METEOROLOGY

by

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PREFACE

THIS book aims at giving a brief sketch of the physical principles underlying the phenomena which constitute 'weather', in so far as this is possible without mathematical analysis. The reader will find in it no attempt at laying down rules for forecasting the weather. This omission is intentional. The author is of opinion that the adequate discussion of the principles and methods of forecasting would double the size of the book. Any short treatment of so difficult a question would give a misleading appearance of definiteness to this aspect of meteorology.

A brief résumé of the historical development of meteorology has been given in Chapter I, largely with a view to showing that the scientific treatment of the subject is of recent growth. Throughout the book an effort has been made to deal with the various problems in as simple a manner as possible, but the author is aware that this cannot always be successful. Meteorology makes use of most branches of physical science, and certain portions of the subject, involving technical concepts, cannot be discussed in a non-technical manner without intolerable circumlocution.

I have to acknowledge my indebtedness to Mr. G. A. Clarke, Aberdeen, for the four cloud photographs reproduced in Figures 8, 9, 15, and 17; to the Royal Meteorological Society for permission to reproduce the lightning photographs in Figures 1 and 19; to the Controller of H.M. Stationery Office for permission to reproduce the Daily Weather Reports in Figures 12 and 13; and to the Director of the Meteorological Office for permission to reproduce from official records the charts in Figures 2, 3, and 14.

I am much indebted to Major A. R. Low, Librarian, Air Ministry, and to my colleagues Captain L. G. Garbett, R.N., and Mr. H. L. B. Tarrant, who have read the book in manuscript or proof, and have helped to remove a number of obscurities.

D. B.

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I

Mainly Historical

THE beginnings of the study of the weather must have come at a very early stage of what we call civilization. When man began to have a conscious knowledge of the course of the seasons, he had already attained the first milepost on the long road to perfect knowledge of the ways of the atmosphere, a road whose end is not yet in sight. The earliest writings contain only the most fragmentary references to weather until we come down to the time of Aristotle in the fourth century B.C. Aristotle wrote a treatise entitled *Meteorologica* in which he discussed at length his own views of the causes of the weather, and in which he incidentally showed that a considerable body of knowledge had already been accumulated concerning the relations of weather and wind. He was followed by his pupil Theophrastus, who wrote treatises on the winds and on weather signs. These weather signs were much like those which we still frequently hear to-day in England, and they appear to have attained general acceptance throughout Europe during a period of close on 2,000 years, for we find that from the time of Theophrastus down to the late seventeenth century, weather science practically stood still. Writings on the subject were limited to translations of and commentaries upon the works of Aristotle and Theophrastus, while the weather signs of Theophrastus were multiplied and occasionally inverted, so that the same weather sign was interpreted in diametrically opposite senses in different parts of the same country. No appreciable progress was made upon the lines laid down by Aristotle, because his physical ideas were insufficient to account for the observed facts, and the true constitution of the atmosphere was not understood until many centuries later.

Nevertheless, the accumulation of the facts of the weather proceeded through the centuries. The invention of the barometer in 1643 by an Italian named Torricelli led to no immediate advance in Meteorology. The wheel barometer invented by Hooke in 1670 was provided with a dial indicating the pressure in inches of mercury, and the association of weather with height of barometer led to the engraving on the dial of the barometer of the word Change at 29·5 inches, Rain at 29·0 inches, Much Rain at 28·5 inches, Stormy at 28·0 inches, and on the other side, Fair at 30·0 inches, Set Fair at 30·5 inches, and Very Dry at 31·0 inches. The barometer thus inscribed was known as a weather glass. Its inadequacy [1] as a means of weather forecasting was recognized immediately by scientific men, and attempts were made to formulate rules in which the changes of pressure were associated with the changes of wind. Fitzroy in England gave forty-seven rules for use with the barometer. None of these attempts was particularly fruitful, simply because a given height of barometer at one station may be associated with widely differing distributions of pressure in the surrounding area.

The first serious attempt to break new ground came in 1686 when Halley (whose name is more generally associated with a comet) put forward a theory to account for the trade winds and monsoons based on the variation of the heating effect of solar radiation from equator to pole, a theory which has held its ground up to the present day. Then in 1735 Hadley indicated the effect of the rotation of the earth upon the trade winds, and showed that these winds must acquire a westward component of motion purely on account of the rotation of the earth.

[1] An example may make this inadequacy clear. On 7th February 1927, with the barometer at 30·47 inches, or 'Set Fair', about one-third of an inch of rain fell in London during the day.

Then came an interval of about a hundred years when no appreciable advance was made. The next important step was made by Brandes, who in 1820 laid down the relation between the wind and pressure distribution, and showed that depressions tended to move from west to east. Brandes discussed in detail a number of well-marked depressions and illustrated his discussions by synoptic charts, i. e. charts on which were plotted simultaneous observations of pressure, wind, temperature, &c.

An American meteorologist, Espy, published in 1841 a valuable contribution to the meteorology of his day in a book entitled *The Philosophy of Storms*, in which he laid down the view that in a depression the air tended to move towards the central region of lowest pressure, which he described as a region of ascending currents. Espy was in fact responsible for the definite formulation of what is to-day called the 'convection' theory of the origin of depressions, and he was the first to state definitely the importance of water-vapour in the atmosphere.

The middle of the nineteenth century was a period of very great activity in meteorological study. Many writers discussed in detail certain particular storms, and formulated general rules as to their motion, and we may say that by 1851 the nature of the circular storm or depression was understood as clearly as it was up to a few years ago. By this time the value of simultaneous observations of pressure, wind, temperature, and weather at a number of places was clearly appreciated, and the time was ripe for a new step forward.

The relation of the wind to the pressure distribution was explicitly stated in 1857 by a Dutch meteorologist, Buys-Ballot, in a form which has come to be known as 'Buys-Ballot's law'. This law states that if in the northern hemisphere you stand with your back to the wind, you will have lower pressure to your left than to your right. In the southern hemisphere the reverse will hold. A glance at any weather chart will readily convince the reader of the truth of this law. It will be found that in general the wind tends to blow around the isobars, or lines of equal pressure, in the direction laid down by Buys-Ballot's law. In the northern hemisphere the winds circulate around centres of low pressure in a counter-clockwise sense, and around centres of high pressure in a clockwise sense. In the southern hemisphere the two cases are reversed, winds blowing clockwise round centres of low pressure and counter-clockwise round centres of high pressure.

If the reader will devote a few minutes to an examination of any of the charts given in later chapters of this book, he will readily recognize the truth of what we have stated above as to the relation between the wind and pressure distribution. Buys-Ballot's law was universally accepted, so much so that we may say it is the one meteorological law which no one questions.

A great impetus was given to marine meteorology by Lieutenant (afterwards Admiral) M. F. Maury of the U.S. Navy, who in 1839 met with an accident which rendered him unfit for further service afloat. He then devoted himself to the collection of observations of winds and currents, with a view to producing charts of the winds and currents of the globe. He soon concluded that international co-operation was essential for the fulfilment of his aims, and he moved the United States Government to invite all the maritime nations of the world to a conference at Brussels in 1853. As a result of this conference a plan of observations to be carried out at sea was adopted, and immediately put into operation by nearly all the maritime nations of the world. Maury's earliest charts led to one of the most striking practical triumphs of meteorology. Up to that time the passage from England to Australia had occupied on an average 124 days, and the round trip about 250 days. Ships sailing via the Cape had rounded the coast of West Africa and followed a fairly direct line to the Cape. This involved beating southward against the south-east Trades, with the expenditure of a very long time after crossing the line. Maury showed that a quicker passage to the Cape would be attained by crossing the equator at a more westerly point, passing close to the coast of South America, and rounding the western edge of the high-pressure belt of the southern Atlantic (see Figure 4). The adoption of Maury's suggestions led to the reduction of the average time for the passage to Australia from 124 days to 93 days. The return journey was achieved in 63 days under favourable conditions.

About the middle of the nineteenth century proposals for the establishment of a 'network' (*réseau*) of meteorological stations, all reporting their weather conditions at certain fixed times to a central service, were put forward in several countries. We need not here go into the details [1] of the order of priority of the various countries concerned. The progress made in the establishment of stations was rapid, the French *réseau* being complete by 1856. The Meteorological Office in London was established in 1854 as a department of the Board of Trade, with Admiral Fitzroy in charge. In 1860 Fitzroy began the collection of daily reports by telegraph, and soon after this began to issue to the daily newspapers forecasts based on his charts. Fitzroy's forecasts were received with favour by the public, but met with very severe criticism from scientific circles in England. He further instituted a system of Storm Signals and weather warnings in February 1861. By this time it was definitely realized

[1] Full details will be found in *Manual of Meteorology*, vol. i, by Sir Napier Shaw (Cambridge University Press).

that synoptic meteorology could only progress by international co-operation, and arrangements had been made to communicate observations from a number of continental stations via Paris to London, and reciprocally to communicate daily to Paris the observations made at the fifteen stations then established in the United Kingdom. International co-operation was put on a sound basis as the result of an international meeting of meteorologists in Leipzig in 1872, and an international congress of meteorologists held in Vienna in 1873, which was attended by official delegates of a large number of countries. The area covered by the British Daily Weather Reports then became approximately what it is to-day. In recent years, however, the exchange of meteorological information has been made to an ever-increasing extent by wireless telegraphy, and now the different central meteorological offices issue the data collected by them in the form of collective code messages sent out at pre-arranged times, and on certain definite wave-lengths, so that all persons or services interested can receive them at the same time.

So far we have only been concerned with the development and use of observations made at the earth's surface. Sporadic efforts were made from the middle of the eighteenth century onward to obtain measurements of temperature or wind in the upper air, and balloon ascents were at various times made by a number of enthusiasts with a view to obtaining upper air observations, beginning with the ascents of Jeffries and Blanchard in 1784. Kites were also used for observations, continuously recording instruments being attached for the first time in 1894. Small free balloons carrying recording instruments were first used in 1893, and the pilot balloon, a small free balloon whose movements are observed by means of a theodolite, was first used in 1909 for the measurement of winds in the upper air. The latter method had been widely adopted before the war of 1914–18. During the war, the need for accurate observations of upper winds was urgent in the artillery and aviation services, and the use of the pilot balloon became firmly established. The development of aviation led to the use of aeroplanes for observations of upper air temperatures, and this method has now almost entirely supplanted all others in meeting the daily requirements of practical meteorology.

The pilot balloon is now the standard method of observation of upper winds over the land. At sea the method has only recently been applied with success, and several British ships are now equipped with all the apparatus necessary for pilot balloon ascents. During the relatively short time that pilot balloons have been in regular use, a large amount of information has been accumulated concerning the winds in the upper air over the continents, and the extension of this method to observations over the ocean may be expected to yield results of considerable value. Such observations are urgently required in connexion with the possible development of air routes for both aeroplane and airship over the oceans, and the meteorologist is to-day required to do for the upper air what Maury did for the surface winds over the oceans.

II

The Atmosphere: its Constitution and some of its Physical Properties

THE atmosphere in which we live forms a thin shell completely covering the whole earth. The highest clouds we see in middle latitudes are seldom more than six miles above the earth's surface, and three-quarters of the total weight of air is within about seven miles of the earth's surface.

Air is a mixture of gases, of which the most important constituents are Nitrogen and Oxygen, which between them account for 99 per cent. of the volume of a sample of 'dry air'; the remaining 1 per cent. being almost entirely accounted for by Argon, one of the inert gases discovered in the atmosphere some thirty years ago. There are also traces of other gases, Carbon Dioxide, Ozone, Hydrogen, Neon, Helium, Krypton, and Xenon, but they occur in such relatively minute quantities that they are of no practical importance in the study of weather.

The list of constituents of air given above does not include water-vapour. This is the one seriously variable constituent of the atmosphere. Apart from the pollution due to factories or other agencies, air from which water-vapour has been removed is always made up of the same gases, mixed together in the same proportions. Water-vapour alone refuses to obey this rule. Its amount is subject to wide variation with both time and place, being increased by evaporation and decreased by precipitation as rain or condensation as dew. The maximum amount of water-vapour which air can take up, or in other words the amount of water-vapour which will saturate air, depends only on the temperature of the air and is independent of its pressure; the higher the temperature of the air, the more water-vapour it can hold. And so we may anticipate that changes of temperature will in general be accompanied by changes in the amount of water-vapour in the air.

Under the temperature conditions which hold for the British Isles in summer, the greatest amount of water-vapour which the atmosphere could contain if it were saturated at all levels would only yield 35 millimetres (about 1·4 inches) of rain if it were all precipitated. In winter the corresponding figure is 15 millimetres (about 0·6 inches). The actual amount of water-vapour in the atmosphere at any time usually falls well below these limits, since air is seldom saturated with water-vapour through more than a fraction of its depth.

It is scarcely possible to over-estimate the importance of water-vapour in the atmosphere, quite apart from the possibility of its being precipitated upon our heads as rain. If all the water in the atmosphere were removed, the whole circulation of the atmosphere would be slowed down very considerably, and possibly its very nature would change. We shall indeed find that much of the study of the weather is a study of the vagaries of water-vapour, whose ever-changing amount causes the medium in which the weather develops to be ever changing its composition and physical properties.

The importance of the amount of water-vapour in the atmosphere makes it necessary to have a standard method of measuring it. A specimen of air which contains all the water-vapour it can hold is said to be 'saturated'. When air is unsaturated the amount of water-vapour it contains is expressed as a percentage of the amount it contains when saturated at the same temperature. This percentage is called the 'Relative Humidity'. The usual method of deriving relative humidity is to measure not only the ordinary temperature of the air, by means of a 'dry bulb' thermometer, but also the temperature of a thermometer covered with wet muslin, which is kept moist by means of an attached wick dipping into a small vessel of water. Such a thermometer is called a 'wet bulb' thermometer. The relative humidity is readily derived from the readings of these two thermometers by the use of humidity tables. Suppose, for example, the dry bulb thermometer reading is 64° F., and the wet bulb thermometer reading is 59° F., or 5° F. lower than the dry bulb reading, then the relative humidity is found from the tables to be 73 per cent.

The use of the wet bulb thermometer is based on the fact that the air passing the wet bulb evaporates some of the water, and so lowers the temperature of the bulb. The drier the air, the more evaporation can take place, and the greater will be the difference between the dry bulb and wet bulb readings. Thus when the air is relatively dry, or the relative humidity low, the difference between the dry bulb and wet bulb readings is large. Conversely, when the air is damp or the relative humidity high, the amount of evaporation from the wet bulb is slight, and the difference between the dry bulb and wet bulb readings is small.

When air is cooled until the amount of water-vapour it contains is just sufficient to saturate it, the temperature it has then attained is called the 'dew point'. If it is cooled below this temperature, some of the water-vapour must condense into water-drops.

One important consequence of the introduction of water-vapour into air previously dry is to reduce its density. Water-vapour is lighter than air, volume for volume, in the proportion of 5 to 8, and when water is evaporated into free air, the water-vapour is not simply added to the dry air within a given volume, but displaces an equal volume of air. Thus the weight of a cubic foot of air containing water-vapour is less than the weight of an equal volume of dry air at the same pressure.

The weight of a given volume of air is small by comparison with the weight of an equal volume of water. At standard temperature and pressure (60° Fahrenheit and 30 inches respectively), a cubic foot of air weighs about 1·2 oz. Or, using the units which are now in more general use among meteorologists, at 0° Centigrade (32° F.) and a pressure of 1,000 millibars (see p. 17), a cubic metre of air weighs 1·28 kilogrammes. The weight of air contained in a column extending from the ground to the top of the atmosphere is, under standard conditions, about 14·5 lb. for every square inch of ground considered. This is what is frequently referred to as a pressure of 'one atmosphere'. It is the weight of such a column which is measured when we measure pressure by means of the mercury barometer. The latter is simply a device for balancing the weight of a column of air extending from the ground to the top of the atmosphere (wherever that may be) against the weight of a column of mercury of equal area of cross-section. If the reading of a mercury barometer is 29·5 inches, then the weight of air above one square inch of ground is equal to the weight of 29·5 cubic inches of mercury. This simple interpretation of pressure is frequently overlooked.

When pressure falls, or, as we normally say, when the barometer falls at a given place, the amount of air above that place is becoming less, and it can only become less by the removal of air. The question of how air is removed from any one place, and where it is removed to, is one of the most important in Meteorology, and its complete answer, if and when it is forthcoming, will be the answer to the vexed question, among others, of the origin of the depressions which are responsible for the uncertain weather of the British Isles.

Like all other gases, air is compressible, and is denser near the ground than it is at some distance above the ground. The density, which we define as the weight of a unit volume of air, decreases with height. The result of this is that, whereas one-half of the total mass of the atmosphere is contained within a layer 18,000 feet thick, the next layer above this of the same thickness contains rather less than a quarter of the total mass. The remaining quarter is spread up through a very considerable height. At the height of the top of Mount Everest the density is about two-fifths of the density at sea-level, and the difficulties of the ascent of the mountain are mainly due to the attenuated atmosphere.

