking the normal approach to the station which on this account becomes closed much earlier than one with a more open location. Similarly any other deteriorations in the fiord, such as heavy frontal showers, may cause the terminal to be closed.

Goose (*Labrador*). Fog and very low cloud bases are rare except during snow conditions in winter; these are usually associated with easterly or north-easterly flow occurring either with the eastward passage of a low on a track lying south of the airport, or with high pressure to the north-west. On some occasions with an easterly flow, ice crystals cause so much radio static that instrument landings become hazardous. Stratus with base at 300 feet or less may be carried in from the Davis Strait in easterly situations in summer, or may occur with frontal passage.

Gander (*Newfoundland*). Fog and very low cloud are frequent in all seasons. Often the fog or low stratus is formed by advection in warm southerly air as it passes over the Labrador Current and is then carried over the airport by wind from a more easterly point; such conditions are especially frequent in spring. In winter the prevailing air mass is polar; conditions are generally good except in the frequent falls of snow, when visibility is seriously reduced. Gales are frequent from October to April. Strong winds and gales from a westerly point in winter cause drifting snow, so reducing visibility, sometimes for days after the actual fall has ceased.

Montreal/Dorval. This airport lies in the track of many eastward-moving depressions and experiences much variability of weather. Snow is frequent from November to April; flying conditons on the whole are poorest during winter, mainly because of snowstorms with their reduced visibility and low cloud bases. Radiation fog is infrequent and occurs only in autumn but visibility may be reduced by smoke from Montreal, expecially in winter.

New York/La Guardia. The influence of the Gulf Stream on the winter climate of New York is limited since the prevailing winds in that season are off-shore. Cyclonic and frontal activity is vigorous in this area in all seasons. Visibility is seriously reduced at the passage of fronts, particularly warm fronts, when it may fall to about 500 metres. Apart from occasional radiation fog, visibility is generally good in anticyclonic conditions in autumn and winter but in spring the visibility deteriorates as the frequency of warm fronts increases. Visibility is also reduced by smoke in suitable conditions. Very low cloud may last throughout the day in autumn and winter. Thunderstorms occur mainly in summer.

Bermuda/Kindley Field. Cloud base seldom falls below 600 feet except in spring and even then the cloud is usually well broken. Fog is very rare, the lowest visibilities occurring in heavy rain. The winter is stormy with severe gales; gales also occur in summer and autumn in association with tropical revolving storms. The latter are liable to affect Bermuda between August and November; on average about one per season passes sufficiently close to give winds of gale force but hurricane winds (force 12) are rarely experienced.

Santa Maria and Lajes (*Azores*). The main weather influence is that of the Azores anticyclone, the position of which varies from day to day. In winter and spring, the Azores lie most frequently within the zone of the disturbed westerlies but in summer they are more often within the area of the Azores anticyclone or in the zone of the north-east trades. Far from being uniformly quiet, the weather here is changeable and cloudy, frontal passages are frequent and wet stormy conditions are common except in summer. Cloud base at the airports often falls below 1000 feet especially at Santa Maria. In strong winds from certain directions, landing conditions are apt to be difficult on account of the varied topography. Radiation fog does not occur and sea fog is uncommon.

169. TRUNK ROUTES TO THE WEST INDIES AND SOUTH AMERICA

Two main routes are concerned here – that from western or north-west Europe via the Azores and Bermuda to the West Indies (which has been partly described in the previous section) and that to West Africa and thence across the central Atlantic to Brazil. The area to be considered therefore extends some 30° of latitude either side of the equator and so includes the two trade-wind zones and between them the equatorial trough. The intertropical front lies north of the equator throughout the year except near the coast of Brazil where its mean position varies between about 5°S in January and 5°N in July; the corresponding variation on the coast of West Africa is from about 5°N to 15°N. The characteristics of the weather over the open sea on these routes are as described in Chapter 22; the weather generally is good and disturbances are unusual except where the route crosses the intertropical front or if there is a hurricane in the vicinity.

African coast

Near the African continent there are however important local variations which require mention. The Canaries Current off north-west Africa flows towards the equator; it is therefore cool for the latitude and the low temperature is enhanced by the effect of the trade winds in driving the surface water off-shore with the consequent upwelling of colder water from below. The result is that a belt of fog forms over the cold water some distance off-shore and this fog then drifts on-shore when the wind is favourable, even the sea-breeze having this effect at times.

Harmattan. Over north-west Africa the harmattan is a dry wind associated with the north-east trade wind. On the coast west of the Sahara, it is pronounced from November to April, this being the dry season in those parts which in summer are influenced by the south-west monsoon. The length of the harmattan period decreases southwards, the Guinea coast being affected only occasionally, during temporary recessions of the monsoon. The harmattan carries much dust, often reducing visibility below 4000 metres and at times even below 1000 metres. The dust haze may be encountered up to great heights and is found aloft where the harmattan overruns the monsoon.

West Indies

The West Indies lie in the path of the trades throughout the year, which here blow from almost due east. Land- and sea-breezes are of daily occurrence. After the long sea track, orographic cloud forms readily on the high ground and the windward slopes receive rain in all months; in summer the rainfall is accentuated by convection and is more generally distributed. There is much topographical variation and very heavy falls occur in exposed places. Hurricanes in this area average about three per season, the great majority occurring between August and October. They sometimes originate within the Caribbean Sea but more generally to the east, occasionally as far east as the Cape Verde Islands. After formation they move towards west or north-west and while some continue into Mexico, most recurve across the islands or the Florida peninsula and pass into the North Atlantic.

South America

The north and east coast of South America as far as about 30°S is subject to prevailing on-shore winds throughout the year. These are the north-east trades north of the intertropical front and the south-east trade winds to the south, although the direction of the latter undergoes some modifications, especially on the south-east coast of Brazil; there the wind blows from east or north-east particularly in summer (January) when there is monsoonal inflow to the heated interior. As the ocean currents off the whole coast are warm, fog and very low cloud base are not characteristic as they are of parts of the west African coast.

On the Atlantic coast north of the equator the northern winter months are on the whole a dry season and the summer a wet season with much cloud and heavy rain. The Amazon basin has heavy rainfall generally, with frequent thunderstorms, although there is a somewhat drier season in the lower basin in the southern winter and spring. From the equator to about 20°S the normal tropical seasons are found with abundant rainfall in the southern summer and autumn while the rainfall in winter and spring is very light. Exceptions are the coastal strip from Bahia to Natal which has a marked winter (July) maximum of rainfall, while inland certain upland areas receive very little rain.

The area to the south of 20°S, including the northern half of Argentina, has rain in all seasons in association with the trailing cold fronts of the travelling lows of higher latitudes and with the shallow lows which form on those fronts. Northern Argentina is known for the 'pampero', a southerly (polar) wind which sets in with violent line-squall characteristics and carries clouds of dust from the pampas. The southern half of Argentina comes more directly under the influence of the temperate lows but as the rain-bearing westerly winds have to cross the Andes, precipitation is very light although well distributed throughout the year. Cloud amounts here are relatively small and marked föhn effects and turbulence are observed with the strong westerlies.

Upper winds

Between Central America and West Africa the upper winds consist essentially of the equatorial easterlies flanked by westerlies on either side. The zone of the easterlies is situated mainly to the north of the equator in July and to the south in January; the speeds are mostly light at least up to 55 000 feet but tend to be rather stronger towards Africa, especially in the northern summer.

Terminal conditions

Gibraltar/North Front. The aerodrome is situated immediately north of the Rock which rises to 1300 feet above MSL. An easterly wind known as the 'levanter' is most frequent from July to October and in March; it is moist and produces the characteristic banner cloud (stratus) which extends from the summit of the Rock a mile or more to leeward. Fog is frequent in these conditions especially after dawn and dusk and in the first two or three days of a levanter spell. The wind most favourable for fog on the aerodrome is a light north-easterly. There is dust haze with southerly winds in summer. Winds between south-east and south-west make landing difficult or even impossible on account of turbulence in the lee of the Rock. The south to south-west sea-breeze, which is frequent from April to September, may reach force 5 and it too makes landing difficult.

Dakar. The intertropical front passes just north of Dakar in summer, making the months July to September a wet and thundery period. The rest of the year is very dry, with less cloud. Visibility is generally good except for periods of harmattan haze in the dry season.

Recife (Brazil). Although the airport is subject to much cloud and rain during the wet season (March to August), the cloud base and visibility are seldom low enough to cause difficulty in landing. Cloud bases below 1000 feet and visibilities less than 5000 metres almost always occur with winds from a southerly point.

Piarco (Trinidad). Cloud base seldom falls below 1000 feet and practically never below 600 feet. Visibility is generally good and fog negligible. The wet season extends from June to December, the rain being of convective type; thunderstorms are most frequent in August and September.

Kingston (Jamaica)/Palisadoes. Cloud bases below 1000 feet are rare except in December, while bases below 600 feet are practically unknown. Fog does not occur. Winter is the dry season here; at other times precipitation is of convective type and thunderstorms reach a maximum frequency in September. Hurricanes are experienced occasionally.

Montevideo/Carrasco. Fog or low stratus occurs with warm fronts to the north. Light or moderate winds from between north-east and south-east give advective fog or low stratus in autumn and winter, and sometimes drizzle. With light westerly winds, mist usually forms over the swamps east of the aerodrome although visibility generally remains above 2000 metres. Cold fronts, approaching from south-west, occur in all seasons; in autumn and winter, weak or occluded fronts become more active while crossing the estuary of the River Plate; in spring and summer, squalls occur with fast moving cold fronts. South-easterly gales develop in winter with a depression centred to the north of Carrasco; they may be accompanied by very low cloud and two to four days of abundant precipitation. Thunderstorms are of common occurrence with a maximum in summer.

Santiago (Chile). The altitude here is about 1700 feet and an outstanding feature is the great frequency of very low cloud; even in summer one day in five has cloud below 600 feet in the early morning, while in winter the frequency is twice as great. Radiation fog occurs at all seasons but is most frequent in winter.

170. TRUNK ROUTES TO ATHENS VIA MUNICH OR ROME AND THENCE TO CAIRO AND BEIRUT

An account of these routes requires a discussion of the meteorology of Europe and the Mediterranean. These two areas have very different climates but the boundary between the two régimes is not so much the coastline as the mountain ranges of southern Europe.

Europe

The weather of Europe is determined primarily by its situation in relation to the Icelandic low, the Azores high and the alternating high- and low-pressure systems of Asia. The general drift of weather is from west to east and the absence of any pronounced north-to-south mountain barrier enables cyclonic systems from the Atlantic to penetrate far into the continent. This region is therefore one of transition between oceanic and continental conditions, the latter becoming gradually more dominant towards the east and south-east.