III
The Standard Meteorological Observations and their Use

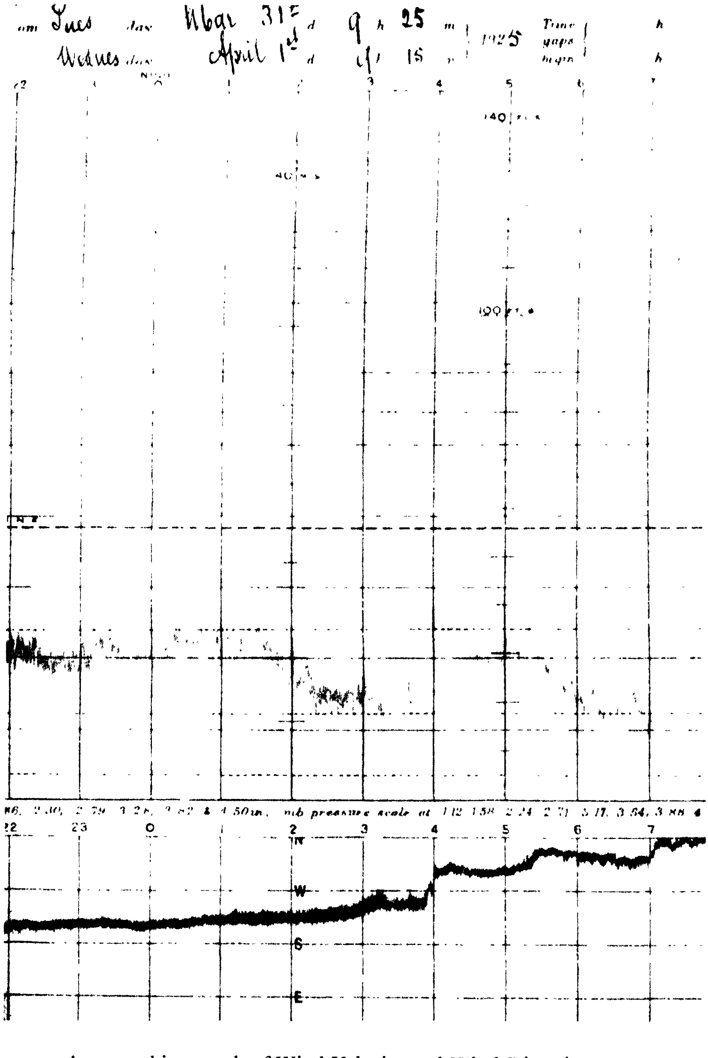
WE may define the aim of Meteorology as the study of the physical processes at work in the atmosphere. One of the consequences of this study is the formulation of certain laws which atmospheric phenomena will follow, and from these laws it is frequently possible to forecast the future course of the weather for some time to come. But the business of forecasting is not the only aim, or even the chief aim, of Meteorology.

The first stage in the study of the atmosphere is the observation of the facts of the weather at certain times each day. The observations which are usually made are those of pressure, temperature, humidity, wind direction and speed, weather and state of the sky, visibility, and amount of rain which has fallen since the last observation. We shall consider these briefly in turn. The reader who desires fuller information is referred to *The Meteorological Observer's Handbook* prepared by the Meteorological Office.[1]

(*a*) *Pressure*.

The pressure should preferably be read by means of a mercury barometer. The reading of the barometer is essentially the measurement of the height of the column of mercury whose weight will balance the weight of a column of air extending from the ground to the top of the atmosphere. The crude reading is corrected for the expansion of the metal tube and of the mercury in the barometer, and allowance is made for the height of the observing station above mean sea-level.

[1] Published by H.M. Stationery Office.



2. Autographic records of Wind Velocity and Wind Direction at Spurn Head 31 March–1 April 1925

Up to about 1914 pressure was commonly measured in inches of mercury, the average pressure being about 30 inches. Since that time, however, a new unit, called the millibar, has come into general use among professional meteorologists. Pressure is a force, and it was considered appropriate to measure pressure in terms of a unit of force, rather than in terms of a length such as an inch. The conversion from one set of units to the other is readily made by the use of the following relationship: 1 mercury inch equals 33·86 millibars; 30 mercury inches equal 1015·9 millibars.

(b) *Temperature and Humidity.*

The temperature is obtained from an ordinary thermometer exposed in a wooden screen whose sides are louvred, to provide for efficient ventilation of the instruments within it. In the same screen is hung a thermometer whose bulb is covered with muslin to which is attached a wick dipping into a small vessel containing water. The difference in the readings of these two thermometers, which are termed the 'dry bulb' and 'wet bulb' respectively, gives an indication of the humidity of the air, and the relative humidity can be obtained directly from the hygrometric or humidity tables.

Thermometers are usually graduated on either the Fahrenheit or the Centigrade scale. On the Fahrenheit scale the freezing-point of water is at 32°, and the boiling-point at 212°; while on the Centigrade scale the freezing-point is at 0°, and the boiling-point at 100°. There is a further scale, the *absolute* Centigrade scale, on which the freezing-point is at 273°, and the boiling-point at 373°. The main advantage of this scale is that for a given pressure the density is inversely proportional to the absolute temperature. It has the further advantage that negative absolute temperatures, i.e. temperatures below absolute zero, cannot occur, and in the subsequent discussion of observations it is a great convenience to be free of all differences in sign.

(c) *Wind direction and speed.*
The wind direction is usually given with reference to the compass bearing from which it blows, a north wind being a wind blowing from true north. If the wind direction and speed are obtained from an instrument, the speed is expressed in any convenient unit, in miles per hour, in feet per second, or in metres per second. The relation between these units is readily remembered in the form

3 feet per second = 2 miles per hour;
1 metre per second = $2\frac{1}{4}$ miles per hour.

These relationships are sufficiently accurate for all practical purposes.

It is possible, however, with practice, to estimate the wind speed with a fair degree of accuracy without the use of any instrument. When this is done, the wind speed is expressed on what is called the Beaufort scale. Such a scale was first used by Admiral Beaufort in 1806, and after some modifications has remained in use to this day. The facing table shows the Beaufort scale now in use. Wind observations at sea are always expressed on this scale, and it is said that seamen, through long practice, become highly skilled in estimating winds, not only from the feel of the wind on their faces, but from the drift of funnel-smoke, and the appearance of the surface of the sea.

Full descriptions of instrumental methods for measuring winds will be found in *The Meteorological Observer's Handbook*. We cannot enter into the details of these methods here, but two records of wind direction and velocity are shown in Figures 2 and 3. In each of these the top trace gives the variation of

The Beaufort Scale of Wind Force, with Specifications and Equivalents.

Beaufort Number.	*General Description of Wind.*	*Specification of Beaufort Scale.*		*Limits of Velocity in miles per hour at about 30 feet above level ground.*
		*For Coast use.**	*For use Inland.*	
0	Calm	Calm.	Smoke rises vertically.	Less than 1
1	Light air	Fishing smack just has steerage way.	Wind direction shown by smoke drift but not by wind vanes.	1–3
2	Slight breeze	Wind fills the sails of smacks, which then move at about 1–2 miles per hour.	Wind felt on face; leaves rustle; ordinary vane moved by wind.	4–7
3	Gentle breeze	Smacks begin to careen and travel about 3–4 miles per hour.	Leaves and small twigs in constant motion; wind extends light flag.	8–12
4	Moderate breeze	Good working breeze; smacks carry all canvas with good list.	Raises dust and loose paper; small branches are moved.	13–18
5	Fresh breeze	Smacks shorten sail.	Small trees in leaf begin to sway.	19–24
6	Strong breeze	Smacks have double reef in main sail.	Large branches in motion; whistling in telegraph wires.	25–31
7	Moderate gale.	Smacks at sea lie to.	Whole trees in motion.	32–38
8	Fresh gale	All smacks make for harbour.	Breaks twigs off trees; generally impedes progress.	39–46
9	Strong gale	—	Slight structural damage occurs; chimney pots removed.	47–54
10	Whole gale	—	Trees uprooted; considerable structural damage.	55–63
11	Storm	—	Very rarely experienced; widespread damage.	64–75
12	Hurricane	—	—	Above 75

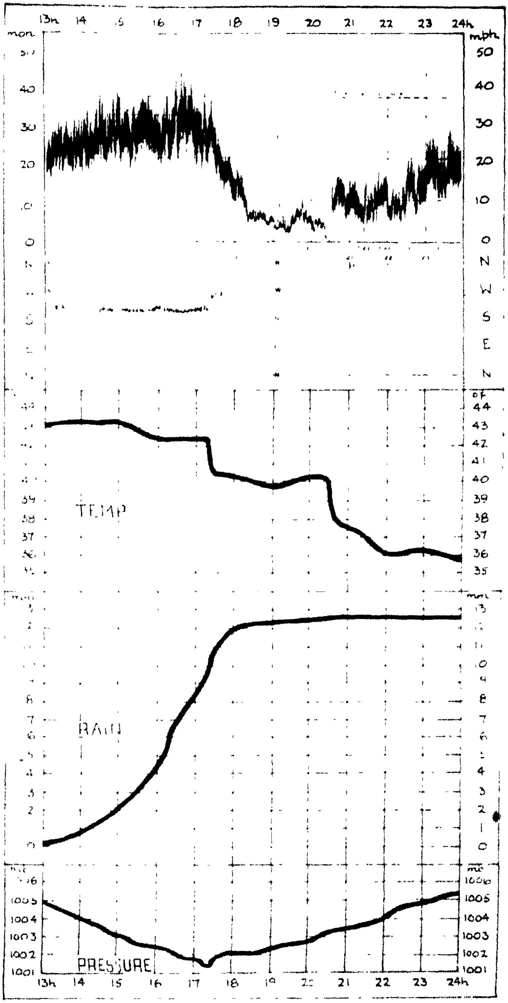
* The fishing smack in this column may be taken as representing a trawler of average type and trim.

the wind velocity during a period of some hours, while the second trace gives the wind direction for the same period. It will be seen that the wind is in no sense to be regarded as a steady current. It comes in a succession of gusts and lulls, with continual variations of direction, though both velocity and direction may maintain the same average value for hours at a time. The lack of uniformity is due to the effect of friction at the ground, which gives rise to eddies and whirls somewhat similar to those which form in a stream of water flowing over an uneven bed. The existence of these eddies in air is made visible by smoke from a chimney or from a garden bonfire.

From time to time considerable changes in the general speed and direction of the wind may set in within the course of a few minutes. In Figure 2 at about 4 a.m. there is recorded a change in the wind velocity from about 15 miles per hour to about 25 miles per hour, with a simultaneous change of direction from WSW. to WNW. In this case the original current of air was replaced by a new current from a different direction. In Figure 3 at about 17 hours 25 minutes (5.25 p.m.) there is shown a rapid change of wind direction from SSW. to WSW., with a steady decrease in the wind velocity. Here again the original current of air was replaced by a new and colder current, as is seen from the temperature trace shown in Figure 3 beneath the wind direction trace. The temperature fell through two degrees when the wind direction changed.

(d) *Weather and State of the Sky.*

A note is made at the time of observation of the presence of fog, mist, or rain; whether the sky is clear or cloudy; and if cloud is present the fraction of the sky which is covered with cloud, together with the type of cloud according to the classification which is briefly discussed below.



3. Autographic records of Wind Velocity, Wind Direction, Temperature, Rainfall, and Pressure, at Eskdalemuir
31 March 1925

(e) *Visibility.*

The visibility is measured by the horizontal distance of the farthest object which is clearly distinguishable.

The Classification of Clouds.

There are four main types of cloud: cirrus, stratus, cumulus, and nimbus.

Cirrus is the highest of all clouds, and in its typical form has a feathery or fibrous appearance. It can, however, assume a wide variety of forms. A fairly typical example is shown in Figure 8.

Stratus is a uniform layer of cloud, resembling a fog, showing little or no form.

Cumulus is the woolpack cloud of summer afternoons, with rounded tops and flat base. Figure 9 shows typical cumulus clouds arrayed in long lines.

Nimbus is a dense shapeless layer of dark cloud, from which rain or snow usually falls.

Between these four main classifications there occur a large number of sub-classifications, which sometimes take a sharply-defined special form. Thus *cirro-stratus* is a thin whitish sheet of cloud, sometimes covering the whole sky, having sometimes no clearly defined structure, but at other times taking the form of a tangled web.

Cirro-cumulus is usually in the form of small irregular globular masses arranged in lines, or it might be defined as a high cloud showing a single or double undulation. The example shown in Figure 15 is a representative one. The cloudlets here appear separated at the edge of the sheet, but in the central part of the cloud the structure may be most clearly described as an undulation.

Alto-cumulus is somewhat similar in appearance to cirro-cumulus, except that the globules are larger and show marked shadows, whereas cirro-cumulus clouds show very slight shadows, or none at all.

Alto-stratus is a dense sheet of grey or bluish colour, sometimes showing fibrous structure.

Strato-cumulus is formed of dark, heavy masses of cloud, irregularly arranged, showing brighter interstices.

Cumulo-nimbus is the thunder-cloud. It frequently takes the form of towers or anvils, with a screen of fibrous texture (false cirrus). In Figure 17 is shown a typical anvil formation, with small wisps of false cirrus at several points. Around the main cloud are shown detached irregular masses, and the dark, heavy nimbus cloud shown at the lower left-hand corner is an extension of the lower portion of the cumulo-nimbus cloud. The base of the cloud appears as a clear-cut horizontal line in the picture.[1]

The Uses of the Observations—Climate and Weather.

When a series of observations, made at the same set times every day over a long period of years, is available, one obvious use to which they can be applied is to evaluate the mean or average values of certain factors for the separate months of the year. Averages for each month of the year may be made of mean daily temperature, maximum day temperature, minimum night temperature, the daily rainfall, the wind force, the frequency of winds from different points of the compass, the number of days when rain, snow, fog, gales, or frost were experienced, the mean pressure for the day, and a number of other factors. This set of averages constitutes a body of information which defines what we may call the climate of the station in question. The climate is in fact nothing more than a statement of the average condi-

[1] The reader who is interested in the fascinating study of clouds is referred to *Clouds*, by G. A. Clarke (Constable & Co.).

tions for each month or each season of the year. The term weather, on the other hand, is used to denote the changes which take place from day to day. In the British Isles we are so obsessed by the variability of the weather from day to day that it tends to become a staple topic of conversation, but there are parts of the world, in lower latitudes, where conditions remain so nearly constant for months on end that they have no weather in the ordinary sense of the word. Thus, for example, Egypt during the months of July and August has such steady conditions from day to day that we may say that during those months Egypt has no weather, but a climate only.

The statistical treatment of observations of different kinds forms a large part of the work of all official meteorological services. The evaluation of average values referred to above is the first step in the statistical process. Then follows the computation of the average deviation of each factor from its mean value, and the selection of the extreme values for each month and year. The amount of information which is thus to be extracted from a set of observations is very considerable, but the reductions required are lengthy, so much so that it is difficult in a few words to give an idea of the labour involved. Take, for example, the observations of temperature, assuming that for each day we have available the maximum temperature, the minimum temperature, and the mean temperature for the twenty-four hours. In addition to the averages referred to above, it is important to know (*a*) the average difference between the observations of any one factor and their mean value, (*b*) the average over all the years of the highest value of each factor in each month of the year, and (*c*) the highest value recorded during the whole of the period covered. The figures thus derived would tell us, among other things, the average daily temperature for each month, the average maximum day temperature, the average minimum night temperature, the highest and lowest temperatures which occur on the average in each month of the year, and the actual highest and lowest temperatures recorded in each month of the year during the period covered by the observations.

Again, it is substantially true for all parts of the world that the temperature of a particular day depends largely on the wind direction, and may be even more dependent on the wind direction than on whether the sky is clear or cloudy, particularly in middle latitudes in winter. It is therefore important to evaluate the frequency of occurrence of winds from each point of the compass.

The data presented in this form are useful to a variety of interests. Agriculture is obviously affected by all the averages and extremes we have mentioned; aviation is affected by winds, both as to speed and direction, and by the weather and visibility; and shipping and fishing are affected by wind and weather. An army which is to carry on warfare in a relatively strange country requires information of the fullest possible nature concerning all the factors of weather. The amount of clothing troops require by day will depend on the day temperature, and the covering required at night will depend upon the minimum night temperature. Military operations may be impeded by fog, continuous rain, or wind from a particular direction; entrenchment may be rendered impossible by continuous heavy frost; and transport may be impeded by a sudden thaw. In short, war in all its aspects may be impeded by unfavourable weather conditions, and it is clearly of importance to know beforehand the chances of encountering these unfavourable conditions.

In certain parts of the world, particularly in the tropics, the climatic conditions present clearly defined differences between one season and another. Such places may have one or two rainy seasons during the year, separated by relatively dry periods. The statistical study of observations which we have described above brings out these climatic features, and the information thus obtained is of value to many interests. For example, an airman wishing to fly across Africa would need to study carefully the times of occurrence of rainy seasons along different parts of his route, so that he might avoid them.

An example of the importance of considering not only the average rainfall for the year, but also the variability of rainfall from year to year, is shown in the case of water-power schemes, which require not only a certain average rainfall, but also a reasonable minimum rainfall each year.

IV

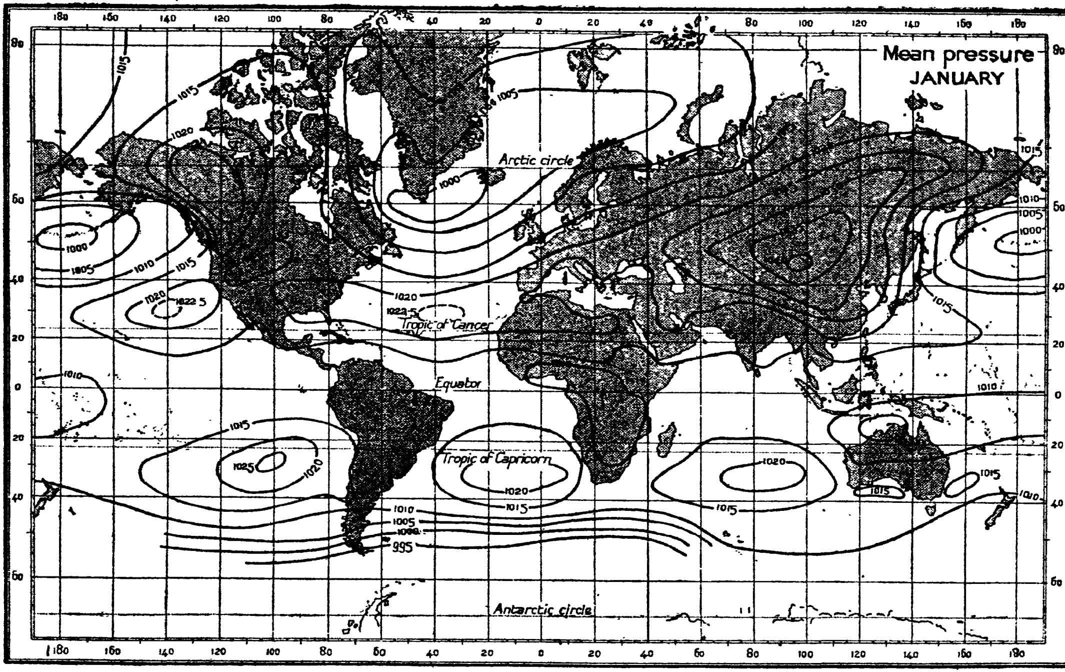
The Idea of a General Circulation of the Atmosphere

THERE is a wider aspect to the use of the average values of various meteorological factors, which is brought out most clearly when similar data for the whole globe are plotted upon charts. It is thus that we obtain charts of the mean pressure distribution, of the mean temperature distribution, of the prevailing winds, and of the average rainfall, over the surface of the globe, and it is by studying such charts, month by month, that we arrive at the conclusion that the atmosphere has what is known as a 'general circulation'. The term requires explanation, as it is not obvious to an individual living anywhere in the middle latitudes of the northern hemisphere that the atmosphere is in any way systematic in its behaviour. When we draw charts on which are represented, at a large number of stations, the most frequently occurring wind directions (in other words, the prevailing winds), we find the movement of the winds of the globe form a system which is much simpler in appearance than might have been expected by one accustomed only to the variability of weather in the British Isles. This system is called the 'general circulation' of the atmosphere. It is true that the winds of the globe do not at any one instant conform accurately to this system, but the conception is useful. It is convenient to regard the general circulation as a background upon which are superposed a certain set of variations in the form of local circulations, so that the general circulation plus local circulations represent the actual conditions prevailing at any instant.

In practice it is found far more convenient to express the wind circulation of the atmosphere in terms of the pressure distribution. When upon a chart of the world we have inserted at a sufficient number of stations the mean pressure for, say, a given month of the year, we can draw isobars, or contour lines of pressure, which are such that at every place lying on one and the same isobar the mean-sea-level pressure is the same. In this way we shall obtain a map giving the mean distribution of pressure for the month, or other period covered by the observations. In Chapter I we mentioned Buys-Ballot's law, which states that the winds blow around the isobars, tending to blow slightly into the low pressure, but on the whole keeping low pressure to their left and high pressure to their right in the northern hemisphere. In the southern hemisphere the winds have low pressure to their right, and high pressure to their left. When in the course of this book no definite statement to the contrary is made, it may be assumed that reference is made to the northern hemisphere.

The Pressure Distribution.

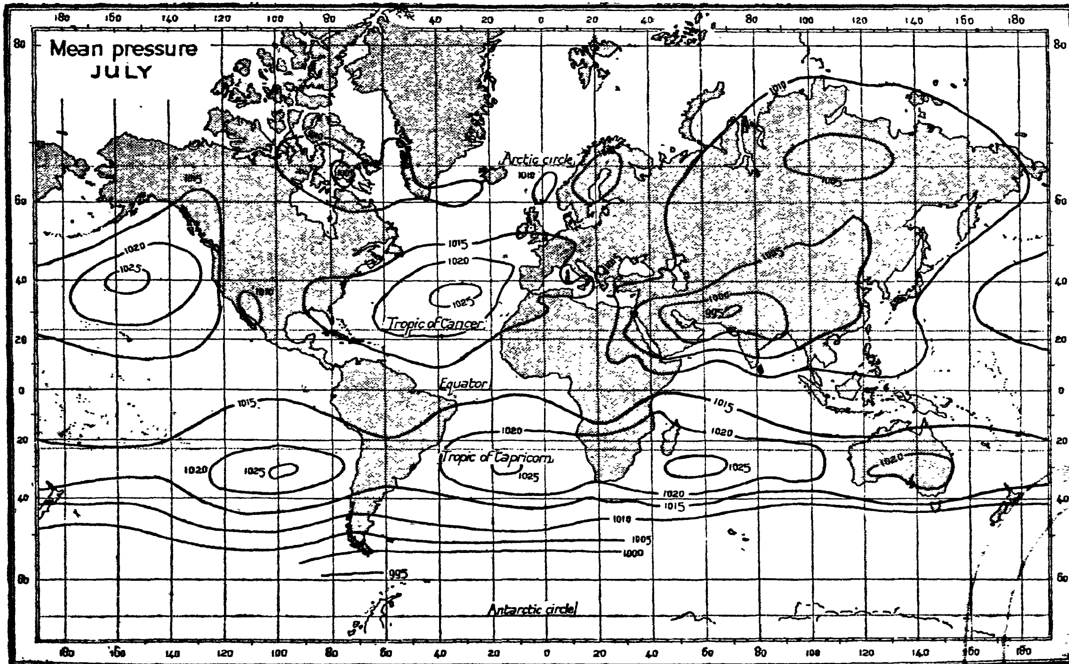
Maps of the mean pressure for January and July for the two hemispheres are shown in Figures 4 and 5. The winds blow



4. Isobars of Mean Pressure for January

around the isobars in the sense indicated in the last paragraph. Certain features of the maps for winter and summer are common to both, particularly in the southern hemisphere. In both summer and winter we find around the equator a region of nearly uniform pressure distribution, in which the winds are very light or variable. This belt is known as the Doldrums, and is a region of intermittent heavy rain and thunderstorms. Its mean position is slightly north of the equator, and it moves northward in the northern summer, and southward in the northern winter. To north and south of this belt there are belts of high pressure encircling the whole earth, having easterly winds on their equatorial sides, and westerly winds on their poleward sides. These belts of high pressure are most intense over the oceans, though they show pressures definitely above normal over the continents. They are generally called the subtropical high pressures or subtropical anticyclones. The easterly winds in each case have a component of motion towards the equator, so that they are north-easterly in the northern and south-easterly in the southern hemispheres. These are the trade winds, and are called the North-east Trade and South-east Trade respectively. They blow around the equatorial margins of the subtropical high pressures, finally dying away into the doldrums, or regions of calm. The air brought towards the equator by these winds ascends in the region of the doldrums, and in part at least returns polewards in the upper air, as a south-westerly current in the northern hemisphere, and as a north-westerly current in the southern hemisphere. The trade winds are relatively shallow, and the reversal of the wind direction at no great height has frequently been observed. The reversed wind currents over the trade winds are called the counter-trades.

It will be seen that in the southern hemisphere there is little difference between the charts for January and July. The region

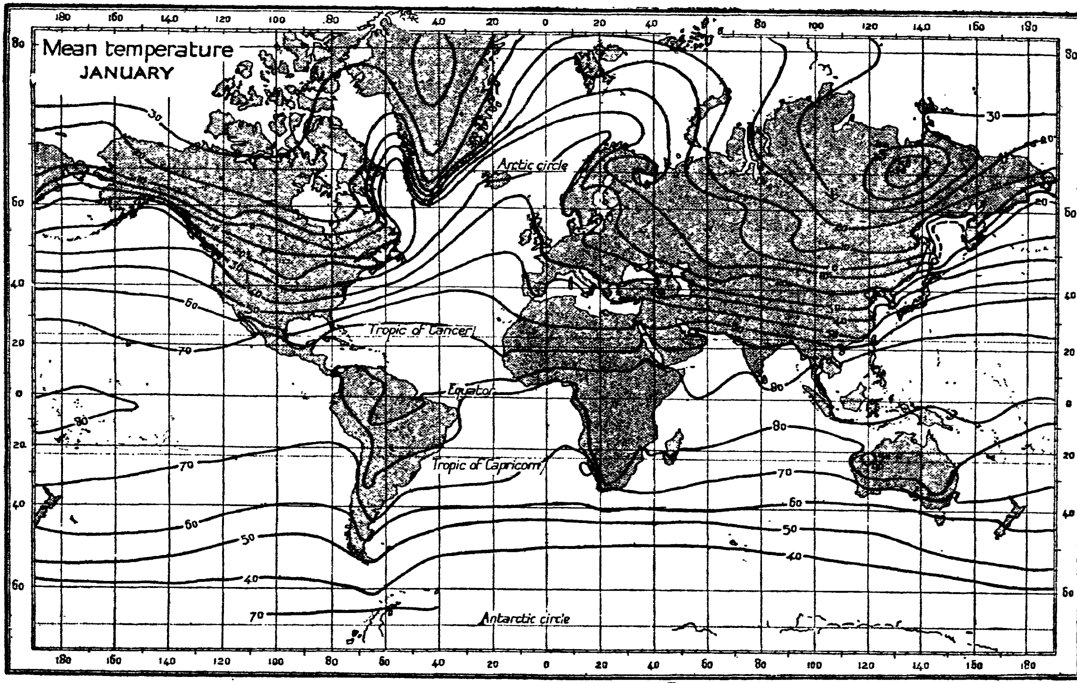


5. Isobars of Mean Pressure for July

of westerly winds is not a region of uniformity from day to day. Centres of high and low pressure move along in the main westerly current, and produce considerable local deviations. But on the whole the form of the isobars of mean pressure is far more symmetrical about the pole than in the northern hemisphere, largely because the southern hemisphere is almost entirely an ocean hemisphere.

In the northern hemisphere we find, from even a casual glance at the maps, that there is a marked tendency for anticyclones to form over the continents in winter, and over the oceans in summer, and for low pressures to form over the oceans in winter and over the continents in summer. These high and low-pressure systems, added to the subtropical belts of high pressure, produce a circulation which is rather more complex than that in the southern hemisphere. The western side of an anticyclone is a region of southerly winds, while its eastern side is a region of northerly winds, so that while the former conveys warm air polewards from low latitudes, the latter conveys cold air equatorwards from high latitudes. On the poleward side of the anticyclones the winds are west or south-west (in the northern hemisphere), up to fairly high latitudes, beyond which there is a cold easterly current.

It is perhaps worth repeating that while these charts represent the average conditions during January and July, the state of things from day to day may differ widely from this. For example, the January chart for the northern hemisphere shows an extensive area of low pressure south of Greenland. This must not be interpreted as meaning that on any day in January there will be a region of low pressure there. The depressions which account for the variability of the weather of the British Isles originate in the Atlantic, and on the average travel in a direction WSW. to ENE. The elongated area of low pressure shown on



6. Isotherms of Mean Temperature for January

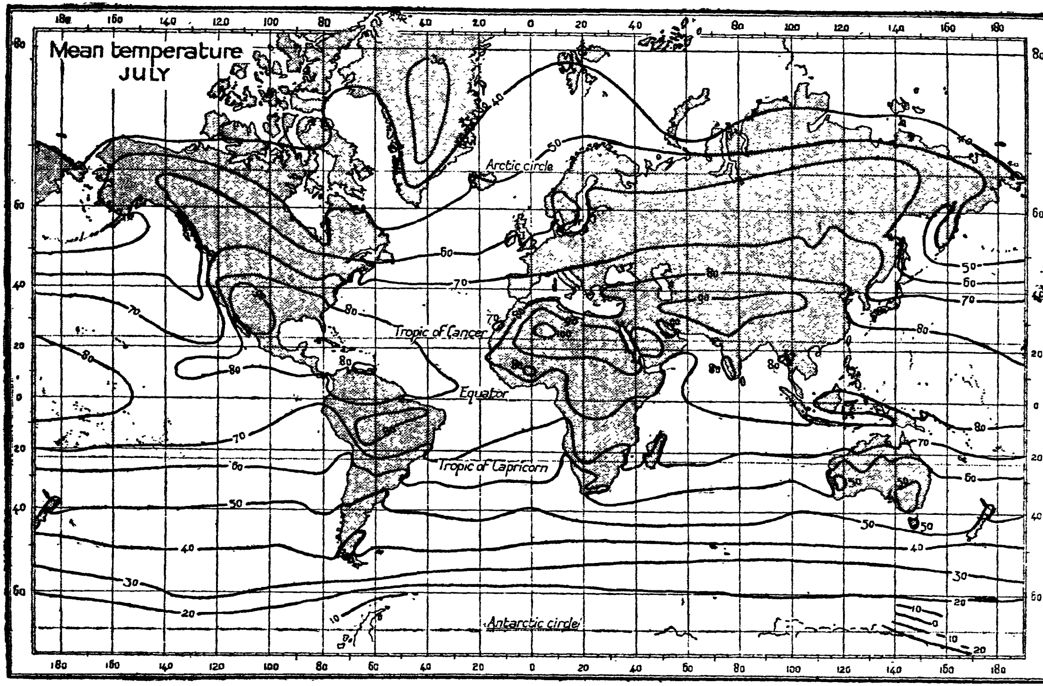
the chart of monthly mean pressures marks the mean position of the tracks of these depressions across the North Atlantic.

In a later chapter we shall consider in a little more detail the functions of the moving cyclones and anticyclones as features of the general circulation. It has already been stated that anticyclones assist in the exchange of warm and cold air between low and high latitudes. The regions of low pressure, or depressions, of middle latitudes perform the same function, and so help to maintain the general circulation of the atmosphere. It is in this way that heat and momentum are conveyed from low to high latitudes.

The Distribution of Temperature over the Globe.

In Figures 6 and 7 are shown charts of the mean monthly temperatures for January and July for the two hemispheres. It will be seen that the highest temperatures occur in the tropics, and that the oceans are warmer than the continents in winter, and colder than the continents in summer. Over the southern hemisphere, so far as observations are available, the distribution of temperature is symmetrical about the South Pole. Over the northern hemisphere the distribution is less symmetrical, and in January the lowest temperatures occur, not at the North Pole, but over northern Siberia. Here the radiation of heat from the huge land area which forms the Eurasian continent results in intense cold in winter, leading to the formation of a large and intense anticyclone. There is a second centre of intense cold over Greenland in January. It will be noted that in winter the isotherms or lines of equal temperature swing downward towards the equator as they approach the centres of the continents.

In summer the lowest temperature in the northern hemisphere is no longer over Siberia, and the general run of the



7. Isotherms of Mean Temperature for July

isotherms is more symmetrical about the pole. The highest temperatures, however, are quite definitely located over the continents.

The Monsoons.

During the winter the intense anticyclone which forms over the continent of Asia extends from northern Siberia as far as the subtropical belt of high pressure, with which its southern edge merges. On the south-eastern side of this anticyclone the surface wind is north-easterly. This wind is the normal North-east Trade, strengthened by the effect of the Asiatic anticyclone. It is, however, frequently referred to as the North-east Monsoon. The word *monsoon* means *seasonal*, and is quite generally applicable to any winds which blow only for a season, and are produced by seasonal changes of pressure. Monsoon winds occur over portions of the coasts of all the continents, blowing from the sea in summer, and from the land in winter; but the word *monsoon* is most commonly used in connexion with the north-east monsoon referred to above, or the south-west monsoon of the Indian Ocean in summer. When we speak of 'the monsoon' without any qualifying word, we refer to the south-west monsoon of India.

With the approach of summer the Asiatic anticyclone dies away, and is replaced by an immense region of low pressure, covering almost the whole of Asia, and extending south-east to the equator and westward half-way across Africa. The circulation of the wind around this depression is on a magnificent scale. At its southern edge it is marked by a broad south-westerly current, blowing from the subtropical high pressure of the southern hemisphere across the equator, across the Indian Ocean and India, into the Asiatic low pressure. This south-westerly current is the South-west Monsoon, a warm and damp current, which on reaching India rises over the mountain ranges, and in so doing gives the heavy monsoon rainfall of the Indian ' summer '. It starts in June and continues until late September, when it dies away as the Asiatic low pressure fills up. During this period the belt of doldrums and the subtropical high-pressure belt of the Indian Ocean are both completely submerged. The monsoon blows as a steady wind, occasionally reaching gale force in places.

The eastern edge of the extensive cyclone over Asia is marked by a southerly wind—the southerly monsoon of the China Seas. At the same time the southern fringe of the low pressure over the Mediterranean is marked by a northerly wind—the Etesian wind of the Mediterranean. These winds are all parts of the wind circulation of the Asiatic cyclone, and all form parts of what is by far the greatest surface cyclonic circulation of the earth's atmosphere.

We have thus arrived at a picture of the major circulation of the atmosphere as controlled by the subtropical high pressures, the continental anticyclones and oceanic depressions of winter, or the continental depressions and oceanic anticyclones of summer. Upon this major circulation, which retains the same main features from year to year, are superposed moving regions of high and low pressure, which we may refer to as the local circulations. The atmosphere thus represents a heat machine, whose parts move in an orderly manner, and whose working material is air, or rather air containing water-vapour. The motive-power of the machine is supplied by the sun, but before we come down to the detail of the local circulations and their possible causes, we must consider more closely the nature of the sun's radiation and its action upon the atmosphere.

V

Solar Radiation and its Reception in the Atmosphere

WHEN a beam of light from the sun is passed through a glass prism it is spread out into a long, coloured band or spectrum in which the primary colours appear in the following order: violet, indigo, blue, green, yellow, orange, and red. Actually there is no definite line of demarcation between one colour and the next, and it would be more accurate to describe the band as made up of an infinite gradation of colours. The physical distinction between the colours is in the wave-length of the vibrations which produce them. The reader is probably familiar with the wave-length of disturbances on the surface of the sea as the distance between successive wave-crests. The analogy is quite a legitimate one to use, but it is perhaps simpler in dealing with light to accept wave-length as a scientific and impersonal method of defining colour. The actual wave-lengths of light are very minute when measured on any everyday scale, and are usually measured in terms of a unit represented by the Greek letter μ (*mu*), whose value is one millionth of a metre, or one thousandth of a millimetre. Visible light has wave-lengths varying from about 0·4 μ at the violet end of the visible spectrum, to about 0·8 μ at the red end.

The visible spectrum terminates by a gradual fading away at each end of the band, but by the use of suitable photographic plates it is possible to show that the solar spectrum has an extension beyond the violet into a region of shorter wave-lengths, known as the ultra-violet, and beyond the red into a region of longer wave-lengths, known as the infra-red, or heat rays.