Cyclonic systems. The most favoured tracks for lows from the Atlantic are as follows:

(i) To the north of the British Isles into the Barents Sea and thence into north Russia.
(ii) East-north-east across the British Isles or France to north-west Russia.
(iii) From Iceland south-eastwards to the Baltic Sea and thence to western Russia.

The areas to the south of the tracks are affected principally by the fronts of the depressions and by alternate warm and cold air masses. A common situation occurs with a low over northern Russia and the cold front trailing across Poland to central France. On approaching the Alps, the cold front often slows down and gives rise to a wide belt of cloud and rain. Waves on the front may develop into small but vigorous lows which move rapidly east-north-east. Sometimes the main low to the north-east becomes very deep and a broad current sweeps over western Europe, carrying the cold front through to the Mediterranean.

Occasionally lows develop over the continent itself, more especially in summer and in a moist unstable air mass. These thermal lows give much rain and thunder with extensive masses of cloud, but outside the rain areas the clouds are usually isolated.

Another type of low develops chiefly in winter and spring between the Alps and the middle Danube in association with warm moist air spreading northwards from the central and eastern Mediterranean. The warm front is better defined in the upper levels than at the surface but it gives rise to extensive low cloud over Germany and Poland, possibly extending to the Low Countries and even to eastern England. Precipitation is also widespread and may reach the ground as snow.

Anticyclonic systems. The anticyclones affecting Europe are mainly of three types:

(i) Extensions of the Azores high – these consist of tropical air and occur mostly in summer.

(ii) Extensions of the Siberian high, consisting of polar continental air and occurring chiefly in winter and early spring.

(iii) Those which approach from north-west following a cold front associated with a depression over northern Scandinavia or Russia – these consist of polar maritime or even Arctic air and occur in all seasons.

Whereas anticylones of type (iii) are usually migratory, those of (i) and (ii) often persist for several days or even weeks.

Wind. Great variability in both speed and direction is found at all levels although the prevailing direction is westerly. Periods of easterly or north-easterly winds are usually associated with a westward extension of the Siberian high. The westerlies generally increase with height; at about 30 000 feet in the jet streams, the axes of which are often situated parallel to the surface fronts but displaced a few hundred miles towards the cold side, speeds of well over 100 knots are common. Winds in summer are generally lighter than in winter.

Cloud and precipitation. With the prevalence of cyclonic activity in winter (especially in the west and north) and of convection in summer, there is much cloud and rain throughout the year. The mean cloud amount varies only from about 6 oktas in winter to 5 oktas in summer and shows little variation geographically, but the annual rainfall decreases steadily from about 1000 millimetres in the extreme west to less than 500 millimetres in the east, except for the increased falls which are to be expected on high ground. Away from the western coastal strips where most rain falls in winter, the wettest period is usually late summer or autumn and the driest period late winter or early spring. Precipitation is liable to fall as snow during the winter months, more especially in the east and south-east where the ground may remain snow-covered for long periods.

Visibility. The greatest difficulty for aviation in Europe is the high frequency of fog and very low cloud. Both of these occur readily in air masses of maritime origin, little cooling being required to produce condensation. The fog may become widespread and dense in anticyclonic conditions and is aggravated by smoke in industrial areas. In summer, fog is infrequent except over the sea and on coasts where it drifts on-shore. Cloud on the surface of hills is liable to occur at any time of year.

Ice accretion. The 0°C isotherm in winter is often at or near the surface, especially in central and eastern Europe; combined with the large cloud amounts, this results in a high frequency of conditions favourable for airframe icing. Even in summer the risk remains high. Severe conditions occur when an unstable maritime air mass passes over a coast or over hilly country with the formation of extensive convective clouds. This may occur at times for example over north-west and central Germany with an air supply from the North Sea; icing is also likely to be severe in a mass of warm-front cloud which develops instability on approaching a mountain range.

Mediterranean

In winter the Mediterranean Sea is warmer than the surrounding lands and it becomes a region of relatively low pressure between the Asiatic high and an extension of the Azores high over north Africa. The Black Sea and the Caspian Sea are similarly regions of low pressure. Various polar and tropical air masses enter the Mediterranean basin and some frontal depressions develop; however, a larger number are formed as lee depressions associated with outbreaks of cold air of which the principal and often the only frontal feature is the cold front preceding such an outbreak. Because of the great dryness of the tropical continental air masses from north Africa, warm fronts and warm sectors are not conspicuous. The vigorous cyclonic activity together with the showers and squalls which develop in the cold air masses, rendered unstable on crossing the warm water, constitute the main difficulty for aviation in this region. Considerable rainfall also occurs, making winter the wet season, but fine periods occur much more often than in the British Isles and visibility is usually good.

In summer the Mediterranean Sea is cooler than the surrounding lands. The Azores high extends over the western Mediterranean, while the monsoon low of Asia is well developed; thus light northerly or north-westerly winds are prevalent at the surface apart from land- and sea-breezes. Cyclonic activity is slight, and quiet cloudless weather predominates. There are however occasional outbreaks of local thunderstorms over the adjacent high ground.

Upper winds. Over the Mediterranean, upper winds are generally westerly throughout the year. In winter the eastern Mediterranean comes under the influence of the northern flank of the subtropical jet stream and the mean vector speed reaches high values, for example, in January at about 40 000 feet it reaches or exceeds 100 knots on the African coast and about 80 knots at Cyprus. The western Mediterranean is hardly affected by the jet stream, and the highest mean vector speeds in any month at Gibraltar and Malta are about 40 knots only.

Local winds. The topography of the Mediterranean basin results in many local winds of importance. Prominent among these are the following:

Mistral. A strong and sometimes violent wind which blows from north or north-west in the Gulf of Lions, adjacent land regions and, in particular, in the Rhône valley; it occurs with outbreaks of polar air often in association with a low over the Gulf of Genoa. Conditions are usually cold and turbulent with little cloud except possibly in association with the initial cold front and with any secondary fronts which are carried south in the stream of polar air. The mistral is most common in winter and early spring and the speed may be intensified locally by katabatic effects and by canalization through the Rhône valley and the Toulouse gap.

Bora. This is a similar type of wind to the mistral, but if affects the eastern shore of the Adriatic. At Trieste the speed has been known to average 70 knots with gusts exceeding 100 knots.

Sirocco. The sirocco is a warm southerly or south-westerly wind in advance of a depression situated in the western Mediterranean. When this wind leaves North Africa it is hot and dry but often dust-laden. On crossing the sea the air is moistened, cooled and stabilized in the lower layers and by the time it reaches the northern waters, coastal and sea fog and perhaps drizzle may occur with extensive but shallow stratus cloud.

Khamsin. A south to south-west wind, similar to the sirocco, which blows over Egypt ahead of depressions passing eastwards along the Mediterranean. As this wind blows from the interior of the continent it is hot and dry and often carries much dust.

Terminal conditions

Marseilles/Istres. Fog is frequent on surrounding low-lying ground and in valleys but rare on the airfield itself. The mistral blows across the airfield from north-north-west.

Rome/Ciampino. Occasions of fog or of cloud bases below 1000 feet are very infrequent, especially in summer. Weather is predominantly good in summer except for an occasional thunderstorm. At other times, certain characteristic winds and their associated weather may be noted. With a low in the Gulf of Genoa, the south-easterly sirocco may give much cloud and rain. The libeccio is a strong squally wind from south-west. Cold air after invading the Gulf of Lions may reach Italy as an unstable north-westerly known as the maestrale and is associated with heavy falls of rain. The tramontana is a north-easterly wind which occurs with low pressure over the Adriatic and a high further west; it usually lasts for several days and is accompanied by fine weather.

Düsseldorf. Shelter is provided by the high ground to the south-east, while to the north-west the airport is exposed to much bad weather. Winds from between south-west and west are particularly gusty. In the sector south-west to east-south-east, winds are canalized from south-south-east along the valley of the Rhine. Visibility is poor in smoke brought from the Ruhr by light north-easterly winds, and the frequency of morning fog is increased when there are light westerly winds in an anticyclone.

Munich/Riem. The airport lies some 30 miles north of the most northern range of the Alps and is sheltered when winds are from south-east to south-west. In winter with winds from other directions cloud bases down to 400–600 feet or even less occur in high pressure situations and at frontal passages. Also in winter there is a risk of sudden fog when pressure is high and wind northerly but it often clears in the early part of the night with wind backing to southerly. In winter and spring with low pressure over Hungary, cloud base may fall to 300–500 feet with precipitation. Haze from Munich with winds west-north-west to north-west may reduce visibility to 2–5 kilometres.

Athens/Hassani. Visibility is good and although there is often much low cloud in the vicinity, over the immediate surroundings of the aerodrome there is usually a marked clearance. Strong winds, particularly from between north-west and north-east cause difficulty in landing aircraft since the winds are very unsteady in direction; this occurs particularly in summer, the period of the so-called Etesian or northerly winds. A strong squally easterly wind causes a similar difficulty.

Cairo. Early morning fog may occur in any month, mostly with a north-easterly wind, and usually disperses before 1000 local time. Low stratus is liable to occur in the early mornings from May to September when northerly winds prevail; it occasionally lowers to the surface as fog. Sandstorms are caused by strong to gale-force winds from between south and west-south-west in association with the passage of depressions during the period December to May.

Beirut. The airport is sheltered from winds between about east and south-east by the high ground but a north-easterly katabatic wind often reaches 15–20 knots during summer nights. Surface winds above 25 knots, especially in thundery squalls, raise sand which may reduce visibility to less than 50 metres. Fog is practically unknown.

Nicosia. Because of the sheltering effect of the Kyrenia and Troodos mountain ranges, the most common wind directions are from between west and north-west and between east and south-east. Radiation fog is mostly confined to the low ground below aerodrome level at the foothills of the Kyrenia range but occasionally spreads to the airport; it occurs most frequently during the summer months in the dawn period.