Light waves are waves in the ether, of the type known as electromagnetic waves. To this same category of electromagnetic waves belong X-rays, radiant heat, and Hertzian waves (including the waves used in wireless transmission). They all travel with a speed of 186,000 miles per second, and differ among themselves only in their wave-length. They may all be represented in one continuous scale of wave-length, or of frequency. The term frequency is used here, as in sound, to denote the number of vibrations per second. The shortest wave-lengths are those of X-rays, which extend up to about $0 \cdot 1\ \mu$. Then comes ultra-violet light extending up to the lower limit of the visible spectrum at about $0 \cdot 4\ \mu$. Infra-red rays extend from $0 \cdot 8\ \mu$ up through a considerable range of wave-lengths, and all waves with a length greater than $100\ \mu$ are classed as Hertzian waves. The differences of frequency involved in such a scale are enormous, and the step from red light to the wireless waves used in broadcasting from Daventry (1,604 metres) amounts to almost exactly thirty-one octaves, which is roughly equivalent to four times the range of frequency of an ordinary piano. The frequency corresponding to Daventry's wave-length is of the order of 200,000, which is about eight times the frequency of the highest note audible to the human ear.

Methods have been developed for obtaining numerical values of the intensity of the radiation in different parts of the spectrum. In the solar spectrum the position of maximum intensity is in the blue, and the intensity decreases rapidly as we go away from this position. Beyond the visible spectrum the intensity decreases rapidly on each side. The theory of radiation shows that the wave-length at which the intensity is a maximum is inversely proportional to the absolute temperature (see p. 17), and the determination of the point of maximum intensity in the solar spectrum gives us the surface temperature of the sun at about 6,000° C. For bodies with lower temperature than the sun, the point of maximum intensity is shifted towards the red, and the radiation from bodies at ordinary terrestrial temperatures is a long way down in the infra-red part of the spectrum. The hotter the radiating body, the shorter are the wave-lengths it sends out.

The term 'high temperature radiation' is frequently used to denote the short-wave radiation of very hot bodies, and the term 'low temperature radiation' to denote the long-wave radiation of bodies at relatively low temperature. Also the infra-red rays are frequently referred to as heat rays simply because such rays are radiated from bodies at low temperature which do not send out any visible rays.

Solar radiation falling upon any body exposed to it may be used up in heating that body; and so we frequently speak of the sun as the source of heat. But it should be noted that, although a considerable fraction of the solar radiation is in the infra-red, more than a half of the total energy is contained within the range of visible wave-lengths. These visible rays, when absorbed by a material body, also contribute to heating it. All radiation from the sun, of whatever wave-length, can be absorbed by material bodies or by the atmosphere, and converted into heat, while only a restricted range of wave-lengths can be observed by the human eye as light. The distinction between light and radiant heat is thus a physiological and not a physical distinction, since all wave-lengths which come within the range of the visible spectrum may also be regarded as heat rays.

When a beam of light is passed through an atmosphere of any kind, rays of certain wave-lengths, corresponding to the periods of vibration of the molecules of the atmosphere, are absorbed by those molecules. We shall not enter here into the physical nature of absorption, but shall content ourselves with saying that the radiation of certain narrowly restricted wave-lengths is absorbed by the molecules, the energy taken away from the beam being converted into heat. This effect is shown in the spectrum by the presence of a number of narrow dark lines running across the spectrum in definitely fixed positions, marking the absence of the light absorbed. Each element in the absorbing medium produces its own set of dark absorption lines. This fact forms the basis of the branch of physics called 'spectroscopy'. The almost innumerable dark lines observed in the spectrum of the sun indicate the presence in the sun of most of the elements known on the earth. These lines are due to absorption in the cooler outer layers of selected wave-lengths of the radiation sent out from the deeper layers of the sun.

Again, when the radiation from the sun passes through the earth's atmosphere, there is a certain amount of absorption by the molecules of the different constituents of the atmosphere, notably by ozone in the highest layers of the atmosphere, and by water-vapour at lower levels.

The Solar Constant is a measure of the intensity of the radiation from the sun at the outer limit of the earth's atmosphere. The usual method of expressing it is to say that it is nearly equal to two gramme-calories per square centimetre per minute; which means that the amount of radiation falling per minute on one square centimetre of a surface set at right angles to the incoming radiation would raise the temperature of one gramme of water through 2° Centigrade. The name 'solar constant' appears to have been rather a misnomer, as this quantity is reported to be subject to variations from time to time.

Radiation on its inward journey through the Atmosphere.

We have thus to picture a beam of short-wave light from the sun reaching the outer boundary of the earth's atmosphere. On its way down the radiation loses some of its intensity in selected wave-lengths through absorption by molecules of the different constituents of the atmosphere. This loss is very slight, for the atmosphere is nearly transparent for the short waves coming from the sun. There is a further loss from the downward beam due to the effect of 'Scattering' caused by the molecules of air and suspended small particles of dust or water-drops. The word 'scattering' can be interpreted here in its literal sense. The effect is to take a certain amount of radiation from the incoming beam, and to send it out in all directions from the scattering particle or molecule, without having any effect on the temperature of the scattering particle. The scattered radiation remains as radiation, in contrast with absorbed radiation, which ceases to exist as radiation, and is converted into heat in the ordinary sense of the word. This scattering effect is greatest for light in the blue end of the spectrum, and the blue colour of the clear sky is explained as due to the scattering of light by molecules of air and water-vapour.

A large amount of incoming radiation is reflected back to the sky by clouds, and by portions of the surface of the earth, and represents dead loss as far as heating effect is concerned. The balance which escapes absorption and reflection by the air is absorbed or reflected by the continents and the oceans, the energy absorbed being converted into heat, and producing a definite increase in temperature, while the reflected energy is lost in space.

Radiation on its outward journey through the Atmosphere.

The heating of the earth's surface goes on daily while the sun is shining, but the earth's surface does not become hotter day after day, because it radiates back up into the atmosphere the energy it absorbs from the solar radiation. The earth's radiation is low-temperature radiation of long wave-length, and this radiation is much more readily absorbed by the water-vapour in the atmosphere than is the direct downward beam. In addition it is subject to a certain amount of scattering by the molecules of air and water-vapour. If we think of a thin horizontal layer of air at any level, we can readily write down its effects on radiation passing upward from the ground. Of the upward beam which enters it at the lower face some is absorbed, some is scattered in all directions, and the balance is passed out again at the upper face. In addition the layer will itself radiate a certain quantity of heat, depending on its temperature and its content of water-vapour. If the temperature of the layer remains constant, there must be a balance between the energy radiated out from the layer and the energy absorbed in it. It is important to note that any layer of air which absorbs the low-temperature radiation from the earth also emits radiation in both upward and downward directions, and so there is in the atmosphere at any moment a downward stream of low-temperature (long-wave) radiation emanating from higher layers of the atmosphere itself. This is the clue to the difference between cloudy and clear nights. When the sky is clear and bright, radiation from the earth's surface passes through the atmosphere out into space, and the surface of the earth becomes increasingly colder through the night. When the sky is cloudy, the radiation from the earth is partly reflected by the clouds, and partly absorbed and then radiated back again to the earth's surface, with the result that the temperature remains relatively steady throughout the night.

We may summarize the above description briefly as follows: The radiation from the sun passing through the earth's atmosphere suffers very little loss by absorption, and only a little more by scattering, so that except for the loss by reflection from clouds, the total loss on the downward passage is small, amounting possibly to 10 per cent. The part which reaches the earth is absorbed by the earth's surface, and an equivalent amount of energy is radiated back to the sky in the form of long-wave or low-temperature radiation. This type of radiation is readily absorbed by water-vapour in the atmosphere, which in turn radiates the energy it receives partly back towards the earth, partly up to the sky. From the point of view of radiation the action of the water-vapour is to impede the passage of the terrestrial radiation back into space.

There is a close analogy between the effect of water-vapour in the atmosphere and the effect of the glass of a greenhouse exposed to sunshine. The incoming short waves of solar radiation pass readily through the glass, and are absorbed by the damp air and the objects inside the greenhouse. These in turn send out long-wave radiation which is absorbed by a thin inner layer of the glass and radiated back again. The glass thus permits solar radiation to pass inward, but impedes the outward passage of the long-wave radiation from the air and the objects within. Hence the temperature inside the greenhouse becomes much higher than the temperature outside.

Over the oceans the phenomena are considerably different. Unless the sun is overhead or nearly so the greater part of the radiation which reaches the sea surface is reflected back to the sky, and is in effect lost so far as heating the water and the atmosphere is concerned. When the sun is sufficiently near to the

8. CIRRUS CLOUD

ɪith, the amount reflected by the sea is small, but in this e the radiation penetrates to a considerable depth before it , completely absorbed. This is borne out by the fact that it is easy to distinguish the pebbles at the bottom of a river provided the observer is so placed that he does not see a reflection of the sky in the water, the pebbles being illuminated by light which passes downward through the water. The heating of the water by the absorption of radiation is therefore spread through a considerable depth. This is in marked contrast with what happens on land, where only a few inches of the top soil are heated to any marked extent, and for this reason alone we should expect the variations of the surface temperature of the sea to be far less than those on land. But there are in addition several other factors which help to keep down the changes of temperature at sea. Some of the energy absorbed is used in evaporating water from the sea surface, and the amount available for heating the water is correspondingly decreased. Again, if the same quantity of heat be communicated to a given mass of water and the same mass of soil, the increase in the temperature of the water is only about one-quarter the increase in the temperature of the soil. When to this fact we add that at sea the amount of radiation available for heating the water is first diminished by reflection and evaporation, and then spread through a great depth of water, we readily see that the increase in the surface temperature of the sea produced by the sun during the day must be slight.

At night, the sea surface readily emits long-wave radiation, ut the decrease in temperature produced by the loss of heat only about one-quarter the decrease in temperature that ould be produced in the same mass of soil by an equal loss of at. When the surface-layer of water is cooled, it contracts ghtly and becomes denser, and therefore tends to descend and spread the cooling effect through a considerable depth. Thus the decrease in surface temperature of the sea produced by the loss of heat by radiation at night is also slight.

We thus conclude that the sea surface is subject to only very slight daily changes of temperature. Observations show that the difference between day and night is at most about 1° F. Precisely the same argument applies to the annual variation of temperature, the sea being less heated than the land in summer and less cooled than the land in winter. The smaller variability of temperature over the sea as compared with that over land is very marked in all latitudes, the oceans being less subject to violent extremes than the continents. This accounts for the distinction between oceanic and continental climates. Outside the tropics the climates of countries bordering the oceans are equable, having relatively cool summers and mild winters, unless the winds blow steadily outward from the middle of a continent, bringing with them the extreme temperatures usual in the middle of land masses. In the British Isles, for example, a south-westerly wind from the Atlantic is warm in winter and cool in summer. But over the eastern coast of Asia, in the same latitude as the British Isles, the winds blow from the north or north-west in winter, bringing with them the extreme cold of the Siberian anticyclone. In the British Isles the effect of wind-direction on the weather is in winter greater than that of direct sunshine, a cloudy day with a south-westerly wind being warmer than a sunny day with a northerly wind.

Generally speaking, in middle latitudes places on the western coasts of the continents have milder winter climates than those on the eastern coasts in the same latitudes. Thus England is in approximately the same latitude as British Columbia, and they both have relatively mild winter climates, British Columbia having more rainfall on account of the high barrier of the Rocky

Mountains. But the lower St. Lawrence is in lower latitudes than the whole of the British Isles and is icebound for many months in winter. The differences here referred to may be ascribed mainly to the fact that in middle latitudes the prevailing winds are westerly, and in winter are therefore mild oceanic winds for places on the western coasts of the continents; while over the eastern coasts of the continents the westerly winds blow from the cold interior of the continents, and are therefore in winter very cold winds.

VI

The Variation of Temperature in the Atmosphere and some of its Physical Effects

The Variation of Temperature with Height; Troposphere and Stratosphere.

BY the use of sounding balloons, kites, or kite balloons carrying self-recording instruments, observations of temperature in the upper air have been obtained in sufficient numbers to enable us to state how, on the average, the temperature varies with height. It is found that from the ground up to a height of several miles the temperature decreases about 3° F. for every 1,000 feet.

Observations have shown, however, that at a height varying from about 50,000 feet above the equator to 30,000 feet above latitude 50°, and 20,000 feet or less above the poles, the temperature suddenly ceases to fall with increasing height above the earth's surface, and remains constant, in some cases even showing at first a slight increase with height. The atmosphere is thus divided into two distinct shells, an inner shell within which temperature falls off with height at the rate of approximately 3° F. per 1,000 feet, and an outer shell within which the temperature remains sensibly constant at all heights. The first, or inner shell, is called the *troposphere*, the second, or outer shell, is called the *stratosphere*, and the boundary between the two is called the *tropopause*. The troposphere is the region within which convection is operative. In the stratosphere, the conditions are determined entirely by the balance between incoming and outgoing radiation, and it has been shown by Gold, Humphreys, and others that this balance demands that the temperature shall remain sensibly constant at all heights in the stratosphere. The stratosphere is colder over the equator than over the poles, and we thus have the remarkable result that the lowest temperatures observed in the earth's atmosphere are those above the equator at heights exceeding 50,000 feet. The lower boundary of the stratosphere is higher in low latitudes than in high latitudes; it is also higher over anticyclones than over cyclones.

The rate of decrease of temperature with height is called the *lapse-rate*, and the average conditions in the troposphere are specified by a lapse-rate of 3° F. per 1,000 feet. The lapse-rate may on occasion differ widely from this average value, particularly in the lower layers near the ground, and its magnitude is of considerable importance in determining whether the atmosphere is stable or unstable, as will be shown below.

Stability and Instability.

If two fluids which do not readily mix are put into the same vessel, the lighter fluid floats on top of the heavier. It is perhaps possible to have the heavier fluid floating on top of the lighter fluid, but this can only be possible so long as the fluids are entirely undisturbed. The slightest disturbance causes the lighter fluid to rise to the top. Thus the only permanent arrangement of two fluids is the one in which the lighter floats on top.

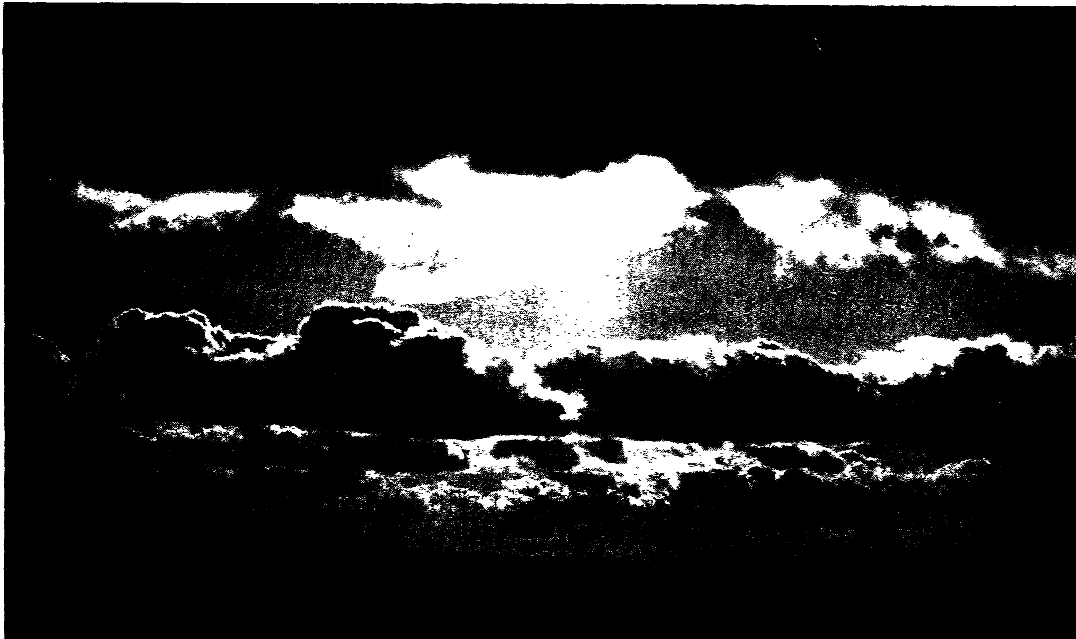
When we come to consider the analogous problem for air we are immediately faced with a complication due to the variation of density with change of pressure. In liquids such as water the changes of density produced by moderate changes of pressure are insignificant. In gases such as air the density is proportional to the pressure, provided the temperature remains unchanged. If in addition the temperature varies, we must make allowance for this in computing density. When air is compressed its temperature is increased. No one who has used an ordinary bicycle pump can have failed to notice that the pump becomes hot. The air which is compressed in the pump is heated, and in turn heats the metal of the pump. Conversely, when air is expanded it cools.

If a mass of dry or unsaturated air rises through its environment, its pressure diminishes and it expands and cools, unless heat is communicated to it from some outside source. In general, we need not take into account the possibility of heat being communicated to the moving air from outside, and the effects of radiation and absorption can be neglected by comparison with those of expansion or contraction with change of pressure. With these reservations it is possible to compute the rate of change of temperature of a rising mass of air. We shall merely state the result of the computation, which is, that when dry or unsaturated air rises, its temperature falls 1° C. for every 100 metres of ascent, or about 5½° F. for every 1,000 feet of ascent. When saturated air rises, the expansion and cooling produce condensation of water-vapour into water-drops. Heat is liberated by condensation, and this heat becomes available for heating the air. The result is to halve the rate at which the temperature falls with ascent, reducing it to ½° C. for every

100 metres, or roughly 3° F. for every 1,000 feet. If now we think of a mass of air which contains some water-vapour, but not enough to saturate it, rising through its environment, its temperature at first decreases by 1° C. per 100 metres, until the stage is reached when the water-vapour it contains is just sufficient to saturate it. From this point upward its temperature will only decrease at the rate of ½° C. for every 100 metres of ascent, and during this stage more and more of its water-vapour content will condense in the form of water-drops, which will either be carried up as a cloud, or will fall as rain.

The rate of diminution of temperature with height in the case of ascending dry or unsaturated air (1° C. per 100 metres, or 5·5° F. per 1,000 feet) is called the *dry adiabatic lapse-rate*, and similarly the rate of diminution with height for ascending saturated air (½° C. per 100 metres, or 2·8° F. per 1,000 feet) is called the *saturated* or *wet adiabatic lapse-rate*. The word *adiabatic* requires definition. Any changes to which an isolated mass of gas is subjected are said to be *adiabatic* when it neither receives heat from, nor gives heat to, any outside body. The word adiabatic might therefore be taken as synonymous with complete *thermal isolation* from the rest of the universe. Air which is not in contact with the ground can only receive heat from, or give up heat to, its environment by radiation or absorption; and, in general, when a mass of air is rising or falling, the changes of temperature produced by radiation and absorption are quite insignificant by comparison with those produced by expansion due to ascent or compression due to descent. Hence the changes may be assumed to be adiabatic, provided it does not mix with the environment. The effects of mixing are considered later, on p. 54.