171. TRUNK ROUTE CAIRO TO JOHANNESBURG

In winter a tongue of high pressure from the Azores anticyclone extends across the Sahara. Although the lows of the Mediterranean bring rain to the north coast of Egypt, the falls diminish rapidly inland and at Cairo the annual total is only 25 millimetres. In the practically rainless belt which extends from Cairo to about 6° N prevailing surface winds are mostly north to north-east; the air is cloudless but often dust-laden. In July the intertropical front reaches its northern limit of nearly 20° N and is associated with the severe duststorms known as haboobs which are frequent in the Sudan at this season. The tropical rain belt at this time extends from the equator to the neighbourhood of Khartoum (16° N) where rainfall averages 150 millimetres in summer, but the rest of the year is entirely dry. Ethiopia similarly experiences a wet summer, the higher parts receiving very heavy falls of rain with much thunder, while at other times the country is mostly dry. Kenya, which lies across the equator, is subject to two wet seasons (March to May and November to December) corresponding to the northward and southward passage of the equatorial trough, but the intervening months are not without rain. In the southern summer the north to north-east winds of north Africa extend far south of the equator in association with the southward displacement of the intertropical front which in January lies across Rhodesia. Thus Tanzania, Malawi and Rhodesia have most of their rainfall in the southern summer, from November to March or April. Throughout the various rain areas there is much convection, and frequent thunderstorms occur. The central plateau of Rhodesia and occasionally the Transvaal are affected by the 'guti' – very low stratus and stratocumulus cloud which occur with moderate to strong south-east winds bringing moist air from the Mozambique Channel. It occurs in spells of one to five days, especially in the dry season from April onwards.

The southern part of the route is very occasionally influenced by tropical cyclones in the Mozambique Channel. Cyclones in the south Indian Ocean originate in the area to the east of the Seychelles between about 5–15° S; they usually travel first south-westwards and subsequently recurve towards south-east. They seldom reach the Seychelles themselves but frequently affect Madagascar, Mozambique, the Comoro Islands, Réunion and Mauritius. They are liable to occur from November to May and are most frequent from January to April.

Upper winds

The main structure comprises two zones of westerly winds separated by the equatorial easterlies. In the north the westerly subtropical jet stream is centred at about 25–30° N in winter; the high speeds reached have been noted in Section 170. The easterlies reach their greatest speed in the summer hemisphere; in particular, in an easterly 'jet stream' at about 15° N the mean speed exceeds 60 knots at 55 000 feet in July. Nairobi (1° S) has light easterlies, with a few exceptions, all the year and at all heights to over 50 000 feet. Direct information is scanty at present for Africa south of the equator, but the main structure appears to be similar to that of the northern hemisphere in the corresponding season although the equatorial easterlies are weaker to the south of the equator in January than they are to the north in July.

Terminal conditions

Cairo. See Section 170.

Khartoum. Sand and dust are almost the only difficulties here. In winter, visibility may be reduced to between 500 and 2000 metres for 24-hour periods by drifting sand with strong northerly winds; still worse visibility can occur for a short time near cold fronts. In summer, haboobs approach from east or south-east and gusts may spring up from near calm to over 40 knots; visibility is often less than 500 metres during the first half-hour of the storm and subsequently less than 2000 metres for periods up to 3 or 4 hours. On a few occasions in summer, strong, south-westerly winds raise sand which reduces visibility to less than 500 metres. Cloud base below 2000 feet occurs only for short periods, usually after late-night thunderstorms in summer.

Nairobi/Embakasi. Low stratus, often falling to the surface as fog, is frequent between 0200 and 0800 local time, especially in the 'long rains' of March to May and the 'short rains' of November to early January. Monthly average frequencies of fog vary from about 18 days in December to only 3 in July. Occurrence of stratus below 400 feet varies from 25 days in November to 7 in August.

Afternoon thunderstorms may occur at any time of the year but are very infrequent in the period June to September. Storms are unlikely to reduce conditions below operational limits for longer than an hour unless they occur at night or in the morning when their extensive rain areas may give widespread very low cloud and poor visibility.

Nocturnal rain is fairly frequent in both 'rains' periods.

Entebbe (*Uganda*). There are two main wet seasons – mid-March to mid-May, and late October to mid December. Thunderstorms, heavy at times, occur throughout of the year, mostly in the early hours and the great majority between 0200 and 0500 local time; they are least frequent from December to February and from June to August. Widespread rain and severe thunderstorms are liable to occur in unstable air from west or south-west, particularly in the wetter seasons. The occasions both of visibility less than 1000 metres and of cloud base below 1000 feet are generally small in number and of brief duration. Visibility may be reduced to about 5000 metres in haze from December to early March with northerly winds. Landing conditions are said to be generally most favourable from 0900 to 1200 and from 1600 to 2300 local time.

Livingstone (*Zambia*). The limiting factor here is mainly low stratus during the wet season (November to April), more especially in January and February. The stratus usually clears after 1000 local time, but there are some occasions associated with the intertropical front when low cloud persists all day. Thunderstorms are frequent during the wet season.

Salisbury (*Rhodesia*). Weather is fine in the cold season, May to mid-August. Smoke haze occurs from mid-August to mid-November; there are also occasional thunderstorms, and rapid local fluctuations of wind on account of turbulence and dust devils. In the wet season, mid-November to April, heavy thunder showers reduce visibility to less than 500 metres. A period of several days of bad visibility may occur at this time if a tropical revolving storm is near; also any influx of maritime air generally gives long periods of low stratus or stratocumulus.

Johannesburg/Jan Smuts. Very low cloud is frequent in the early mornings during the wet and thundery period October to March; cloud bases below 500 feet tend to occur with north-north-easterly surface winds. The thunderstorms usually approach from between south-west and south-south-east although the surface wind at the time is often north-westerly. Fog occurs on a few days each month throughout the wet season, visibilities of 800 metres or less being associated with a north-north-westerly surface wind, but dispersal takes place early in the day. Fog also occurs at times in winter with easterly winds, while smoke may reduce visibility to 2 to 5 kilometres in south-south-westerly winds on winter mornings.

172. TRUNK ROUTES FROM BEIRUT AND CAIRO TO BASRA AND BAHRAIN

The area concerned here is largely arid. The winter weather is controlled by the Asiatic high and by the occasional lows which move eastwards from the Mediterranean and pass by way of the Persian Gulf to Afghanistan and northern India. Away from the coast of the Mediterranean there is little rainfall; what there is occurs in association with the travelling lows and in spring it is often thundery. Over the deserts, widespread duststorms occur with the southerly or south-easterly winds ahead of the lows. In summer the area is almost entirely rainless. Over Iraq the north-westerly wind known as the 'shamal' is very persistent and carries clouds of dust by day.

In winter the subtropical westerly jet stream covers the Persian Gulf area and much of Arabia; in places the mean speed exceeds 100 knots at heights between about 35 000 and 50 000 feet. In midsummer the low-level north-westerlies of the Persian Gulf give way to light easterlies above about 12 000 feet.

Terminal conditions

Beirut, Cairo. See Section 170.

Basra. The aerodrome is subject to frequent dust haze throughout the year. June to August is a dry and almost cloudless period of prevailing north-west to west-north-west wind which, probably on account of sheltering by an extensive belt of date palms, rarely exceeds 16 knots; dust often obscures the sky and reduces horizontal visibility to 2000–3000 metres and to below 1000 metres on the rare occasions of strong wind. Thick rising dust also occurs during other months whenever the winds are strong enough from a north-westerly or south-easterly point. Thick fog in early morning occurs during periods in winter and early spring, while with very light winds smoke haze from nearby towns may reduce visibility in the early morning to about 2000 metres. Except for lifted fog, the cloud base is usually well above 2000 feet, even in rain.

Bahrain. Radiation fog occurs occasionally in autumn and winter but is seldom persistent. From June to August visibility is often reduced by dust and for long periods there may be considerable haze; duststorms can occur at any time but are most persistent in summer. Cloud base below 1000 feet is rare except in lifting fog in autumn. The land-breeze from west-south-west and seabreeze from east-north-east are pronounced in summer and autumn.

173. TRUNK ROUTES FROM BASRA AND BAHRAIN TO KARACHI OR BOMBAY AND THENCE TO CALCUTTA

These routes traverse an area where the weather is dominated by the monsoons of Asia. In winter when the Siberian high is established the interior of the continent becomes a source region of polar continental air – dry and cold – which flows out over the surrounding areas; in summer the high gives way and a low to the north-west of India draws in moist equatorial air from the Indian Ocean. Thus the direction of the low-level circulation in summer is largely the reverse of that in winter, and spring and autumn are transitional seasons in which there is no very definite circulation.

During the north-east monsoon (December to February) the routes are sheltered by the mountains to the north from the direct outflow of continental air. Weather is predominantly fine with little cloud and good visibility, but the shallow depressions which occasionally move eastwards from the Mediterranean bring some cloud and rain to the routes. In the transitional period (March to mid-June) the weather is mainly hot, dry and dusty; the weak lows which continue at intervals to traverse northern India from west to east are at this time associated with severe squalls and duststorms which in Bengal are known as 'nor'westers'.

Meanwhile pressure begins to fall in north-west India and the south-east trade winds cross the equator to feed the south-west monsoon which lasts from about mid-June to mid-September. The onset of the monsoon takes place suddenly at any one place; it reaches the west coast of India in early June and then extends north-eastwards. After its long sea passage the air is moist and convectively unstable and heavy orographic and convectional rain become widespread. The monsoon cloud and rain just reach the coast of Baluchistan, but further west and in the Persian Gulf conditions remain arid. Tropical cyclones form in the Arabian Sea and Bay of Bengal during the advance and retreat of the monsoon, that is, when the inter-tropical front lies over these seas; other cyclonic storms form over the Bay during the monsoon. Some of these move north-westwards and give copious rain in the Ganges valley. Here the airstream is south-easterly but as it progresses inland it dries out and north-west India gets little of the monsoon rains. Flying conditions in the south-west monsoon are difficult on account of the severe turbulence, the massed clouds and rain and the obscuring of high ground; however, visibility is good outside the cloud and rain; the bad weather tends to be patchy and some days are better than others, even to the extent of occasional 'breaks' in the monsoon. After the retreat of of the monsoon during September, the weather generally improves.

Upper winds

The winter anticyclone over Asia is shallow and westerlies are established over the routes by 10 000 feet. The axis of the subtropical jet stream lies at about 25–30° N and from about 40 000 to 50 000 feet in January the mean speed of the westerlies exceeds 100 knots near Basra and again near the north-east border of India.

The summer low over north-west India also declines with increasing height and hardly extends above 15 000 feet. The south-west monsoon current in this range becomes more westerly aloft with little change in speed; above about 20 000 feet there are light or moderate easterlies on the routes, the axis of the 'equatorial' easterlies here in summer being at about 15° N.

Terminal conditions

Basra, Bahrain. See Section 172.

Karachi. Early morning fog occurs occasionally from October to March and usually clears at about sunrise. Dust haze is frequent and dust is raised from the ground by winds exceeding 20 knots, usually in summer. Much low stratus and stratocumulus occurs during summer while occasional convection clouds with thunderstorms and heavy showers occur in the south-west monsoon which affects the airport during July and August.

Bombay/Santa Cruz. Fog is very infrequent but a few occasions may be expected at night in February and March with light winds from north-west or south-east. Cloud base seldom falls below 600 feet even in the south-west monsoon, but neighbouring hills are obscured at times. Thunderstorms are most frequent in June, September and October and to a lesser extent in May; they drift over the aerodrome from east or south-east with associated squalls. A marked sea-breeze blows from west or north-west, and there is a north-easterly land-breeze except during the south-west monsoon.

Calcutta/Dum Dum. During winter, fog occurs with light winds from between south-east and south-west; early morning fog is prone to thicken in the presence of light north-west to north-east wind. During late winter and the early hot season, low stratus, base 200–800 feet, occurs with moderate to strong south to south-west winds when an inversion is present. In the hot season, dust haze occurs with west to north-west winds in association with an upper inversion. Nor'westers occur during this season and approach mainly from between west-north-west and west; the wind speed often exceeds 50 knots when the squalls arrive in the evening.

In the south-west monsoon, squally thunderstorms are associated with wind changes from east-north-easterly to south-westerly or vice versa as the axis of the monsoon moves northwards or southwards across the aerodrome. Steady winds from between north-east and south-east at this time are usually associated with good visibility and broken cumuliform clouds.

174. TRUNK ROUTES FROM CALCUTTA TO RANGOON AND SINGAPORE

The weather of these routes is controlled by the monsoons which have been referred to in the previous section. During the north-east monsoon the first part of the routes lies in the lee of the mountains and flying conditions are excellent with little cloud and good visibility. Malaya, however, is exposed to the north-east monsoon which arrives there after a long passage over the South China Sea during which heat and moisture have been absorbed; hence there is much convectional cloud and precipitation and frequent thunderstorms. The west coast of Malaya however is somewhat sheltered and provides the better route between Rangoon and Singapore at this season.

During the south-west monsoon, the cloud and rain are accentuated orographically over the Arakan coast but Malaya is to some extent sheltered by the high ground of Sumatra. The better route now follows the east coast of the peninsula. There is however no dry season in Malaya – convective cloud and precipitation are plentiful throughout the year.

The bulk of the cloud is convective consequent upon day-time heating, and therefore there is a diurnal variation of cloud amount with thunderstorms occurring in mid- or late afternoon. This diurnal pattern, although evident well inland on most occasions, is complicated by such factors as the formation of sea-breezes in coastal regions, Sumatra storms, the presence of hills and valleys, and the passage of convergence lines. In general, cloud is at a maximum well inland between midday and midnight, and over the sea it is at a maximum between midnight and midday whilst coastal stations tend to have more cloud during the early and latter parts of the night than during the day.

The occurrence of tropical and other cyclones in the Bay of Bengal has been noted in the previous section. On occasions between July and November a tropical cyclone moves from the China Sea into the Gulf of Siam and very rarely one crosses the Malay peninsula into the Bay of Bengal. During the south-west monsoon the Malacca Strait is subject to violent thundery squalls in which the cumulonimbus assumes a characteristic arched shape; these squalls are known as 'Sumatras' and usually occur at night since they are initiated by katabatic winds. Over the Malay peninsula the wettest periods are the transitional seasons when the intertropical front moves across the area.

Upper winds

In winter, upper winds are westerly in the north and light easterly in the south of the routes. In summer the winds are mostly light westerly in the lower levels but easterly above 20 000 feet with mean speeds increasing near Rangoon to about 70 knots at 55 000 feet in July.

Terminal conditions

Calcutta. See Section 173.

Rangoon/Mingaladon. Early morning fog or low stratus occurs occasionally during the dry season (late October to April); the fog or low cloud usually disperses within a few hours after sunrise. Cloud base below 1000 feet often occurs during the wet season (May to early October). Of the mean annual rainfall of 2500 millimetres at Rangoon, all but some 175 to 200 millimetres falls in the wet season.

Singapore. Low cloud, poor visibility and strong winds occur in general only in association with rain, which is mainly in the form of scattered showers or thunderstorms and occurs on about one day in two throughout the year. The showers may be heavy at times, reducing the visibility to a few hundred metres and cloud base to a few hundred feet, and with associated gusts of about 25 knots (60 knots have been recorded). Prolonged periods of rain are rare except occasionally during the north-east monsoon. The diurnal variation of rainfall shows a late afternoon peak during the north-east monsoon from November to March, but during the south-west monsoon from June to September the peak rainfall occurs around dawn or during the forenoon, mainly as a result of the 'Sumatra' storms referred to above.

175. TRUNK ROUTE FROM SINGAPORE TO DJAKARTA AND DARWIN

This route has an equatorial rainy climate except near Darwin where there is little rain from May to September. In January, the north-east monsoon reaches the area moist and unstable after a long sea track; south of the equator it veers to become the north-westerly monsoon of northern Australia which then lies within the equatorial trough. In July, Australia is in the belt of subtropical high pressure; south-east trade winds blow from the continent and from the Pacific Ocean towards the East Indies and veer to become the south-west monsoon north of the equator. Generally there is much convective cloud with frequent heavy showers and thunderstorms but the topography of the islands is extremely varied and there are often marked contrasts of climate between the windward and lee sides. Land- and sea-breezes are a regular feature. That part of the south-east trade wind which originates over Australia in the southern winter is dry and dusty over the continent and is associated with a period of haze and reduced rainfall in the islands south of about 5° S. Generally the wettest periods at any place occur when the intertropical front is in the vicinity, that is, in one or other of the transitional seasons. Tropical cyclones known as 'willy-willies' develop in or near the Timor Sea from January to March. They usually move at first south-westwards and after recurving pass into north-west Australia; they are accompanied by heavy rain and strong winds or gales.

Upper winds

In this region the upper winds are mostly the equatorial easterlies except that winds with a westerly component occur north of the equator in the northern summer monsoon, which extends up to about 15 000 feet.

Terminal conditions

Singapore. See Section 174.

Djakarta. In the west-south-west monsoon period from October to March there are frequent prolonged heavy showers but visibility is otherwise good. The east-south-east monsoon from April to September is comparatively dry but hazy. Throughout the year, surface winds rarely exceed 15 knots and the sea-breeze comes in from the north almost every afternoon.

Darwin. In the dry season (May to October) the south-east trade wind prevails and at times reaches 25 knots. Dust haze is prevalent and may reduce visibility to less than 2000 metres. The period of the north-west monsoon (November to April) is a very wet season; winds are light except for the sea-breeze which blows strongly in the afternoon from north or north-west, even in the dry season. In the wet season heavy rain occurs in squalls and also from isolated convective clouds the base of which may fall to the ground for a short period. Thunderstorms are frequent during this season.

176. TRUNK ROUTE FROM DARWIN TO SYDNEY AND MELBOURNE

In winter (July) Australia comes under the influence of the subtropical high and except in the south-east and south-west, weather is mostly dry with little cloud. The lows of temperate latitudes pass for the most part well to the south of the continent but are associated with troughs and secondaries which give disturbed weather over the southern extremity of the routes; there is also orographic rain in this area from the south-east trade winds as they rise over the highlands near the coast. With the approach of spring the pressure systems shift southwards and the disturbances in the south become less frequent; at the same time in the north of the route thunderstorms begin to occur as the temperature rises. In the southern summer the north-east monsoon of Asia crosses the equator to become the north-west monsoon of northern Australia where it arrives moist and unstable. The inter-tropical front also moves south until in January it lies across the route at about 15°S. Throughout the summer, rain falls in the north in heavy thundery showers, but the amounts decrease towards the interior and the central part of the route has arid conditions all the year.

The subtropical high of the southern hemisphere is not an unchanging system but consists of a series of eastward-moving anticyclones separated sometimes by cols, at other times by troughs of low pressure so that, although fully developed cyclonic lows are rare over the continent, conditions are variable from day to day. Hot dry air from the interior is associated with severe duststorms, while the passage of a cold-front trough brings a sharp fall of temperature and strong to gale-force winds which constitute the well-known 'southerly busters' of New South Wales.

Upper winds

Except near Darwin in summer, upper winds are mainly westerly throughout the year. In January the axis of the subtropical jet stream lies across New South Wales and the westerlies average about 60–70 knots at 40 000 feet. In July the axis lies over the centre of the continent and average speeds are in the neighbourhood of 100 knots at the same height.

Terminal conditions

Darwin. See Section 175.

Sydney/Mascot. Bad weather occurs typically with one or other of two situations in the summer half-year (November to April). The more serious is that in which south-easterly winds follow a well-defined cold front; within about three hours of the frontal passage, broken stratus with base 400–800 feet drifts in from the sea and is later followed, as the airstream becomes deeper, by showers and further low stratus up to 8 oktas in amount. The other and less frequent situation is the 'black north-easter' which develops ahead of a trough; low cloud forms as in the south-easterly situation but conditions generally are less severe. The sea-breeze also comes in from north-east but this is a shallow current which is not associated with low cloud.

Thunderstorms occur mostly in summer with the passage of cold fronts. The latter are usually of the southerly buster type; the main cloud base is generally at about 5000 feet with broken low stratus below. Dust with north-westerly winds in summer may reduce visibility below minimum requirements for landing. Turbulence is most severe with strong south-westerly winds blowing from the high ground, and also with strong north-westerly winds on hot days.

Melbourne/Essendon. Light winds from the northerly sector bring fog from valleys in winter. With light winds from between 130° and 170°, industrial haze from Melbourne with visibility 1–2 kilometres is frequent from December to March. Stratus at 200–500 feet occurs with light southerly winds. Cold fronts from between south and west are liable to be followed by prolonged bad weather when the air-stream is of southern origin.

177. TRUNK ROUTE FROM SYDNEY TO AUCKLAND

This route lies just within the zone of disturbed westerlies most of the year but comes under the influence of the subtropical high in summer. Changeableness is the general characteristic. Weather is controlled by a succession of lows which mostly move south-eastwards between, or to the south of, Australia and New Zealand. Each low brings its frontal cloud and precipitation; the cold front in particular is usually pronounced and is followed by a cold anticyclone with temporarily improved conditions. In summer (January) the emphasis is on the anti-cyclones which move eastwards in the zone of the subtropical high to the number of about five per month; successive highs are separated by cols or troughs so that some variability in conditions is maintained even in this season. Also in summer, tropical cyclones form over the western Pacific south of the equator; some of these recurve near Australia where they bring heavy rain to the east coast and then cross New Zealand as very deep extratropical cyclones.

On account of the generally sustained movement of the weather systems in the area, prolonged periods of low cloud and precipitation or fog are uncommon and flying conditions on the whole are good.

Upper winds

Upper winds are moderate westerlies except in winter (July) when the route lies on the southern fringe of the subtropical jet stream and the westerlies average 60–70 knots at 40 000 feet.

Terminal conditions

Sydney. See Section 176.