If the lapse-rate from the ground upward is greater than the dry adiabatic lapse-rate—the air being unsaturated—then any



9. CUMULUS CLOUDS

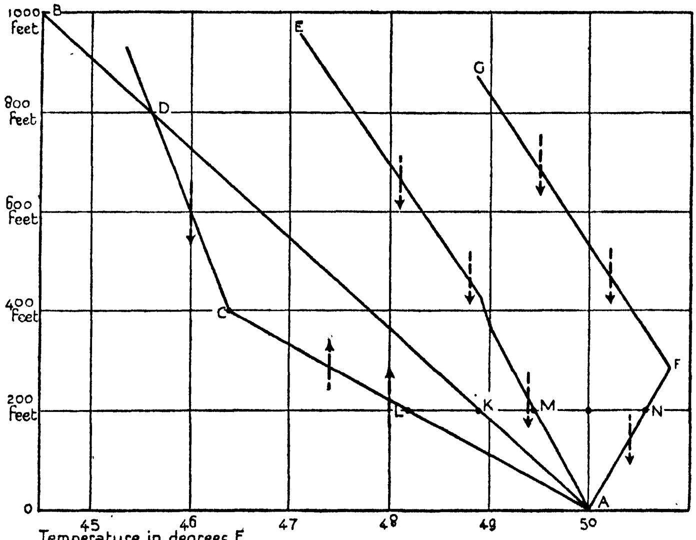
disturbance will cause the air near the ground to rise, its place being taken by air descending from higher levels. In other words, the air is then unstable. If the lapse-rate is less than the dry adiabatic, the air is stable; and if we were to move upward any isolated mass of air from any level, then when it got to any higher level it would be cooler, and therefore denser than its environment at the same level. It would therefore fall back again to its original level. The tendency of the air in a stable condition is to retain its original distribution of levels. Whereas, if the air is unstable, having a lapse-rate greater than the adiabatic, then if we displace upward an isolated mass of air it will be warmer and lighter than its environment in its displaced position and so will tend to rise farther. Thus any disturbance of air which is unstable must result in complete churning up through all levels to which the instability extends.

The last paragraph refers only to dry or unsaturated air. If the air is saturated to begin with, then the limit between stability and instability is a lapse-rate equal to the wet or saturated adiabatic. The air will then be stable if the lapse-rate is less than the saturated adiabatic ($\frac{1}{2}°$ C. per 100 metres), and unstable if the lapse-rate exceeds this limit.

What we are here concerned to emphasize is that we shall not expect to find lapse-rates in excess of the dry adiabatic persisting for any length of time in dry or unsaturated air, or lapse-rates exceeding the wet adiabatic persisting in saturated air. Lapse-rates greater than these limiting values may (and do) occur for short intervals of time, but they represent an unstable condition which cannot persist indefinitely.

In Figure 10 the temperature at different heights is represented graphically, temperature being measured along the horizontal axis, and height along the vertical axis. The adiabatic condition is represented in this diagram by a line whose back-ward slope gives a decrease of temperature of about $5\frac{1}{2}°$ F. for each 1,000 feet. Such a line is AB, which has been drawn for a ground temperature of 50° F. For any other ground temperature the line would be drawn through the corresponding point on the axis, parallel to AB. If any mass of air starting with an initial temperature of 50° at the ground ascends, its temperature at different stages of its ascent can be read off the diagram. Thus at 400 feet its temperature will have dropped to 47·8°, while at 1,000 feet its temperature will be $44\frac{1}{2}°$. The lines ACD, AE, AFG represent observations taken under varying conditions. When the relation of temperature to height is that given by ACD, the lapse-rate is greater than the adiabatic from the ground up to C, i.e. to a height of 400 feet. This follows at once from the fact that the slope of AC from the vertical is greater than the slope of AB, the slope of the line being a measure of the lapse-rate. If, then, a mass of air starting from the ground moved upward through the atmosphere whose temperature-height relation is shown by ACD, at each stage of its ascent the rising air will be warmer than its surroundings by an amount equal to the separation of the lines AC and AB. For example, at a height of 200 feet, the ascending air has a temperature of 48·9°, corresponding to the point K, while its surroundings at that level will have a temperature of 48·2°, corresponding to the point L. The displaced air will thus be warmer, and therefore lighter, than its surroundings, and will continue to ascend. It will in fact ascend until it reaches the height of 800 feet, corresponding to the point D, since this will be the height at which it will have the same temperature as the surrounding air at the same level. Thus the air will be unstable under the conditions represented by the line ACD, up to a height of 800 feet.

The irregular line AE has everywhere a slope from the vertical less than the adiabatic line AB, and represents a stable condition. Air which is caused to move upward from the ground will at each stage of its ascent be colder than the surrounding air at the same level. If, for example, it could be caused to rise to



10. Some specimen Lapse-rates, with arrows indicating the directions of Transfer of Heat by Turbulence (*vide* p. 55)

a height of 200 feet, its temperature would be 48·9°, whereas the surrounding air would have a temperature of nearly 49·5°, corresponding to the point M. It would therefore be colder and denser than its surroundings, and if left to itself would fall back to its original level. The conditions represented by AE are therefore stable conditions.

The line AFG represents a state of things which frequently holds in winter after a clear night. The temperature increases with height up to the height of the point F, and then decreases. The increase of temperature with height is called an *inversion.* Such a condition is extremely stable. The temperature at a height of 200 feet is shown by the point N (about 50·6°), and if a mass of air starting from the ground with a temperature of 50° could be caused to ascend to a height of 200 feet, its temperature would be lower than that of the surrounding air at the same level by an amount equal to KN, which is greater than the difference KM found in the case of the ascent of air to a height of 200 feet through an atmosphere whose conditions are represented by AE. Thus an inversion is an extremely stable condition.

We have so far only considered the ascent of air in discussing Figure 10. When air descends, its temperature rises by 5½° F. per 1,000 feet of descent. Even if the air is originally saturated, the heating due to descent raises its temperature so that it immediately ceases to be saturated, and the rate of increase of temperature with descent is 5½° per 1,000 feet. If the descending air carries with it a supply of water-drops which evaporate as the temperature rises, and so maintains saturation at all stages, the rate of increase of temperature with descent is the saturated adiabatic rate. In all other cases the temperature of descending air rises 5½° for every 1,000 feet of descent.

The First Effects of Direct Sunshine.

We can now return to the subject of Chapter V and consider in further detail the effect of solar radiation on the atmosphere. We shall begin by considering the effect of sunshine in the morning after sunrise; for we can safely assume that the atmosphere will be in a stable condition at the end of the night.

The first effect of sunshine is to heat the ground, which in turn heats the air in contact with it. This process goes on until the lapse-rate attains the adiabatic limit, and probably indeed until it has passed beyond this limit. Any further heating of the air in contact with the ground now produces an unstable condition, and any disturbance must tend to churn up the whole layer through which the instability extends. The motion of air over the ordinary unevenness of the ground is usually sufficient to produce the necessary churning up which brings about a redistribution of heat through the layer concerned, giving in effect a drop of temperature in the lower levels and a rise of temperature in the upper levels. This process is known as the 'convection' of heat, and it must be emphasized that it only begins to operate when the lapse-rate exceeds the adiabatic. While it is obvious that some heat must be conveyed from lower levels to higher levels by radiation, it is generally considered that the major portion of the transfer takes place by convection, i.e. by the ascent of warm air heated by contact with the ground. The convection here visualized may consist in a general churning up through a certain height of the air over a considerable area, or it may consist in the ascent of isolated columns of air over restricted areas, the place of the ascending air being taken by air settling down from higher levels, and pushing in sideways into the funnels of the ascending columns. If the air in an ascending column attains a sufficient height to bring its temperature down below saturation-point there will be condensation of water-vapour into water-drops, and the top of the ascending column will be indicated by the formation of cloud. The cumulus or wool-pack clouds of summer afternoons are examples of clouds formed in this manner.

The air rising in a convection current carries its surplus heat with it, and through mixing with air at higher levels, it produces an increase of temperature at those higher levels. It is thus that the effect of solar heating is conveyed to air which is not in contact with the ground. The sun first heats the ground, which in turn heats the air in contact with it, and this air eventually rises and mixes with air at higher levels, and so conveys the effect of solar heating upwards. This view is borne out by the observations. The range of temperature between day and night is greatest at the ground, and decreases with increasing height, becoming practically negligible at a height of 6,000 feet. The highest temperature during the day occurs earlier at the ground than in the free air, as might be anticipated if we assume that the direct heating occurs at the ground, and is then transmitted upward by convection. If we had available detailed observations of temperature at the ground, and at the top of a tower say 1,000 feet high, we should expect the temperature at the ground to start rising about sunrise, and to rise steadily until the early afternoon, while at the top of the tower the temperature should not begin to rise until the difference between the temperatures at the top and the base had surpassed the limit of 5½° F. (p. 49). It is possible that the effects of radiation and absorption might cause the temperature at the top to start rising before this limit is reached, but no continuous detailed observations have yet become available for testing this.

So far we have restricted the discussion of the transmission of heat upwards to the direct effects of heating of the surface-air by the ground. There is another process to be considered, one which is not readily explained in simple terms. In Chapter III it was mentioned that air is always turbulent on account of friction produced by irregularities at the ground, a current of air being not a steady flow but a collection of whirls and eddies. An eddy may consist of air which is forced into a higher or a lower level by the irregularity of motion, and we have now to consider what will be the effect of the eddies on the transfer of heat upwards or downwards. If the lapse-rate of temperature is greater than the adiabatic, any eddy which moves upward will carry its surplus heat with it, and will therefore contribute to the upward current of heat, and any eddy which moves downward will at its new level be cooler than the air at that level, and by mixing with the air at lower levels will lower the temperature there. So long as the lapse-rate is greater than the adiabatic the effect of the turbulent motion of eddies will be to produce an upward transfer of heat. If, on the other hand, the lapse-rate is less than the adiabatic, any eddy which moves upward will find itself colder than the air at its new level, and so by mixing with the latter will lower its temperature; while any eddy which moves downward will be warmer than the air at its new level, and by mixing with this air will raise its temperature. Thus the effect of turbulence in an atmosphere whose lapse-rate is less than the adiabatic will be to produce a downward flow of heat. If we return to the consideration of the changes of temperature at the top of a tower, we can now say that once the difference in temperature at the top and base of the tower has surpassed the adiabatic limit of 5½° F. per 1,000 feet, eddy motion or turbulent motion will assist the convection of heat upward. But so long as the difference of temperature is less than this limit, the effect of turbulence is to produce a transfer of heat downward from higher levels towards the ground. In Figure 10, p. 51, the direction of flow of heat produced by turbulence is indicated by arrows.

The difference in the changes of temperature from day to night over land and sea were discussed in the last chapter, where it was stated that over the sea the daily variation of temperature is in practice negligible. But it is worthy of remark that the temperature changes which occur in different types of soil vary considerably. The temperatures of (1) a macadam road, (2) bare soil, (3) grassland, (4) bare sand, and (5) woodland soil show enormous differences on a sunny day, with the result that temperature conditions may vary very widely from place to place, both at the ground and at some height above it.[1] These differences may in part account for the turbulence of the air on a sunny day, and are certainly the cause of the 'bumpiness' noted in flying over woods, roads, or rivers.

The amount of the variation in temperature between day and night depends on a large variety of causes, but more particularly on whether the sky is clear or cloudy. When the sky is clear during the day the air temperature near the ground rises steadily during the morning, attaining its highest value about two hours after noon, then falls, at first slowly, then more rapidly towards sunset. If the following night is clear, the loss of heat from the ground by radiation causes a steady fall of temperature until early morning, the lowest temperature recorded being as much as 40° F. below the afternoon maximum temperature. If the day is overcast, little of the sun's radiation penetrates down to the earth's surface, and the temperature remains steady during the whole day. Again, if the night is overcast, the outward radiation from the earth is in part reflected back by the clouds, and in part absorbed at the lower surface of the clouds and re-radiated downwards, so that there is little or no change of temperature during the night. The difference in the diurnal changes of temperature in clear and cloudy weather is so marked that it is easily possible to pick out from a set of temperature (thermograph) records those which refer to overcast weather.

[1] In June 1925 the maximum afternoon temperatures recorded at a station on Salisbury Plain averaged 71·2° F., 108·6° F., 88·0° F., 84·7° F., and 95·1° F., in air, macadam, bare soil, grassland, and sand respectively.

The following table gives the average range of variation of temperature from day to night at Greenwich, for the period 1841–1905:

January	9·4° F.	July	20·9° F.
February	11·1° F.	August	19·7° F.
March	14·8° F.	September	18·2° F.
April	18·2° F.	October	14·3° F.
May	20·3° F.	November	11·2° F.
June	20·8° F.	December	9·3° F.

We append a table giving at the ground and at heights of 3,300 feet and 10,000 feet the mean temperature for each quarter of the year in south-east England:

	Jan.–Mar.	*April–June.*	*July–Sept.*	*Oct.–Dec.*
Ground	38° F.	54° F.	59° F.	45° F.
3,300 ft.	30° F.	43° F.	49° F.	36° F.
10,000 ft.	13° F.	23° F.	33° F.	21° F.

The Heating and Cooling of Air by Motion over the Earth.

When cool air moves over warm land, or warm sea, the lowest layer is warmed by contact with the surface, and heat may be conveyed upward by convection currents or by eddies. In such circumstances the air tends to acquire a distribution of temperature in which the lowest layers are too warm for stability, and if the original difference of temperature is marked, the convection may become sufficiently great to produce rain. Currents of air originating in high latitudes, which we may, in accordance with recent custom, call polar currents, will in the course of their motion southward be warmed by contact with warmer land or sea. It is in such currents that clearing showers in the rear of depressions are formed, being directly due to instability produced by the warming of the air from below by contact with the earth.

The converse process takes place in currents of air originating in low latitudes (equatorial currents), which move northward over colder land or sea, and are cooled from below by contact with the surface. Equatorial currents are therefore stable, and so long as they move horizontally over the surface of the earth they do not produce rain.

The Formation of Rain.

It has been stated above that if an ascending mass of air is cooled below its dew-point, the water-vapour will be partly condensed into water-drops, which will at first remain suspended in the form of cloud, but will fall as rain if the process of condensation is continued for a sufficient time. Practically all rain is formed in this manner, though the ascent of the damp air in which the rain is formed may be due to a variety of causes.

The formation of showers in unstable air is due to ascending currents produced by instability, as has already been stated. Rain is also formed when a warm, damp current of air is forced to climb up over a colder and therefore denser current. Most of the rain which falls during the passage of depressions may be ascribed to this cause. Such rain will not take the form of passing showers, but will be continuous so long as there is a supply of warm air available. We shall return to the consideration of these conditions in the next chapter, in dealing with the nature of the depressions of middle latitudes.

The third outstanding type of rain is produced when a damp current of air is forced to rise over mountain ranges. The most striking illustration of this is the rainfall of India during the South-west Monsoon, when a warm current which has passed over several thousand miles of sea, and has so taken up a plentiful supply of water-vapour, impinges on the mountain ranges and climbs up over them. The rainfall in this case is almost continuous for a period of three to four months, and only ceases when the south-westerly current is cut off. The same phenomena occur on a smaller scale in all mountain districts, such as the Rocky Mountains or the Scandinavian peninsula, and to a less extent in parts of the British Isles.

Inversions.

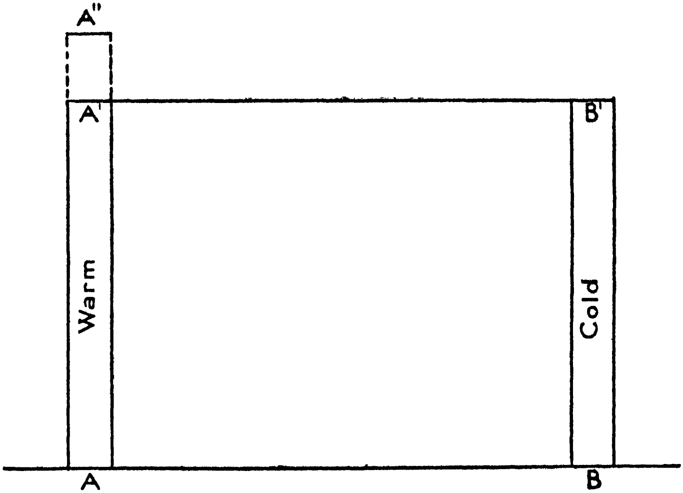
At the beginning of this chapter we stated that the air is in a stable condition if the lapse-rate is less than the adiabatic. The considerations adduced above suffice to show that the smaller the lapse-rate the more stable is the air; and on those occasions when the temperature increases from the ground upward, the air must possess a very high degree of stability. Such a condition of things, with temperature increasing with height, is called an *inversion*. An inversion is most readily produced during a clear night in winter, when the ground cools rapidly by radiation, and cools the air immediately above it to such an extent that it becomes colder than the air at some distance above the ground. Inversions extending up to 5,000 feet are by no means uncommon in winter. Similar inversions are frequently formed at sea by the passage of a relatively warm current of air over a colder sea surface. In each case the cooling is from below by contact with cold land or cold sea. If the air before being cooled is damp, then in the course of cooling its temperature may pass below the dew-point, with the result that some of the water-vapour is condensed into water-drops. These water-drops remain suspended in the air, in the form of fog. It is thus that fogs form during clear nights in winter. The fogs which form over the Great Banks of Newfoundland are similarly explained by the cooling of relatively warm, damp air through contact with the cold surface of the sea. The passage of a warm current of air over cold land can produce fogs in precisely the same way. These fogs are all associated with a very stable condition in the lower layers of the atmosphere, which prevents any churning up of the different layers, so that the cooling is restricted to a shallow layer. Such fogs are therefore usually shallow.

When an inversion occurs from a height of a few hundred feet upwards, with no wind, the smoke of a town rises to the base of the inversion and spreads out horizontally, forming a thick canopy. Such conditions over London produce almost complete darkness, though there may be little or no fog to impede horizontal visibility.

Land and Sea Breezes.

A current of air from sea to land, known as the sea breeze, sets in during the morning, and the reverse, a land breeze, blowing from land to sea, sets in during the evening, in consequence of the daily changes of temperature over the land. In each case there is a return current at no great height flowing from land to sea during the morning and from sea to land in the evening. The formation of these breezes can be explained briefly as follows.

In the diagram of Figure 11, A and B are two points at mean sea-level, and A′, B′ are two points vertically above A and B respectively. Then at a time when the temperature of the air above A and B is the same, if the pressures at A and B are equal, the pressures at A′ and B′ are equal, the weights of columns AA′, BB′, taken with unit cross-section, being equal. If now the air above A is heated, it expands, and lifts the column above it bodily, so that the air which was previously at A′ is now raised to A″. The pressure at A′ will now be greater than the pressure at B′, and a current from A′ to B′ will accordingly set up. This in turn produces an increase in pressure at B, which causes a current from B to A along the surface. The same argument will apply to the case where the temperature above B falls below that at A, and the phenomena can be summarized briefly by saying that along the surface there is a flow from the cooler region to the warmer region, and above there is a flow in the reverse direction.



11. Diagram illustrating the formation of land and sea breezes

We may therefore expect that when the air over the land is rapidly heated by the sun, a current from sea to land will set in at the surface, with a current in the reverse direction at some height above the surface. When the land is rapidly cooled in the evening, the process is reversed, and the surface current is from land to sea, and the upper current from sea to land.

The vertical extent of the land and sea breezes is of the order of 500 to 1,000 feet, and the horizontal extent of the order of a few miles. Since the fall of temperature in the evening is less rapid than the rise during the morning, the land breeze is much weaker than the sea breeze, and it is doubtful whether the return current above the land breeze has ever been observed.

Monsoon Winds.

Over the coasts of most of the continents there is a well-marked alternation of wind from the sea in summer and from the land in winter. The theoretical explanation of these winds is probably similar to that of land and sea breezes, except that in this case it is the annual variation of temperature over the land which provides the motive force. Also, since the annual variation of temperature extends to far greater heights than the diurnal variation, the monsoon winds are much deeper currents than the land and sea breezes, and extend horizontally over much greater distances. The wind current of the South-west Monsoon of India is of the order of 10,000 feet in depth, and flows over several thousand miles of ocean.

There is, however, one marked distinction between the monsoon winds and land and sea breezes, in that the former flow along the isobars, except at the surface, where they drift across the isobars. This is in accordance with the natural tendency of air to flow around the isobars (*vide* p. 64), and the reason why land and sea breezes flow directly from high pressure to low pressure is that they are on such a small scale that the air has not time to attain the balance between the forces at work, which is connoted by flow around the isobars.

VII

The Weather Map

Drawing the Weather Map.

IF for a network of stations there are available observations of pressure, temperature, wind force, wind direction, and weather, all the observations being taken at the same time, we can best obtain an idea of the conditions over the area covered by the network by plotting these observations on a chart. At each station the wind direction is represented by an arrow whose point is at the position of the station on the chart, and the wind force is shown on the Beaufort scale by the number of barbs on the arrow. Alongside the position of the station is written the barometric pressure, below this is written the temperature, and below the temperature are written the Beaufort letters representing the weather. Having plotted the observations in this way for every station, we proceed to draw the isobars, or lines of equal pressure, for any convenient interval of pressure (2 mb in the examples in this chapter). The mean-sea-level pressure at each place on an isobar has the same value. It will generally happen that the pressures at the points of observations are not equal to the exact values selected for drawing isobars, and the isobars are then drawn to pass between the observed values. If, for example, two adjacent stations have pressures of 1015·5 and 1017·4 respectively, we should draw the isobar of 1,016 mb to pass between the two stations, its distance from the second station being about three times its distance from the first station (i.e. as 0·5 to 1·4). To avoid unnecessary crowding, the pressures at individual stations have not been represented on the charts here reproduced, and the isobars only are shown. In the chart shown in Figure 14, the weather is indicated by the Beaufort letters, b representing blue sky, c cloudy sky, o overcast sky, r rain, and d drizzle.

The Relation of Wind to Pressure Distribution.

Having drawn the isobars, we find on examining the chart that there is a very definite relation between the wind and the isobars. The wind blows slightly across the isobars, from high pressure to low pressure, but mainly round the isobars, in such a manner that it has lower pressure to the left than to the right. This law of relation between wind and pressure is the most definite in meteorology. We can go a step farther and add that the wind is strongest where the isobars are most closely crowded together. See, for example, Figure 12, where the wind is much stronger to the south-west of the centre of the depression than to the north-west.

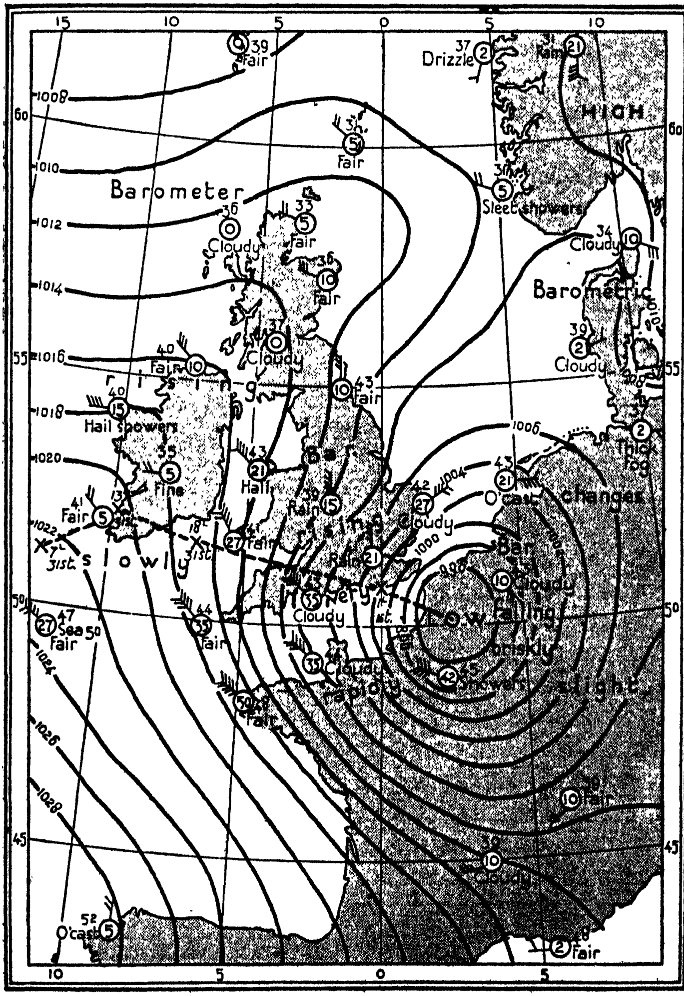
In an examination of the paths followed by surface air currents, Shaw and Lempfert found that currents of air moved over long distances without much change of velocity, and appeared to move under balanced forces. If we assume that the forces acting on a wind current are balanced, we can readily find the relation between the motion and the distribution of pressure.

The forces to which air in motion is subject are of three kinds.

(*a*) The natural tendency of air to flow into the region of lowest pressure, which is measured by the rate of change of pressure horizontally across the isobars, or, in other words, by the *gradient of pressure*;

(*b*) a force due to the rotation of the earth which is directly proportional to the velocity and acts at right angles to the motion; and

(*c*) the effect of friction and turbulence.



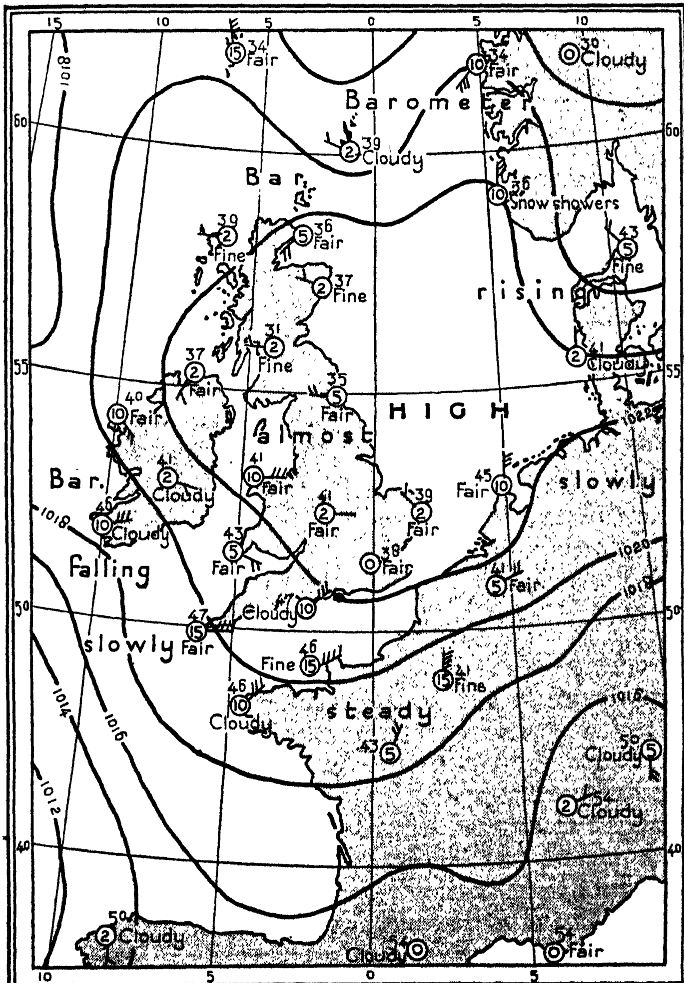
12. The Depression of 1 April 1927

These three forces are balanced by

(*d*) the acceleration of the air, which is at right angles to the direction of flow, and is measured by (velocity)$^2 \div$ radius of curvature of the path which the air follows.

We shall consider first the case where frictional effects at the ground can be neglected, so that our results will apply to the wind at a height of about 1,500 feet above the surface of the earth. Then, neglecting (*c*) above, (*a*), (*b*), and (*d*) must balance each other. Now (*b*) and (*d*) are both at right angles to the direction of motion, and (*a*) is at right angles to the isobars. If these three are to balance, then (*a*) must act along the same line as the other two, or, in other words, the direction of motion is along the isobars. A very simple illustration may make this clear. Imagine a log placed on smooth ice, with two boys pulling it in the same direction. A stronger boy wishing to prevent them from moving the log must pull in exactly the opposite direction, or the log will slip sideways. The wind which, blowing along the isobars, gives a balance between the forces at play, is called the *gradient wind*. The value of the gradient wind which is derived when the curvature of the path is neglected is called the *geostrophic wind*. Thus the geostrophic wind alone would call into play a deviating force exactly counterbalancing the pressure gradient. It is directly proportional in magnitude to the pressure gradient.

We shall not enter into the mathematical discussion of the question raised above. All we are now concerned with is to satisfy ourselves that in the case of steady motion the wind at some distance above the ground blows around the isobars with a velocity which is greatest where the isobars are most closely crowded together. Near the ground the effect of friction is to slow down the wind, so that it flows partly across the isobars from high to low pressure.



13. The Anticyclone of 1 May 1927

Forms of Isobars.

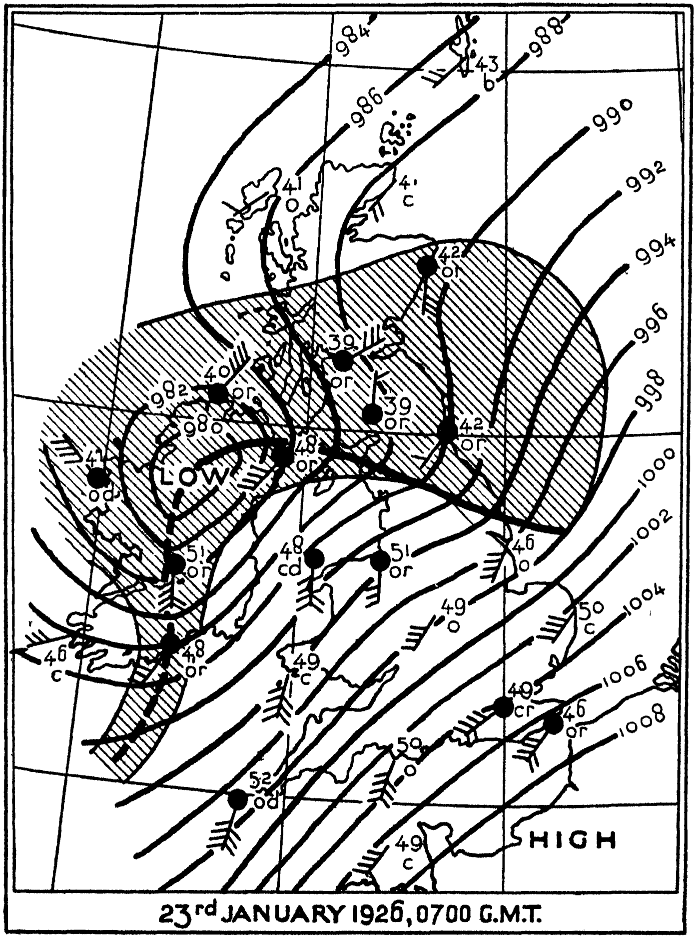
We shall frequently find on our weather charts that the isobars take the form of closed curves, and a family of closed isobars may enclose a centre of high pressure, or of low pressure. The centre of high pressure, with its system of closed isobars, is known as an anticyclone, and the centre of low pressure, with its system of closed isobars, is known as a cyclone or depression. Typical examples are shown in Figures 12 and 13. The distinctive features of any weather chart can be expressed in relatively few words by a statement of the positions of the cyclones and anticyclones which are to be found within the area of the map. These two types of pressure distribution are the fundamental ones, and though there are other so-called types of pressure distribution, they can usually be expressed as regions between two of the fundamental types. For example, a wedge of high pressure is a region between two adjacent cyclones; a *col* is a region between two anticyclones; and a ridge of high pressure is a region between two extensive but not closely adjacent cyclones.

The Anticyclone.

The moving anticyclone is a region of relatively light winds, at any rate in its central part, and while it is usually described as a fine-weather system, it may in practice yield almost any kind of weather, in winter giving fog, rain, or snow almost as readily as blue sky. The motion of anticyclones is usually slow, and irregular. Hanzlik classified anticyclones according to the motion of their centres. He stated that moving anticyclones were cold, and stationary anticyclones warm; that so long as an anticyclone continued to move it remained cold, but that when it became stationary it became warmer, while the pressure rose at its centre.

It is difficult to give a satisfactory theory of the origin of the moving anticyclone of mid latitudes. It appears to be a quiescent mass of air, not so much a part of the circulation of the atmosphere as a dumping-ground for air which has been temporarily withdrawn from that circulation. Exner endeavoured to account for this type of anticyclone as due to the motion of solid currents of air from tropical regions, bringing with them, as it were, a section of the whole atmosphere from sea-level up into the stratosphere. This theory would account for the observed fact that the stratosphere is higher over anticyclones than over cyclones, but it is not clear how a section of the atmosphere can dissociate itself from the remainder, nor must it be forgotten that air can only flow to a particular place and stay there when conditions both along the route and at the stopping-place have been adjusted to receive it.

In the lower layers immediately above the ground there is a general trend of wind outwards across the isobars, tending therefore to annihilate the anticyclone. The effect which the outflow produces is equivalent to a quarrying-out of the lower layers, and a consequent subsidence of the higher layers. Looked at from this point of view, the anticyclone is a region of very slow descending currents, and is thus not an accurate counterpart of the cyclone, in which ascending currents may attain a considerable magnitude. The rate of descent is so slow that during the time air would take to descend through 1,000 feet so much could happen to it by horizontal motion that the actual descent would have no important effect upon the physical condition of the air. Sir Napier Shaw estimates the velocity of descending currents as of the order of 300 feet per day in the Atlantic anticyclone and of from three to five times that amount in smaller anticyclones. This rate of descent is so slow that it is questionable whether we could assume the effects of radiation



14. A depression showing a strongly marked Polar Front, the Rain Area being shown shaded

and absorption to be negligible by comparison with the effect of compression due to descent.

The Depression.

The moving depression or cyclone is in most respects the counterpart of the anticyclone. On the weather map the moving cyclone appears as a region of closed isobars, having the lowest pressure at the centre. The surface winds blow round the centre in a counter-clockwise sense (in the northern hemisphere), with a drift across the isobars from high pressure to low pressure. The depressions of middle latitudes move from WSW. to ENE. on the average, though in individual instances the motion may deviate widely from this. This rate of motion may vary from nothing up to 600 miles a day, and even more in exceptional cases.

The depression is associated with strong winds and, at least in a part of its area, with persistent rain. The heaviest rain is usually in the south-eastern quadrant of a depression moving from west to east.

Earlier writers supposed the cyclone or depression of middle latitudes to have a core which was warmer than the surrounding outer region, but it was first clearly enunciated by the American meteorologist Bigelow, that in the surface-layers depressions cannot be said to have either warm cores or cold cores, the actual conditions being more aptly described by saying that the centres of depressions are usually to be found at a boundary separating warm and cold currents. Thus we should expect any depression shown on our charts to have a warm sector and a cold sector, the line of separation passing through the centre of the depression. While this expectation is realized in a large percentage of the depressions which we find on our daily charts, it often happens that no such line of separation of warm and cold currents is traceable, the distribution of temperature at the ground being roughly symmetrical about the centre of the depression. Again, W. H. Dines found a close correlation between pressure and temperature in the upper air, which would lead us to expect symmetry of temperature about the centre of a depression. The line of separation can, however, be traced upon the weather map in many depressions. Its identification need not commit us to any particular theory as to the nature of the association between the depression and the line of separation. All we need admit at the moment is that there is frequently an association of the centre of depressions with the lines of separation of cold and warm air as shown on our weather maps.