Auckland/Whenuapai. Cloud base may fall to 300 feet and visibility to 1000 metres in moist gradient winds from north-east. In pre-frontal conditions with winds from north or north-west it is rare for cloud base to fall to 300 feet or visibility to 1000 metres; a clearance takes place if the wind backs to south-west after the frontal passage but not necessarily if the shift in only to west. In calm conditions early morning radiation fog is likely; it mostly clears about 2 hours after sunrise but later if there is a light wind from east or east-north-east.

178. TRUNK ROUTES FROM SINGAPORE TO HONG KONG AND JAPAN

These routes come under the influence of the Asiatic monsoons, with two brief transitional seasons. The intertropical front lies across the route in summer; its position is ill defined but in July it probably lies east of Korea to southern Japan. As the winter monsoon sets in so the intertropical front retreats, eventually to well south of the equator. For conditions at the Singapore end of the route reference should also be made to Section 174.

Winter

The monsoon blows from north-west or north over the northern parts and from north-east over the southern parts of the routes. It becomes established gradually during September and October in successive bursts of cold air behind a series of cold fronts. In the north the air is of Siberian origin – dry and cold while over the mainland – but in the crossing to Japan it is warmed and moistened and this process is continued while passing over the warm Kuro-Siwo current (the Pacific counterpart of the Gulf Stream) towards south China and Indo-China. In the north, bad weather is mainly confined to the areas in the rear of the cold fronts, the areas west of the mountains of Japan being particularly affected. From January to April the Siberian air interacts with maritime tropical air giving long periods of low stratus and drizzle with mist or fog, a condition known as the 'crachin'. It affects any part of the coast from Cape Cambodia to Shanghai and in south China it penetrates well inland. The winter monsoon is generally overrun at 7000 feet by a deep westerly current. This is strongest and most persistent in the subtropical jet stream, the axis of which extends from Tibet to southern Japan. In January the mean speed near Yokohama amounts to 150 knots at 40 000 feet and individual wind speeds of over 200 knots in this area are not uncommon at high levels.

Summer

In May, tropical air begins to advance northwards and is heralded by cyclonic activity and the widespread 'plum' rains of China and Japan. The season of the established monsoon over China and Japan is July and August. The air is mostly hot and sultry with much convection cloud and thunderstorms, but there are some minor disturbances which produce widespread rain. Sea fog forms over the cold sea current off the north-west coast of Japan in this season. The summer monsoon may be as much as 15 000 feet deep; it is overlaid in the south by easterlies, and north of about 25°N by westerlies. The China seas and coasts as far south as about 15°N are liable to typhoons from May to November, although July to September is their main season; many of these recurve and move north-eastwards across Japan, but occasionally one enters the Gulf of Siam.

Terminal conditions

Singapore. See Section 174.

Hong Kong/Kai Tak. During periods of crachin (January to April) there is much stratus with base down to 600 feet over the aerodrome and 300 feet over the sea, light rain or drizzle and occasional sea fog. The north-east monsoon period (October to December) is mainly dry with little low cloud. The south-west monsoon period (May to September) has fair periods, heavy showers and occasional thunderstorms; rainfall averages 250–400 millimetres in each of these months. Tropical revolving storms are liable to occur in the vicinity from June to November.

Tokyo. Visibility may be reduced to less than 800 metres by smoke from Tokyo and the local industrial areas; this effect is most noticeable during winter mornings and evenings with light winds. Fog and smoke haze are also frequent in the mornings from April to October with light southerly winds. Cloud base often falls below 1000 feet in association with the warm front of a depression moving eastwards along, or to the south of, the south coast of Japan. Snow is experienced occasionally in winter. The area may be affected by typhoons once or twice a month from June to November.

179. NORTH POLAR AIR ROUTES

Within the Arctic Circle some regions have a continental type of climate, others a maritime climate, and there are variations from place to place and from year to year which make it difficult to generalize. The winter lasts five or six months but the summer only about three months; conditions change rapidly in the brief transitional periods. A period of continuous darkness is centred at the winter solstice and one of continuous daylight at the summer solstice.

Pressure

The north polar region appears to be influenced by lows alternating with highs more or less throughout the year. However, in winter, cyclonic activity is at its lowest since both the arctic and polar fronts are then at their farthest south and the polar highs are correspondingly rather persistent. In this season lows move northeastwards into the Arctic Ocean in an almost continuous series between Greenland and Norway, but east of Novaya Zemlya they stagnate and fill up. Other small depressions form on the arctic front near the New Siberian Islands whence they move eastwards but generally fill up before reaching Alaska. During the summer, lows may penetrate to any part of the arctic and cyclonic activity is quite high in the area from Iceland to Novaya Zemlya.

Temperature

In winter, surface temperature is well below 0°C almost everywhere except in those parts of the Greenland, Norwegian and Barents Seas which are warmed by the Gulf Stream. The sea in the north polar region is permanently frozen and surrounding this ice is a variable field of drift or pack ice. The temperature over the ice fields remains near 0°C even in summer, but elsewhere the summer is mild and overland even warm. At inland places the seasonal variation in surface temperature may be considerable (see Section 156).

The estimated distributions of mean temperature in the vertical in midwinter and midsummer are shown in Fig. 13. In winter the arctic troposphere is on the whole considerably colder than that of the international standard atmosphere and the stratospheric temperatures are also somewhat colder than the standard. In summer the mean temperatures in the troposphere are little different from standard but the stratospheric temperatures are usually appreciably higher than the standard.

Precipitation

Mean annual precipitation decreases northwards and is almost everywhere less than 250 millimetres of equivalent rainfall. Precipitation generally is greatest in summer and least in winter with the exception of the Greenland, Norwegian and Barents Seas where it is usually greatest in winter. The dominant period for snow is from November to April, but in the north it may occur at any time and near the pole practically all precipitation falls as snow. Thunderstorms and hail are almost unknown except in latitudes farthest from the pole and there they are infrequent and occur only in summer.

Cloud

With the exception of most of the region covered by the Greenland, Norwegian, Barents and Kara Seas, conditions are usually much more cloudy in summer than in winter; the latter is the period when clear skies are most frequent. The seasonal variation is well marked in the north and over the pack ice, and the frequency of clear days there during the winter is quite high. In the north polar region and over the pack ice, overcast days are frequent in summer and low stratus cloud is extensive during this period. Over most of the sea areas mentioned there is little seasonal variation of cloud and amounts are large throughout the year; this region is regarded as one of the cloudiest in the world.

During bad weather, cloud tops have frequently been found to exceed 18 000 feet. There is some evidence that the vertical extent of orographic cloud is liable to be much greater than one would normally expect elsewhere. It is probable that here, as elsewhere, the vertical extent of cloud is limited by the tropopause.

Visibility and fog

In the absence of fog and precipitation, the visibility in most of the arctic is usually exceptionally good. However in areas where there is loose snow, even a moderate wind will raise sufficient snow to reduce visibility drastically. The combination of a strong wind and falling or drifting snow results in the well-known blizzard conditions in which visibility is practically nil.

By far the most common type of fog in the arctic is advection fog; radiation fog does occur but is mainly confined to inland areas in Alaska, Canada and Siberia. Over the open sea and the pack ice and in coastal districts, fog is very frequent in summer and infrequent in winter. Over the pack ice and most of the open waters of the arctic seas, fog every other day during the summer is normal, while in some years fog is reported practically every day during the summer in some places. The frequency of these summer fogs diminishes rapidly inland from the various coastlines. Well inland, as would be expected, fog is most frequent during the autumn and winter but these inland radiation fogs are generally much less frequent than the summer fogs of the pack ice and of the sea and coastal districts.

A phenomenon peculiar to the arctic is 'ice fog' in which the fog particles are ice crystals. Ice fog appears to be confined mainly to the land areas but it may occur over the pack ice; it is reported that this type of fog is most prevalent and thickest in inhabited regions. According to reports it is usually shallow; while vertical visibility is reasonably good, horizontal visibility may be very poor; this is clearly of importance to aircraft landing in ice fog conditions. However, a deep layer has been known to form with serious reduction of vertical visibility. Cases have occurred where the take-off of an aircraft has caused ice fog to form on an aerodrome, and motor transport has been known to produce the same effect.

Ice accretion on aircraft

It appears that ice accretion is less of a problem in the arctic than it is in temperate latitudes. Clear ice is infrequent but is most likely to be experienced when flight is in cloud over mountain ranges such as the Greenland ice cap. Rain ice may also occur; in the north this is likely to be very infrequent and confined to the summer months; in the south its frequency will be greater, especially overland, but it will probably occur only during the winter. Rime will be by far the most common type of ice formation. Severe airframe icing does not appear to occur frequently, but once airframe icing has formed, it can persist for a long time after the aircraft has left the icing region and the performance and range of the aircraft may then be adversely affected.

Aurora

The aurora borealis, commonly called the northern lights, is of importance to aircraft operations in the arctic for two reasons. First, the aurora could very well illuminate a target at night-time; when the aurora is at its brightest, the illumination is sufficient for the reading of large print. Secondly, auroral displays are almost invariably associated with serious interference with radio communications, complete interruptions being common.

APPENDIX I

DERIVATION OF SOME FORMULAE USED IN THE TEXT

Characteristic gas equation

For a given mass of gas there is a fundamental relation between the pressure p, volume v and absolute temperature T of the form

$$pv = RT, \qquad \dots . \quad (1)$$

where R is a constant for any one gas. Since for unit mass the density ρ is the reciprocal of the volume, the equation is equivalent to

$$p = R\rho T, \qquad \dots . \quad (2)$$

which is the more convenient form for meteorological applications. The value of R for dry air is $2 \cdot 87 \times 10^{-3}$ when p is in millibars, ρ in grammes per cubic metre and T in degrees Kelvin. When C.G.S. units are used – pressure in dynes per square centimetre and density in grammes per cubic centimetre – then R is $2 \cdot 87 \times 10^6$. Equation (2) leads to expressions for the density of dry and of moist air, as shown in Section 20.

Pressure and height (Section 7)

A second fundamental equation in meteorology derives from the rule stated in Section 5 that the pressure at any level in the atmosphere is equal to the weight of air above that level in a column of unit cross-section. In Fig. 1, which represents

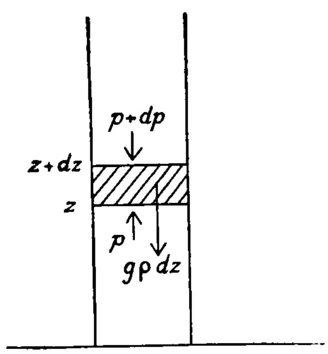


FIG. 1.

a column of air standing on the ground, consider a layer extending from height z where the pressure is p, to height $z + dz$ where the pressure is denoted by $p + dp$. The pressure on the upper surface of the layer is less than the pressure on the lower surface by the weight per unit area of the layer itself. The difference of pressure is $p - (p + dp)$ or $-dp$, while if ρ is the density of the layer, the mass per unit area is ρdz and the weight is $g\rho dz$, where g is the acceleration due to gravity. Therefore

$$g\rho dz = -dp.$$

Eliminating the density by means of equation (2), we obtain

$$dz = -\frac{RT}{gp}\,dp. \qquad \ldots . \quad (3)$$

This result shows how an increment of height dz is related to the corresponding increment of pressure dp, the negative sign indicating that pressure decreases with increase of height. Using the C.G.S. value of R and putting $g = 981$ centimetres per second per second, we obtain the thickness dz in centimetres; or on dividing by 30·5 we obtain the thickness in feet. Finally, for the thickness (in feet) of a layer extending over a pressure range of 1 millibar we must put $dp = 1$ and therefore

$$dz = \frac{2\cdot 87 \times 10^6}{981 \times 30\cdot 5} \times \frac{T}{p} = 96\frac{T}{p} \qquad \left[29.25\frac{T}{p}\text{ metres}\right]$$

where p is in millibars and T in degrees Kelvin.

The height interval between two pressure levels p_1 *and* p_2 is obtained by integration of equation (3). The result takes a simple form when the temperature is constant, thus

$$\int_{h_1}^{h_2} dz = -\frac{RT}{g}\int_{p_1}^{p_2}\frac{dp}{p}$$

whence $$h_2 - h_1 = \frac{RT}{g}(\log_e p_1 - \log_e p_2).$$

With the units used above, the value of R/g is again 96, while if the logarithms are changed from base e to base 10, the conversion factor 2·303 must be introduced. This gives

$$h_2 - h_1 = 221\cdot 1\ T(\log p_1 - \log p_2)\text{ feet}$$

$$[h_2 - h_1 = 67\cdot 4T(\log p_1 - \log p_2)\text{ metres}].$$

Adiabatic relations

An equation given in Section 14 states the relation between the changes in pressure and temperature of a mass of air which is thermally insulated from its surroundings. This result is derived from the first law of thermodynamics which is an expression of the conservation of energy and of the equivalence between energy and heat. It may be stated for our purposes as follows:

If an amount of heat is taken up by a gas, some is converted into internal energy and the remainder is used up in work done on the environment as the gas expands.

The internal energy refers to the kinetic energy of the molecules and is known to be proportional to the absolute temperature; for a given rise of temperature, the increase of the internal energy is therefore proportional to the change of temperature. If the rise of temperature is accompanied by expansion, then work is done in pushing back the environment so that some of the heat supplied is consumed in this way; or if the gas contracts, work is done on the gas by the environment and an additional amount of heat is made available. Thus for a given increase of temperature a definite amount of heat is taken up in changing the internal energy, while a variable amount (which may even be negative) is taken up in changing the volume. By the definition of specific heat we may write:

heat supplied = (specific heat) × (increase of temperature).

In the special case when the volume remains constant, all the heat goes to raising the temperature. The specific heat is then written as c_v to draw attention to the constancy of volume; it is called the specific heat at constant volume. Another special case arises if the pressure is kept constant; some heat is now consumed by the work done in expansion, so the specific heat has a value greater than c_v; it is denoted by c_p and is called the specific heat at constant pressure. Thus for a small rise dT in temperature, the heat dQ taken in should be equated to $c_v dT$ if the volume is constant, or to $c_p dT$ if the pressure is constant.

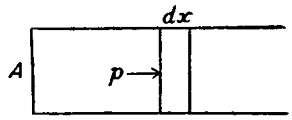


FIG. 2

Next consider the work done by the gas as it expands and pushes back the environment. If in a cylinder of cross-section A (Fig. 2) the expansion pushes a piston outwards through a small distance dx, the work done (force × distance) is $pA.dx$ or pdv, where dv is the increase in volume. Thus of an amount of heat dQ applied to the gas, an amount $c_v dT$ is accounted for by the increase of temperature and an amount pdv by the change in volume. The law of conservation is therefore expressed by the equation

$$dQ = c_v dT + pdv. \qquad \text{. . . . (4)}$$

From equation (1), by making a small variation we obtain

$$pdv + vdp = RdT. \qquad \text{. . . . (5)}$$

Therefore (4) becomes

$$dQ = (c_v + R)dT - vdp. \qquad \text{. . . . (6)}$$

Now an adiabatic change is by definition one in which no heat is supplied or removed; in this case dQ is zero and equation (6) gives

$$(c_v + R)dT = vdp = \frac{RT}{p} dp\,,$$

or

$$\frac{dT}{T} = \frac{R}{c_v + R} \frac{dp}{p}. \qquad \text{. . . . (7)}$$

Then by integration from pressure p_0 to p_1, corresponding to temperatures T_0 and T_1, we obtain the 'adiabatic' equation

$$\log T_0 - \log T_1 = \frac{R}{c_v + R}(\log p_0 - \log p_1). \qquad \ldots . \quad (8)$$

Again, from equation (6), if the pressure is for the moment assumed constant so that $dp = 0$, then dQ, as we have seen, may be replaced by $c_p dT$ and we have

$$c_p dT = (c_v + R)dT ,$$

or

$$R = c_p - c_v. \qquad \ldots . \quad (9)$$

The first factor on the right of (8) then becomes $(c_p - c_v)/c_p$, or if the ratio of the specific heats, c_p/c_v, is put equal to γ, the factor becomes $(\gamma-1)/\gamma$. The value of γ for dry air is $1{\cdot}402$ so that equation (8) becomes finally

$$\log T_0 - \log T_1 = 0{\cdot}287 (\log p_0 - \log p_1). \qquad \ldots . \quad (10)$$

Potential temperature

This has been defined in Section 14 as the temperature acquired by a parcel of air when its pressure is changed adiabatically to 1000 millibars. Let it be denoted by θ and let p_0 in equation (10) refer to 1000 millibars. Then

$$\log \theta = \log T + 0{\cdot}287 (\log p_0 - \log p), \qquad \ldots . \quad (11)$$

or

$$\theta = T\left(\frac{1000}{p}\right)^{0.287}$$

where T and p now denote the temperature and pressure of the air initially.

Dry adiabatic lapse rate (Section 14)

Equation (7) may be rewritten as

$$\frac{dT}{T} = \frac{\gamma-1}{\gamma}\,\frac{dp}{p}. \qquad \ldots . \quad (12)$$

It determines the change in temperature dT resulting from a small change in pressure dp when the process is carried out adiabatically. This result will now be applied to the adiabatic ascent of a parcel of dry air. On substituting for dp/p from equation (3) and dividing by dz, we find

$$\frac{dT}{dz} = -\frac{\gamma-1}{\gamma}\,\frac{g}{R}. \qquad \ldots . \quad (13)$$

The expression on the left is the rate of change of temperature with height in air ascending adiabatically, in other words it is the dry adiabatic lapse rate and its value as given on the right is seen to be constant. With the figures already used, this constant value becomes 0.287/96 in degrees Celsius per foot, or 3 degC (5.4 degF) per 1000 feet (9·8 degC per kilometre), the negative sign in (13) showing that the temperature decreases with increasing height.

A discrepancy in the above argument should be noted. Whereas equation (12) describes the conditions within an isolated parcel of air, equation (3) concerns the relationship between pressure and height in the environment. Thus in combining these equations in order to derive (13) it has been tacitly assumed that the conditions of the ascending parcel are the same as those of the environment at each level. This is true of the pressure but not in general of the temperature which, after a large displacement, may differ considerably from that of the environment at the same level. The formula for the adiabatic lapse rate is therefore strictly true only at the start of the displacement, but in practice its use does not seriously affect the computed temperature provided the change in height is less than about 10 000 feet. For changes greater than this, equation (10) should be used.

The derivation of an expression for the saturated adiabatic lapse rate is much more complicated and will not be given here.

Entropy (Section 15)

We have seen that if a small variation takes place in the state of a gas, the amount of heat taken up is given by equation (4). Now suppose the state of the gas changes by a finite amount. Since any two of the variables p, v and T are sufficient to define the state, we may represent the change on, for example, a p–v diagram (Fig. 3) by a continuous line from the point A representing the initial state (p_1, v_1) to the point B

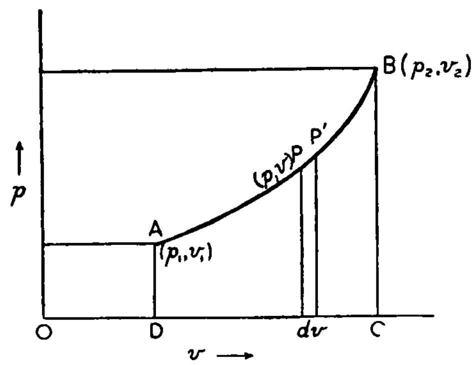


FIG. 3

representing the final state (p_2, v_2). Any intermediate point P on the curve AB represents the intermediate state (p, v) given by the co-ordinates of that point. In order to find the total amount of heat taken up by the gas in passing from the state A to the state B, the expression (4) must be summed for all stages represented by the path-curve AB. Thus the total amount of heat taken up is given by

$$Q_2 - Q_1 = c_v(T_2 - T_1) + \int_{v_1}^{v_2} p\,\mathrm{d}v. \qquad \ldots . \quad (14)$$

Now $p\mathrm{d}v$ is represented by the element of area in the column under PP′, so that the integral on the right of (14) is equal to the whole area (ABCD) enclosed between the curve AB and the v-axis; this area is clearly dependent on the shape of the curve AB as well as on the actual positions of A and B themselves. Hence the amount of heat which must be supplied when the gas changes from the state p_1v_1 to any other state p_2v_2 depends on the precise manner in which the transition is brought about at each stage. This property of the amount of heat taken up is very inconvenient for purposes of calculation – it would be more convenient to have a related quantity whose value depends only on the initial and final points of the transition. Actually it is easy to derive such a quantity. All we need do is to divide equation (4) throughout by the temperature before integrating. Thus

$$\frac{\mathrm{d}Q}{T} = c_v\frac{\mathrm{d}T}{T} + \frac{p\mathrm{d}v}{T},$$

which by use of equation (1) becomes

$$\frac{\mathrm{d}Q}{T} = c_v\frac{\mathrm{d}T}{T} + \frac{R\mathrm{d}v}{v},$$

whence $$\int_{Q_1}^{Q_2} \frac{\mathrm{d}Q}{T} = c_v \log\frac{T_2}{T_1} + R \log\frac{v_2}{v_1}. \qquad \text{. . . . (15)}$$

Hence the result of integrating $\mathrm{d}Q/T$ is to obtain an expression which depends only on the initial and final states of the gas and not at all on the intermediate stages. The expression $\mathrm{d}Q/T$ is written as $\mathrm{d}S$, where S is termed the entropy. Thus from equation (15) the general expression for the entropy of unit mass of gas is obtained as

$$S = c_v \log T + R \log v + \text{constant}, \qquad \text{. . . . (16)}$$

where the constant depends on the temperature and volume in a state of zero entropy.