The surface of separation of cold and warm air is known as the *Polar Front.* The example shown in Figure 14 has been specially selected as showing a well-marked *polar front.* Here the line of separation of cold and warm air is shown passing through the centre of the depression. The front or eastern portion of the line separates the cold air from the warm air which is seeking to overtake it. This part, shown in Figure 14 by a continuous line, is called the *warm front.* The rear portion of the line of separation is the boundary where the cold air tends to overtake and to push under the warm air. This is called the *cold front* and is shown in Figure 14 by a broken line. Observations have shown that the cold air at both warm and cold fronts is wedge-shaped, so that at the warm front the warm air tends to run up over the cold air, while at the cold front the warm air is pushed up by the cold air.

Between the two fronts there is a sector of warm air which is continually diminished in extent by the cold front overtaking the warm front, the cold air in the rear pushing under the warm air and lifting it off the ground. When this stage is completed and the whole of the warm air has left the ground, the depression is said to be '*occluded*'. In general, after this stage has been attained, the depression remains stationary, or moves only very slowly, and its intensity diminishes.

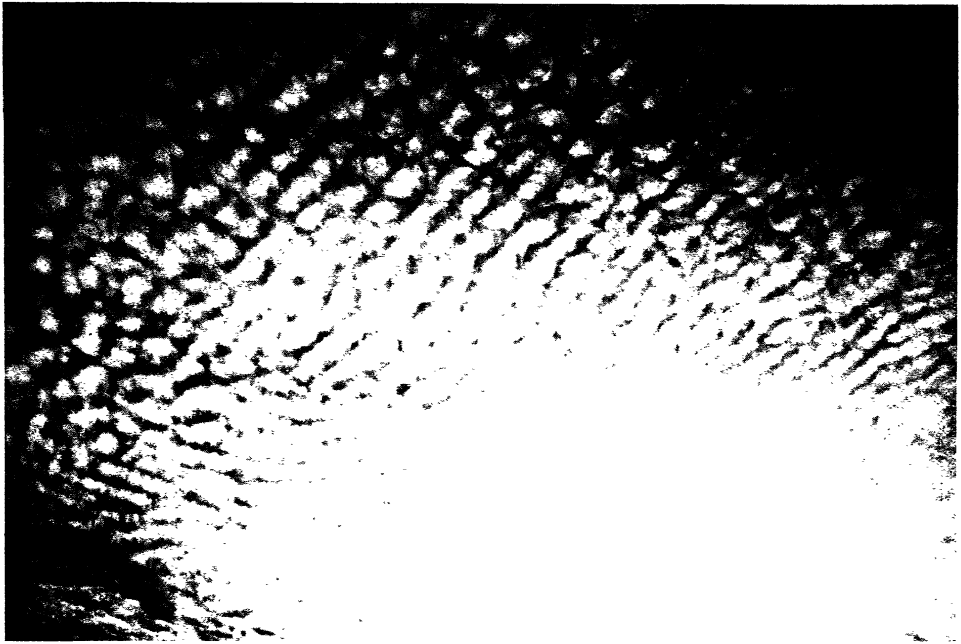
The polar front is marked on the synoptic chart by considerable differences in wind force and direction. Further, the cold air has usually originated in high latitudes, and has low temperature and low relative humidity, generally with good visibility. The warm air has usually originated in lower latitudes, and so has high temperature, high relative humidity, and relatively poor visibility. On account of their difference of origin, the cold and warm air are called 'Polar air' and 'Equatorial air' respectively. The method of analysing synoptic charts by the examination of surface temperatures and winds, outlined in the last few paragraphs, is usually known as the 'Polar front method'. It is sometimes wrongly called the 'Polar front theory', but it is not a *theory* so much as a *description* of the association of weather with different parts of the polar front.

One of the most useful results derived from this method is the establishment of a very well-marked tendency for depressions to move parallel to the directions of the isobars in the warm sector of the depression. Previous to the formulation of this rule, it was customary to deduce the direction of motions of a depression from the rates of fall of the barometer at different stations, the centre of the depression being supposed to move towards the stations where the fall of the barometer was most rapid. The two rules are not necessarily contradictory, but for depressions which move in somewhat irregular paths, the direction deduced from the isobars in the warm sector is the more reliable.

We have already seen that rain is formed by the ascent of damp air. In the picture we have drawn of the cyclone, the warm and cold fronts are the two regions in which warm, damp air is forced to ascend. We should, then, expect to find a definite association of rainfall with these fronts. This expectation is borne out by the facts. Usually the warm front defines the position of a broad band of rain, while the cold front defines the position of a narrower belt of rain. In the main there is not much systematic rainfall in the warm sector, but in the rear of the cold front passing showers are frequent. The latter is in accordance with the views expressed in Chapter VI, since the cold air is supposed to have originated in high latitudes, and to have been warmed by contact with warmer sea on its journey southward. Heating at the surface, if continued sufficiently long, produces instability, and the ascent of the unstable air produces showers.

The weather map shown in Figure 14 depicts a well-marked depression centred over northern Ireland. The temperatures in the south-eastern quadrant of this depression are from 8° to 10° higher than in the remaining quadrants, and it will be readily seen that the region of heavy and continuous rain is that to the north of the warm front, or the part of the line of separation of cold air running eastward from the centre of the depression. There is a well-marked narrow band of rain along the cold front (shown by a dotted line), while in the warm sector only intermittent rain fell. The isobars are drawn nearly straight in the warm sector and do not flow smoothly into the run of the isobars in the cold sectors. The isobars are more closely clustered together in the warm sector, which is the region of strongest surface winds. This particular depression moved in a north-easterly direction, and at 1 p.m. on the same day was centred over the Firth of Forth. Its motion was thus along the direction of the isobars in the warm sector.

The charts shown in Figures 2 and 3 refer to days when depressions passed over the stations. Thus Figure 2 gives the



15. CIRRO-CUMULUS CLOUD

record of wind, velocity, and direction at Spurn Head. At 4 a.m. on 1st April a cold front passed the station, and the wind veered from WSW. to nearly NW. in the course of a few minutes, while the velocity jumped from about 15 miles per hour to about 25 miles per hour, with gusts exceeding 30 miles per hour. Figure 3 shows the passage of this cold front across Eskdalemuir at about 5.20 p.m. on the 31st March. The wind veered from WSW. to nearly W., but the velocity showed no sudden change, and later the wind fell steadily. The third curve in Figure 3 shows the temperatures recorded for the same period, and it will be seen that at the time that the wind veered the temperature fell through about two degrees. The arrival of the cold current was quite sudden and clearly marked. The fourth curve in the same figure gives the total rainfall from 1 p.m. up to each successive time covered by the record. Steady rain fell from 1 p.m. up to 6 p.m. and then practically ceased with the arrival of the cold air. The lowest curve indicates the changes of pressure, marked by a steady fall up to the time of arrival of the cold air, after which there was a steady rise of pressure. The actual arrival of the cold air is marked by a sudden jump upwards of the pressure-trace.

Although we are not here concerned with the formulation of rules for forecasting the weather, we may in passing note that the picture of a depression as a region of conflict of warm and cold air provides a far more convincing picture of the physical processes which are at work than could previously be formed. The picture is naturally incomplete, more especially as regards what happens in the upper air, and so it is not yet possible to forecast with absolute certainty the course of the weather for twenty-four hours ahead.

From time to time instances occur of depressions which at no stage of their history show any association with a contrast of temperature. It is of course possible that a contrast of temperature exists in the upper air, but is completely lacking at the ground. These instances, however, suggest the possibility that the cyclonic depression should not be regarded as a local disease of the atmosphere, as specific as appendicitis, but rather as being of the nature of influenza, one attack of which may bear little resemblance to the last one.

The reader who is interested in the method of analysis of weather charts suggested above is recommended to draw in the warm and cold fronts on such charts as those issued by the Meteorological Office, or even on the small charts published by some of the morning newspapers. He will find it of interest to trace the phenomena which follow the passage of the different parts of the depression. The marked association of steady rainfall with the warm front, of squally winds with the cold front, of poor visibility with warm air, and good visibility with the colder air, can be noted in most depressions. The cyclonic depression presents itself as an alluring subject for study, as it is the most rapidly changing feature to be found on the weather chart.

The central problem of forecasting weather is that of forecasting the future development of the distribution of pressure, which usually means the forecasting of the development and motion of cyclonic depressions. When the pressure distribution is known, the winds can be pretty accurately estimated by the use of Buys-Ballot's law, and the weather for each individual type of pressure distribution can be stated with fair accuracy, so far as its general nature is concerned. The drawing of the polar front, to indicate the regions where contrasts of temperature exist between neighbouring currents, has led to a closer appreciation of the physical changes which take place in the air.

The use of synoptic charts for weather forecasting, as contrasted with the forecasts made by an isolated observer who has only his own observations of wind, pressure, and temperature, together with his observations of the appearance of the sky, is equivalent to increasing the effective horizon of the observer to cover the whole area of his chart. It is not maintained that this is always an advantage. Indeed the difficulty of using a synoptic chart covering a wide area may occasionally be that the forecaster has too much detail shown to enable him to obtain a clear mental picture of the whole, while a keen observer of the local sequence of cloud changes may be able to deduce with considerable accuracy the course of the weather locally for some hours ahead. The ideal conditions for forecasting the details of the weather at a given place would be provided if a keen observer had at his disposal the use of a synoptic chart, and lived largely in the open air. An enthusiastic meteorologist so situated could readily achieve a degree of success in forecasting his local weather which no official meteorologist stationed at a central office could ever hope to achieve. The latter, even when he has very full telegraphic details of the weather for a very wide network of stations, can scarcely hope to appreciate the individual eccentricities of each locality, and even if he could do so, he could scarcely give effect to his appreciation, since he is called upon to sum up his views of the coming weather over a wide area in from twenty to thirty words. This limit does not permit of the description of local peculiarities. The present writer has frequently recommended as a useful exercise for critics, that they should endeavour to describe yesterday's weather at one place in twenty to thirty words.

VIII

Theories of the Origin of Cyclonic Depressions

THE moving cyclonic depression is a region of low pressure, having a system of winds blowing round the centre in a counter-clockwise direction. It is necessary to emphasize the fact that the depression is a region of low pressure. A lowering of pressure over a given region involves the removal of air from above that region, since the pressure at any point is due to the weight of air above that point. If, then, we are to put forward any theory to account for the origin of the depression, we must in the first place find a mechanism which shall be capable of removing masses of air from one region to another in such a way that there is no compensating inflow into the region from which the air is removed.

The most obvious method of removal which we should consider is the horizontal outflow of air from a region. Now it has already been stated (p. 64) that any body or any mass of air moving freely over the earth's surface tends to swing round to the right (in the northern hemisphere). Hence, if air flowed away horizontally outwards from any restricted area, it would acquire a motion to its right in addition to its outward flow. This air would therefore acquire a spin round in a clockwise direction, which is opposite to that in a cyclone. The suggestion of forming a cyclone by this method thus fails completely, and we have to find some other method than purely horizontal motion to explain the formation of a cyclone. The only alternative is vertical motion, and we must first examine whether vertical motion can in any way give rise to a cyclonic depression.

The question suggested at the end of the last paragraph can be most readily considered by limiting ourselves, in the first place, to a case where air is removed from the layer, say 1,000 feet deep, nearest the earth's surface. For the sake of simplicity, we shall suppose the air is originally at rest and that over a small area some air is removed through a non-material vertical chimney, leaving out of consideration for the moment the method of removal. As the air is taken away from the layer in question, air from all sides will stream inward to fill the space which would be left vacant by the removed air. In moving inward the air acquires a motion to the right, and this, as is readily seen, is equivalent to acquiring a motion round the centre of the region in a counter-clockwise direction. The motion is thus in the direction required to produce the system of winds appropriate to a depression, throughout the whole of the 1,000-foot layer next to the ground.

If the depression is to extend to a height of, say, 20,000 feet, then some air must be removed from each successive layer up to 20,000 feet. The axis along which air is removed need not be supposed vertical. It may, so far as we can say at the moment, slope in any direction. But it is worthy of remark that if we are to obtain a wind system blowing in the proper sense (counter-clockwise), round its centre, we must give the air a horizontal drift inward towards that centre.

The centre of the depression formed in the manner which we have described is at rest, and does not move across the map. The moving depression can be formed in the same way, the air which is taken upwards being removed from a particular portion of the current, instead of over a fixed area of the earth's surface. The depression so obtained is a revolving column of air, whose motion will be determined by the motion of the upper current. Except possibly in the lowest layers, the temperature should be nearly symmetrical about the centre. At the surface, currents of air from different latitudes and of different temperatures are brought into the system, and may give rise to sharp surfaces of separation of warm and cold air whose intersection with the earth's surface give lines of discontinuity as described in the last chapter.

Thus far we have not considered what happens to the pressure distribution. If we remove air from low levels to higher levels over the same region, we shall produce no appreciable change of pressure, unless it be an increase in pressure due to the flowing inward of air in the lower levels. We therefore have to postulate some method of getting rid laterally of the air which has been carried upwards. The only plausible method is to suppose the existence of a strong current in the upper air by which the superfluous air is carried away.

It thus appears that two conditions are necessary for the formation of a cyclonic depression: firstly, some physical cause which is sufficient to remove vertically large masses of air, and secondly, a strong upper current which can carry off the ascended air sideways, so as to get rid of sufficient air to produce a lowering of pressure over the region in question.

The converse proposition is true for anticyclones, which are regions of high pressure, in which the wind circulates clockwise round the centre. To account for these we require to have a current in the upper levels bringing up supplies of air, which are allowed to settle down and spread outward. In spreading outward the air acquires a clockwise rotation round the centre on account of the deviation to the right produced by the rotation of the earth.

While several theories have been advanced to account for the ascent of masses of air, little or nothing has been written concerning the possibility of the subsequent horizontal removal by currents in the upper air of the air which has ascended. We shall have to leave the latter aspect of the question in this unsatisfactory state and devote ourselves to the consideration of the vertical removal of air, for which two main theories are available.

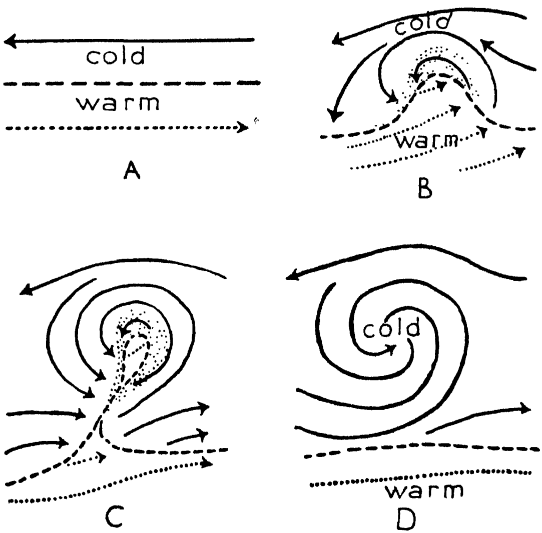
The first theory is the 'local heating' theory, which assumes that the air over a restricted region is heated to a higher temperature than its surroundings, or is heated from below till it becomes unstable (*vide* Chapter VI), so that it becomes capable of rising to very considerable heights. The motive power here is the diminution of density, and this may in part be due to higher temperature, and in part to higher humidity. Suppose the surrounding air has a lapse-rate less than the dry adiabatic, but greater than the saturated adiabatic. Then if some of the surface air becomes saturated and heated above the temperature of its environment, it will ascend, and at succeeding stages of its ascent it will become increasingly warmer than its environment. Thus it is not difficult to visualize the effect of local heating combined with saturation as yielding a vigorous ascent. There is a further point to consider. The ascending mass of air does not retain its heat content intact. The turbulence at its boundary causes partial mixing with some of its environment, and the air from the environment thus drawn into the ambit of the ascending mass will itself become warmer than the normal environment at the same level, and so will ascend.

The problem thus resolves itself into that of finding a physical cause sufficiently powerful to produce the heating required, over a limited area. Over the oceans the effect of solar radiation is slight, as we have already shown in an earlier chapter, and since most of the depressions of middle latitude originate over the ocean, we cannot suppose that these depressions are produced by the intensive local heating of some of the air to a temperature far above that of the surrounding air. On the other hand, currents of air originating in high latitudes and moving southward are heated from below by contact with warmer seas, until an unstable stage is reached, in which large-scale convection readily takes place. This idea is in agreement with observation, since it is not infrequently noted that depressions originate in polar air, with no sign of a line of separation between warm and cold air appearing at the surface.

The alternative to this theory makes the formation of depressions of middle latitudes somewhat analogous to the formation of eddies at the boundaries of a stream flowing into a still millpond. On each side of such a stream eddies are formed, revolving in opposite directions, and showing up as small dimples in the surface of the water.

In the earth's atmosphere the motion of warm currents from low latitudes, and of cold currents from high latitudes, brings into juxtaposition currents of air of widely differing temperatures. Helmholtz showed that such currents could flow side by side, separated by a sharp surface of discontinuity, and that this system could remain stable. Numerous writers have at different times suggested that the depressions of middle latitudes were formed at these surfaces of discontinuity, but it was only during the war of 1914–18 that a serious attempt was made by Norwegian meteorologists to apply this idea in detail to the practice of forecasting. The picture which the Norwegian meteorologists give of the formation of a cyclone starts with a straight line of separation between a wedge-shaped cold easterly current and a warm westerly current. The first step is a bulge of the warm air into the cold air, and as the bulge increases, a well-marked system of closed isobars develops, surrounding a centre of low pressure situated at the tip of the tongue of warm air. The warm air in the tongue impinges sideways on the cold air in front of the tongue, and rises over it; while the cold air in the rear pushes under the warm air, and lifts it off the ground.