Thus the change of entropy, which may be defined as the sum of successive small heat increments each divided by the corresponding absolute temperature, possesses the required property of depending only on the state of the gas and not on how that state was reached. Further, if a process takes place adiabatically, $dQ = 0$; it follows then that $dS = 0$ so that the entropy is constant and the process is said to be isentropic. Consequently the dry adiabatic lines are lines of equal entropy and are therefore represented as a set of parallel straight lines on the tephigram.

Another expression for the entropy of dry air can be given in terms of potential temperature. From equation (11) we obtain

$$\log \theta = \log T - \frac{\gamma - 1}{\gamma} \log p + \text{constant}.$$

By use of equation (1) and by putting $\gamma = c_p/c_v$ this becomes

$$\log \theta = \log T - \frac{c_p - c_v}{c_p} \log\frac{T}{v} + \text{constant},$$

or $$c_p \log \theta = c_v \log T + R \log v + \text{constant}.$$

Therefore

$$S = c_p \log \theta,$$

apart from an additive constant. Thus the entropy is proportional to the logarithm of the potential temperature. The dry adiabatics on the tephigram (Appendix II) are shown for every 10 degC of potential temperature, the position of any line being determined by its entropy, which is on a linear scale.

Humidity mixing ratio and dew-point (Section 31)

Consider a volume v containing 1 gramme of dry air and r grammes of water vapour. If the total pressure is p and the vapour pressure e, then the gas equation for the dry air alone is

$$(p-e)v = RT$$

and for the water vapour is

$$ev = rR_wT = \frac{8}{5}rRT.$$

Hence the humidity mixing ratio in grammes vapour per gramme dry air is given by

$$r = \frac{5e}{8(p-e)}. \quad \ldots\ldots (17)$$

For saturated air, e becomes the saturation vapour pressure e_s. Since the value of e_s is known experimentally for any temperature, it follows that the saturation mixing ratio at any pressure and temperature is obtainable from equation (17) when e is replaced by e_s. Further, as the humidity mixing ratio is the saturation mixing ratio at the dew-point temperature, the dew-point temperature is partly dependent on pressure. In unsaturated adiabatic ascent, the humidity mixing ratio remains constant while both temperature and dew-point decrease, as may be verified from the tephigram.

Coriolis force (Section 24)

In deriving the formula for the geostrophic wind the direction and magnitude of the Coriolis force was assumed. To prove this result it is necessary first to consider how the rotation of the earth affects any part of its surface. The earth rotates about its polar axis from west to east with angular velocity denoted by Ω. The linear velocity of a point P in latitude ϕ (Fig. 4) is therefore $\Omega R\cos\phi$, and the velocity at

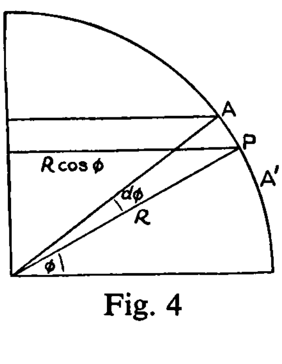


Fig. 4

a near point A in latitude $\phi + d\phi$ is $\Omega R\cos\phi + d(\Omega R\cos\phi)$. Thus the velocity at A relative to that at P is $\Omega Rd(\cos\phi)$ or $-\Omega R\sin\phi . d\phi$ and similarly the relative velocity at A′ in latitude $\phi - d\phi$ is $+\Omega R \sin \phi . d\phi$. Since the distance from P to A or A′ along the earth's surface is $R d\phi$, it follows that the surface in the neighbourhood of P rotates counter-clockwise about the vertical at P with angular velocity $\Omega \sin \phi$.

Let us now consider air moving horizontally over the earth in north latitude ϕ with velocity V in a direction OA which is fixed in space (Fig. 5); in a small

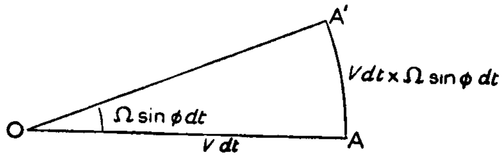


FIG. 5

time dt the air travels a distance OA equal to Vdt. Meanwhile the rotation of the tangent plane to the earth's surface at O with angular velocity $\Omega \sin \phi$ carries the point A to A′, where the angle AOA′ is $\Omega \sin \phi . dt$. While moving along OA in space, the air therefore travels a distance represented by A′A (in that direction) across the earth's surface; this distance A′A equals $Vdt \times \Omega \sin \phi . dt$ or $\Omega \sin \phi . V (dt)^2$. Since the motion at O is directed along OA, the transverse motion may be regarded as accelerated from rest. If we denote this acceleration by f, the usual formula gives

$$A'A = \tfrac{1}{2} f (dt)^2$$

whence

$$f = 2\Omega V \sin \phi .$$

This is the Coriolis acceleration; its direction is seen to be perpendicular to and to the right of the wind velocity. In the southern hemisphere the corresponding latitude is to be taken as negative so that there the direction of the rotation and of the Coriolis acceleration are reversed.

Determination of displacement, Z_n, from altimeter observations (Section 30).

Consider an aircraft flying on a constant heading along an isobaric surface (i.e. at a fixed height as indicated by the pressure altimeter) and maintaining constant airspeed; then consider the progress of the aircraft over a short section of its track PQ (Fig. 6), such that the wind over it may be regarded as constant.

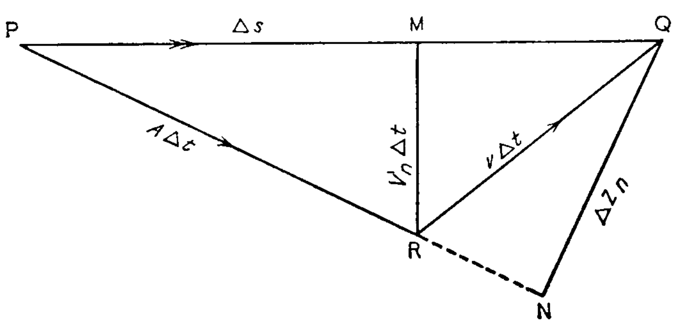


FIG. 6

A = true airspeed of aircraft $\quad v$ = wind speed
Δs = length of PQ $\quad \Delta t$ = time taken to traverse PQ
H_P, H_Q = height at P and Q respectively, of the isobaric surface on which the flight is made. $\Delta H = H_Q - H_P$
V'_n = component of wind perpendicular to ground track
ΔZ_n = component of displacement, due to wind in time Δt, perpendicular to the aircraft heading PR

In Fig. 6, PR ($=A\ \Delta t$) represents the displacement which the aircraft would have had in still air in the interval Δt; RQ ($=v\ \Delta t$) represents the displacement of the aircraft due to wind in the same period; the result is a net displacement PQ ($=\Delta s$); MR ($=V'_n\ \Delta t$) represents the displacement perpendicular to the aircraft's ground track.

The triangles PNQ and PMR are similar, and thus QN/PQ = RM/PR or

$$\frac{\Delta Z_n}{\Delta s}=\frac{V'_n\ \Delta t}{A\ \Delta t}; \qquad \ldots . \ (18)$$

then, if the wind satisfies the geostrophic relation,

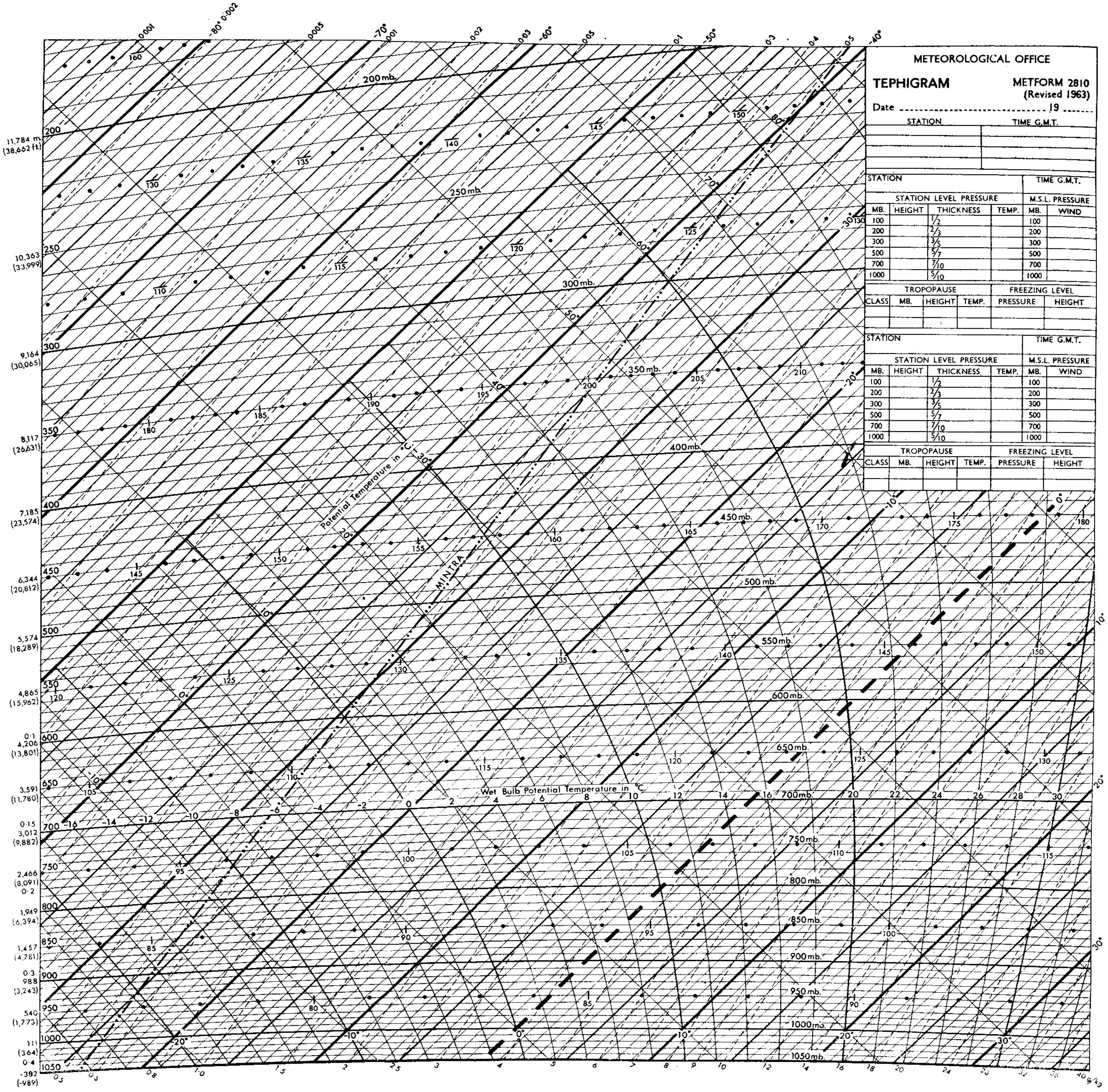
$$V'_n=\frac{g}{2\Omega\sin\phi}\frac{\Delta H}{\Delta S}. \qquad \ldots . \ (19)$$

Combining (18) and (19)

$$\Delta Z_n=\frac{g}{2\Omega\sin\phi}\frac{\Delta H}{A}.$$

Then, if the airspeed, A, the latitude, ϕ, and the heading of the aircraft are regarded as constant over a flight path from P to Q

$$Z_n=\frac{g}{2\Omega\sin\phi}\frac{H_Q-H_P}{A}.$$


METEOROLOGICAL OFFICE
TEPHIGRAM
METFORM 2810
(Revised 1963)
Date 19
STATION
TIME G.M.T.
STATION LEVEL PRESSURE
M.S.L. PRESSURE
MB.
HEIGHT
THICKNESS
TEMP.
WIND
TROPOPAUSE
FREEZING LEVEL
CLASS
PRESSURE
200 mb.
250 mb.
300 mb.
350 mb.
400 mb.
450 mb.
500 mb.
550 mb.
600 mb.
650 mb.
700 mb.
750 mb.
800 mb.
850 mb.
900 mb.
950 mb.
1000 mb.
1050 mb.
Potential Temperature in °C
Wet Bulb Potential Temperature in °C
MINTRA
11,784 m.
(38,662 ft.)
10,363
(33,999)
9,164
(30,065)
8,117
(26,631)
7,185
(23,574)
6,344
(20,812)
5,574
(18,289)
4,865
(15,962)
4,206
(13,801)
3,591
(11,780)
3,012
(9,882)
2,466
(8,091)
1,949
(6,394)
1,457
(4,781)
988
(3,243)
540
(1,773)
111
(364)
-392
(-989)

APPENDIX III

BOOKS FOR FURTHER READING

The following list of publications, all of which are readily available, is intended as a guide for readers who wish to pursue their studies beyond the limits covered by the handbook.

Chapter 5

SAWYER, J. S.; Theoretical aspects of pressure-pattern flying. *Met. Rep., London,* No. 3, 1949 (Reprint 1956).

GRAYSTONE, P.; Equivalent headwinds at heights of 30,000 feet and 40,000 feet along air routes. *Met. Rep., London,* No. 20, 1958.

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AANENSEN, C. J. M. (Editor); Gales in Yorkshire in February 1962. *Geophys. Mem., London,* **14**, No. 108, 1965.

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ANDERSON, R. K., FERGUSON, E. W. and OLIVER, V. J.; The use of satellite pictures in weather analysis and forecasting. *Tech. Notes, Wld. met., Org., Geneva*, No. 75, 1966.

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Part III. Coding, decoding and plotting. Fourth edition, 1963.

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Chapters 19 *and* 20

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PETTERSSEN, S.; Weather analysis and forecasting. Second edition. Vol. I Motion and motion systems. Vol. II Weather and weather systems. New York and London, McGraw-Hill, 1956.

Chapter 21

London, Board of Trade, Civil Aviation Department. UK Air Pilot. London, HMSO, 1948 (with amendments).

Montreal, International Civil Aviation Organization. Procedures for air-navigation services—Meteorology. Fourth edition. (Doc. 7605—MET/526/4). Montreal, ICAO, 1966.

Chapter 22

HEASTIE, H. and STEPHENSON, P. M.; Upper winds over the world. Parts I and II. *Geophys. Mem., London*, **13**, No. 103, 1960.

FROST, R. and STEPHENSON, P. M.; Mean streamlines and isotachs at standard levels over the Indian and West Pacific Oceans and adjacent land areas. *Geophys. Mem., London*, **14**, No. 109, 1965.

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The challenge of the atmosphere. London, Hutchinson, 1962.
SUTCLIFFE, R. C.; Weather and climate. London, Wiedenfeld and Nicolson, 1966.

APPENDIX IV

CONVERSION TABLES

Table I—Conversion of degrees Celsius to Fahrenheit

°C	·0	·2	·5	·8	°C	·0	·2	·5	·8
		degrees Fahrenheit					*degrees Fahrenheit*		
40	104	104	105	105	0	32	32	33	33
					− 0	32	32	31	31
39	102	103	103	104	− 1	30	30	29	29
38	100	101	101	102	− 2	28	28	27	27
37	99	99	99	100	− 3	27	26	26	25
36	97	97	98	98	− 4	25	24	24	23
35	95	95	96	96	− 5	23	23	22	22
34	93	94	94	95	− 6	21	21	20	20
33	91	92	92	93	− 7	19	19	19	18
32	90	90	91	91	− 8	18	17	17	16
31	88	88	89	89	− 9	16	15	15	14
30	86	86	87	87	−10	14	14	13	13
29	84	85	85	86	−11	12	12	11	11
28	82	83	83	84	−12	10	10	9	9
27	81	81	81	82	−13	9	8	8	7
26	79	79	80	80	−14	7	6	6	5
25	77	77	78	78	−15	5	5	4	4
24	75	75	76	77	−16	3	3	2	2
23	73	74	74	75	−17	1	1	1	0
22	72	72	73	73	−18	0	− 1	− 1	− 2
21	70	70	71	71	−19	− 2	− 3	− 3	− 4
20	68	68	69	69	−20	− 4	− 4	− 5	− 5
19	66	67	67	68	−21	− 6	− 6	− 7	− 7
18	64	65	65	66	−22	− 8	− 8	− 9	− 9
17	63	63	63	64	−23	− 9	−10	−10	−11
16	61	61	62	62	−24	−11	−12	−12	−13
15	59	59	60	60	−25	−13	−13	−14	−14
14	57	58	58	59	−26	−15	−15	−16	−16
13	55	56	56	57	−27	−17	−17	−17	−18
12	54	54	55	55	−28	−18	−19	−19	−20
11	52	52	53	53	−29	−20	−21	−21	−22
10	50	50	51	51	−30	−22	−22	−23	−23
9	48	49	49	50	−31	−24	−24	−25	−25
8	46	47	47	48	−32	−26	−26	−27	−27
7	45	45	45	46	−33	−27	−28	−28	−29
6	43	43	44	44	−34	−29	−30	−30	−31
5	41	41	42	42	−35	−31	−31	−32	−32
4	39	40	40	41	−36	−33	−33	−34	−34
3	37	38	38	39	−37	−35	−35	−35	−36
2	36	36	37	37	−38	−36	−37	−37	−38
1	34	34	35	35	−39	−38	−39	−39	−40

TABLE II—CONVERSION OF KNOTS TO METRES PER SECOND

1 knot = 0·51479 metres per second

Knots	0	1	2	3	4	5	6	7	8	9
					metres per second					
0	0·0	0·5	1·0	1·5	2·1	2·6	3·1	3·6	4·1	4·6
10	5·1	5·7	6·2	6·7	7·2	7·7	8·2	8·8	9·3	9·8
20	10·3	10·8	11·3	11·8	12·4	12·9	13·4	13·9	14·4	14·9
30	15·4	16·0	16·5	17·0	17·5	18·0	18·5	19·0	19·6	20·1
40	20·6	21·1	21·6	22·1	22·7	23·2	23·7	24·2	24·7	25·2
50	25·7	26·3	26·8	27·3	27·8	28·3	28·8	29·3	29·9	30·4
60	30·9	31·4	31·9	32·4	32·9	33·5	34·0	34·5	35·0	35·5
70	36·0	36·6	37·1	37·6	38·1	38·6	39·1	39·6	40·2	40·7
80	41·2	41·7	42·2	42·7	43·2	43·8	44·3	44·8	45·3	45·8
90	46·3	46·8	47·4	47·9	48·4	48·9	49·4	49·9	50·4	51·0
100	51·5	52·0	52·5	53·0	53·5	54·1	54·6	55·1	55·6	56·1

TABLE III—CONVERSION OF INCHES TO MILLIMETRES

1 inch = 25·4 millimeters

Inches	·0	·1	·2	·3	·4	·5	·6	·7	·8	·9
					millimetres					
0·0	0	3	5	8	10	13	15	18	20	23
1·0	25	28	30	33	36	38	41	43	46	48
2·0	51	53	56	58	61	64	66	69	71	74
3·0	76	79	81	84	86	89	91	94	97	99
4·0	102	104	107	109	112	114	117	119	122	124
5·0	127	130	132	135	137	140	142	145	147	150
6·0	152	155	157	160	163	165	168	170	173	175
7·0	178	180	183	185	188	191	193	196	198	201
8·0	203	206	208	211	213	216	218	221	224	226
9·0	229	231	234	236	239	241	244	246	249	251
10·0	254	257	259	262	264	267	269	272	274	277

TABLE IV—CONVERSION OF FEET TO METRES

1 foot = 0·3048 metres

1000's of feet	0	1	2	3	4	5	6	7	8	9
					metres					
0	0	305	610	915	1219	1524	1829	2134	2438	2743
10	3048	3352	3657	3962	4267	4572	4877	5182	5486	5791
20	6096	6401	6706	7011	7351	7620	7924	8229	8534	8839
					kilometres					
30	9·1	9·5	9·8	10·1	10·4	10·7	11·0	11·3	11·6	11·9
40	12·2	12·5	12·8	13·1	13·4	13·7	14·0	14·3	14·6	14·9
50	15·2	15·5	15·9	16·2	16·5	16·8	17·1	17·4	17·7	18·0
60	18·3	18·6	18·9	19·2	19·5	19·8	20·1	20·4	20·7	21·0
70	21·3	21·6	22·0	22·3	22·6	22·9	23·2	23·5	23·8	24·1
80	24·4	24·7	25·0	25·3	25·6	25·9	26·2	26·5	26·8	27·1
90	27·4	27·7	28·0	28·4	28·7	29·0	29·3	29·6	29·9	30·2
100	30·5									

Note: Metres are rounded off to the nearest whole metre; kilometres are rounded off to the nearest tenth of a kilometre; 500 feet = 152 metres.

INDEX

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