The result is to narrow the warm tongue, until eventually all the warm air has been pushed up off the ground. From this stage onward the depression decreases in intensity. This process of development is shown at different stages in Figure 16, where



16. Stages in the development of a depression at the boundary between warm and cold currents

the dotted line is the line of separation of cold and warm air at the surface, the rain area being shown shaded, and the centre of the depression is situated at the point of the tongue of warm air.

Before proceeding farther, we must review very briefly the situation as we now see it. In the course of our brief sketch of the 'local heating', or 'revolving fluid' theory, it was stated that the column of revolving fluid produced by the convergence of air to take the place of the ascending air would, by its action on its surroundings, bring into juxtaposition warm and cold currents of air. The column of revolving fluid was there visualized as the cause, and the line of sharp contrast of temperature as the effect. When we come to consider the practical methods of forecasting developed by the Norwegian school of meteorologists, we find that the line of sharp contrast of temperature, called the *Polar Front*, is visualized as the cause, and the depression as the effect. The polar front method of analysing synoptic charts has been proved to be an extremely valuable aid to forecasting, and we may grant to the exponents of this method our tribute of admiration. At the same time it must be emphasized that there is a wide distinction between a method of analysing charts and a meteorological theory. A polar front *theory* of the origin of depressions should give a description of the physical forces which come into play in order to produce depressions, and in particular should give an explanation of how and why sufficient air is removed to allow for the observed decrease in pressure. Hitherto these points have not received any satisfactory explanation. It is suggested that the air in the warm sector rises over the cold air, but no adequate explanation of this ascent has yet been given.

It is probable that the only way to make advances in this part of the subject is by the detailed examination of individual depressions. Much of the difficulty which has been met in the past is undoubtedly due to the attempt to bring all depressions within the scope of one diagram. It is beyond question that some depressions have originated at polar fronts, where there was a sharp contrast of temperature; and it is equally beyond question that other depressions have arisen in regions where no polar front could be detected. In view of this it appears to the present writer that the most hopeful method of advancing our knowledge of the physical processes which lead to the formation of depressions is, not to attempt to draw one picture of their formation, but to take an individual depression and analyse the phenomena which occur within it. This would obviate the difficulty, which many writers have met, of postulating a reasonable distribution of wind and temperature in their model depression. It is in fact possible that we require not one but two or even more models of depressions.

IX

Other Travelling Disturbances in the Atmosphere

IN discussing the features of the weather map we saw that in middle latitudes the maps were usually of a form determined by the positions of rather slowly moving anticyclones, and of more rapidly moving cyclonic depressions. The latter move in general in the direction of the upper currents in the atmosphere. We thus regard the cyclone as a wind system embedded in the general circulation of the atmosphere, and moving with the speed and direction corresponding to the general circulation. We also find that within the zone of influence of a large depression there may be secondary depressions, which move around the isobars of the main depression.

The features enumerated above do not by any means exhaust the list of phenomena which may be regarded as travelling disturbances in the atmosphere. We shall discuss rather briefly some of the others, viz. the tropical cyclone, the hurricane, tornado, waterspout, and line-squall.

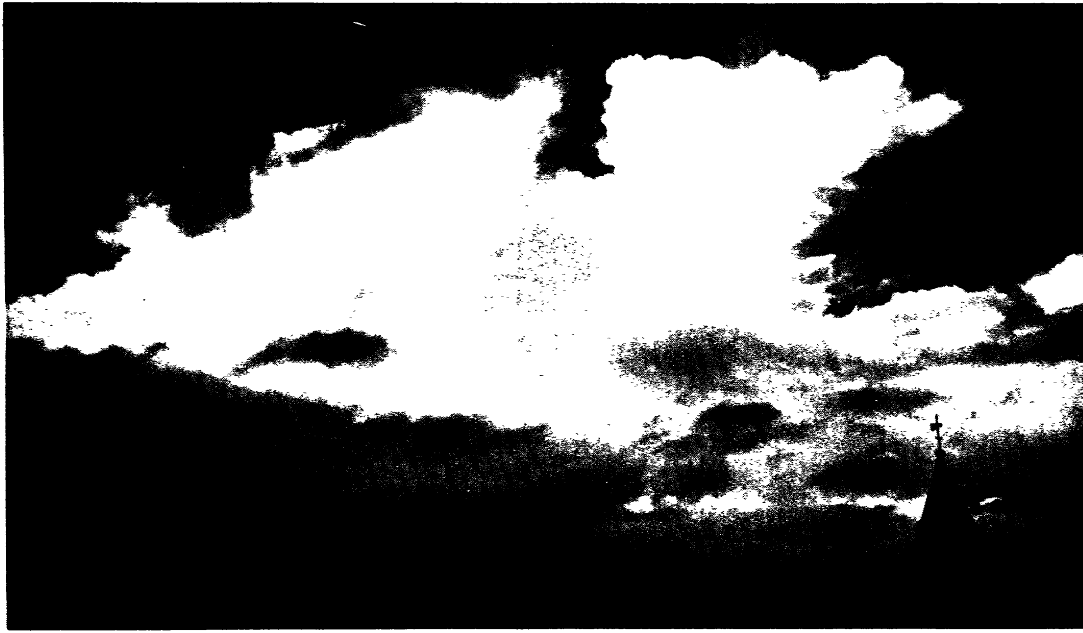
The Tropical Cyclone.

Tropical cyclones are small cyclonic depressions, having nearly circular isobars, and very strong winds, circulating counter-clockwise in the northern hemisphere, clockwise in the southern hemisphere. They originate in the tropics, generally between latitudes 6° N. and 20° N., or between 6° S. and 20° S. They originate over the oceans, and are to be found over all oceans except the South Atlantic. They are variously named in different parts of the world, being known as *Cyclones* in the Indian Ocean and the Mozambique Channel, as *Hurricanes* in the West Indies, and as *Typhoons* in the China Seas.

Most of the regions in which tropical cyclones originate are near the eastern shores of one or other of the continents, and are studded with small islands. They are also regions of conspicuously high sea-surface temperatures. The air in which these cyclones originate has usually travelled over a long stretch of warm sea.

Referring back to Figures 4 and 5, we see that the regions in which tropical storms originate are on the equatorial side of the subtropical anticyclonic belts. When formed, the cyclones travel around the isobars of the anticyclonic belt. The motion is first W. or NW., then through N. to NE., in the northern hemisphere. In the southern hemisphere they move first to W. or SW., then through S. to SE. The path is frequently called 'parabolic', but in general the motion can be quite simply characterized as following the trend of the isobars. It follows that we may regard the tropical cyclone as embedded in the upper current of air, with which it moves. In the Bay of Bengal and the Arabian Sea the tracks are nearly straight.

The tropical cyclone has a strongly marked centre of low pressure. At the centre is a small region of calm or light, variable winds, and surrounding this is a whirl of hurricane winds. The



17. CUMULO-NIMBUS CLOUD

Nimbus Cloud on lower left-hand side of the photograph

horizontal diameter of the cyclone may be anything from 20 miles up to several hundred miles, and the winds may exceed 100 miles per hour. The velocity of travel is usually slow, being about 12 miles per hour. The height to which the tropical cyclone extends is usually estimated at less than 6,000 feet.

At sea the signs of approach of a tropical storm are as follows. The sky becomes covered with a thin cirrus haze, giving red sunsets, and halos or rings about the sun or moon, while a long rolling swell appears on the sea surface. As the cyclone approaches, the barometer begins to fall, the wind rises steadily, and the cirrus haze changes to true cirrus. The rate of fall of the barometer increases, and as the wind becomes very strong the sea is lashed into fury, and torrential rain falls. In the centre or eye of the storm the wind falls to a calm or becomes light and variable, the sea is excessively turbulent, and the sky clear. The 'eye' of the storm has on an average a diameter of 14 miles, but considerable variations occur in different storms. The region of strong winds outside the inner calm are marked by heavy clouds and torrential rain.

The torrential rain in tropical cyclones is in itself evidence of the occurrence of convection of damp air, and this, added to the fact that they are usually found to originate over regions of unusually high sea temperatures, and in currents of air which have travelled for very long distances over tropical seas, points to their formation being due to the convection of heated air. The convergence of air inward to replace the rising air produces the necessary wind circulation in the way we described in Chapter VIII when dealing with the local heating theory. The horizontal removal of the air carried up by the convection currents requires that at some height there shall be a marked change of wind direction, so that the air which rises to this level shall be removed from the column in which it ascended. Thus two conditions are necessary for the formation of a tropical storm: firstly, we require a supply of warm, moist air capable of rising through its environment; and, secondly, we require a marked change of wind at some height in the free air. If the first condition alone were sufficient, it would be difficult to understand why tropical cyclones are not much more frequent. It is the necessity for the second condition which apparently limits their frequency. In the absence of favourable conditions in the upper air, it is probable that the effect of thermal convection is merely to produce a thunderstorm.

The Tornado.

The name Tornado is given to a class of circular storm which is of frequent occurrence in the United States of America, an intense whirl of very small horizontal extent, in which the winds attain a violence far in excess of anything experienced even in the tropical cyclone. The winds in the tornado circulate in a counter-clockwise sense round a centre, with an inflow towards the centre. The velocity of the wind increases towards the centre of the whirl, and it is estimated that wind velocities of as much as 300 miles per hour are attained in some tornadoes. It is only possible to give an estimate of the winds, since they are so violent that no structure of any kind can withstand them, and any recording instrument which happened to be in the path of a tornado would be destroyed completely.

The centre of the whirl, towards which the inflowing air converges, is a region of violent ascending currents, which add their quota to the destructive effect of the whole. The pressure at the centre is reduced considerably below the normal pressure, and this again produces destructive effects, the walls, roofs, or windows of buildings being blown outward as a result of the excess of pressure inside over the outside pressure produced during the near passage of the centre. Heavy rain or hail, accompanied by thunder and lightning, usually occurs away from the central ascending current.

Fortunately the tornado is short-lived, usually persisting for not more than about an hour. The horizontal diameter, or the total width of the path of destruction, is usually less than a quarter of a mile wide, and is frequently much less than this. In Abercrombie's *Weather* there is given a description of the 'Delphos' tornado of 30th May 1879, which did enormous damage and caused some loss of life, although its diameter was only 43 yards.

In the United States tornadoes occur most frequently in the central lowlands. They are formed near the trough or cold front of a V-shaped depression. In the rear of the front the winds are between north and west, with low temperatures, often with rain or snow. In advance of the front the wind is south-westerly, coming as a warm, damp current from the Gulf of Mexico. Thus the cold front is a line of very strongly marked contrast of wind and temperature which produces squally winds and heavy masses of cloud, with frequent thunderstorms. Tornadoes form in the warm air, but near to the boundary between the cold and warm air. The general conditions near the boundary are very disturbed at all times, as is shown by the frequent occurrence of thunderstorms. The formation of tornadoes requires strong localized convection, and this condition appears to be fulfilled most completely in the Mississippi Valley. We do not find tornadoes associated with every cold front passing this region, but when one is formed, conditions appear to favour the formation of a series of tornadoes, which move along essentially parallel paths. This appears to be especially the case in spring. The most disastrous day on record, 19th February 1884, produced no less than sixty tornadoes, leading to a loss of 800 lives, injuring 2,500 persons, and destroying over 10,000 buildings.

The decrease of pressure in the central part of the tornado causes adiabatic cooling of the air, giving condensation of water-vapour. Consequently the core of the tornado most frequently appears as a funnel-shaped pendant below the violently agitated cloud, advancing towards north-east with a speed of from 20 to 40 miles per hour. In some cases the whirl does not reach down to the ground, and it then passes over without causing any damage. When it *does* reach the ground, it leaves nothing standing within its narrow path, whereas objects a few yards outside this narrow zone may remain undisturbed.

The reader who is interested will find in an article by Professor de Courcy Ward in the *Quarterly Journal of the Royal Meteorological Society* for 1917, pp. 317–29, an illustrated description of the effects of a number of tornadoes.

Tornadoes are not always associated with the passage of cold fronts. The *Scientific American* for December 1926 contains a description of tornadoes started over an oil fire which destroyed nearly six million barrels of oil at San Louis Obispo, California. And over the fires which completed the destruction of Tokyo after the earthquake of 2nd September 1923, numbers of small whirls were formed. In these cases the fires produced violent convection, which in turn brought into being the violent whirls of wind.

Tornadoes are not unknown in the British Isles, though they are rare, and of less violence than those which occur in the United States. On 27th October 1913 a severe thunderstorm swept the west of England and Wales from south Devon to Cheshire and developed locally into a violent tornado. Its full violence was only developed along four stretches of its track, one in south Devon, one in Glamorganshire, one in Shropshire, and another in Cheshire. In places the track of destruction was only 50 yards wide, in others 300 yards wide, but the destruction was not comparable with that of the most violent American tornadoes, serious though it may have been in places.

Waterspouts.

When the conditions for the formation of a tornado occur over the sea, they produce what is known as a waterspout, which consists essentially of a small whirl of violent winds whose core is made visible by the condensation of water-drops due to adiabatic cooling produced by the lowering of pressure in the core. Waterspouts are physically the ocean counterpart of tornadoes, being of quite small horizontal dimensions, though the winds within the whirl may attain enormous velocities. They tend to occur at the boundary of warm and cold currents, and, like tornadoes, they may occur in families along a wide front, when they appear as separate pendants from a single squall cloud.

Waterspouts are more frequent in low than in high latitudes, though they are by no means rare even in the British Isles.

Line-squalls.

In the course of the discussion of the origin of thunderstorms and of tornadoes or waterspouts, it has been pointed out that the passage of a cold front may be associated with a variety of phenomena of differing degrees of violence. For example, we have seen that, whereas a cold front moving across the British Isles may produce nothing more serious than a belt of rain, or at most thunderstorms, a cold front moving across the United States may, under suitable conditions, produce a succession of tornadoes of devastating intensity. There is a phenomenon bound up with these called the line-squall, which may be regarded as practically synonymous with the cold front, but which under favourable conditions gives a long roll of cloud extending along the front, and a very strong squall of wind at the onset of the front. Figure 18, illustrating the formation of a thunderstorm, may be used to illustrate this phenomenon. The onset of the cold front is marked by the formation of a cloud which is restricted to the boundary between the ascending and descending currents, this being the roll-cloud of the thunderstorm. The descending cold current sets in as a sudden squall, often attaining gale force, accompanied by a rise of pressure and a sudden fall of temperature. Rain or hail may fall.

The phenomena can at least in part be explained by the form of the wedge of cold air. The friction at the surface of the earth retards the cold air in contact with it, while the cold air at some height above the ground moves forward almost unimpeded. The result is to give the cold air the form of a wedge with its point raised some distance above the ground. Under the overhanging front of cold air there is warm air, and so the overhanging portion is unstable. The breakdown of the unstable arrangement gives the squally winds which are always associated with the passage of a cold front, and probably accounts for the more intense phenomena of the line-squall. It will be noted that essentially the same conditions may produce the rain and squally winds of the passage of a cold front, the line-squall, the thunderstorm of class *c* of Chapter X, as well as tornadoes or waterspouts.

X

Thunderstorms

A FULLY developed thunderstorm may be described as one in which heavy rain or hail, lightning, and thunder occur. All thunderstorms are associated with the formation of towering cumulus clouds, with a more or less level base, and irregular cauliflower-shaped heads, having very sharply defined edges (*vide* Figure 17). They frequently advance across country in a narrow belt moving at right angles to its length. The typical conditions during the onset of a thunderstorm are as follows.

First there will be noted the formation and approach of the thundercloud. The wind freshens, blowing at first towards the advancing storm, while the barometer falls slowly. As the thundercloud arrives overhead, the wind changes in direction, blowing out from the storm in a forward direction, and at about the time when the reversal of wind direction takes place, the barometer rises rapidly through from one to three millibars. For a brief period very strong gusts of wind may occur, falling off as the reversal of direction is completed. While the thundercloud is overhead heavy rain or hail, or a mixture of the two, falls, and thunder and lightning are then most intense. After some fluctuation during the earlier stages the barometer becomes steady, usually at a higher level than before the onset of the storm. The rain may continue for a considerable time after the passage of the more violent front of the storm, but this rain is generally steady as opposed to the intermittent heavy downpour of large drops associated with the front part of the storm.

The first aspect of thunderstorms which we shall consider is the mode of formation of the thundercloud. The essential condition is the production of rising currents of damp air on a sufficient scale. This convection can only arise from the occurrence of instability in the atmosphere on a large scale. The instability may be produced in any one of the following three ways:

(*a*) Strong surface-heating of the air by contact with ground heated by the sun's rays on clear days with light surface winds; or, conversely, cooling of the upper layers over the sea while the surface temperatures remain unchanged.

(*b*) A cold layer of air coming in at high levels, over surface-layers which are relatively very warm, producing a state of things in which the temperature decreases with height at more than the critical rate of 5·5° F. per 1,000 feet (see p. 49).

(*c*) The underrunning of warm, damp air by colder air which causes it to ascend.

We shall consider these in turn.

(*a*) *Convection due to bright sunshine on clear days.*

It is only on clear days, with light winds and a somewhat irregular pressure distribution, that the surface-layers can be sufficiently heated to produce the required degree of instability to cause convection currents on a large scale. The ascending column of air will continue to rise so long as it is surrounded by air of greater density than itself, and when the ascending motion ceases, the rising air spreads out sideways. The rising air will attain very great heights if the upper air is cold, and so thunderstorms formed in this way can occur most readily in early summer (May or June), when the upper air is still cold, though the sun is sufficiently powerful to produce very high temperatures at the ground. It should be noted in passing that at 20,000 feet the air over south-east England attains its highest summer temperature at the end of July or the beginning of August.

The height at which condensation takes place in the ascending air depends on its humidity, but if the air is not originally excessively dry it will attain its dew-point at some stage of its ascent, so forming a cloud by condensation. If there is an upper current of air, the cloud will be carried forward by this current; the indraught of air into the cloud will be from the front, and if this air is warm it will ascend in turn, while colder air from higher levels will descend in the rear. Once the rising and descending currents are established, they will remain in existence so long as there is available a supply of warm air to maintain the ascending current.

Thunderstorms produced by convection in the manner outlined above will naturally occur at those times of day when convection is most active, i.e. on summer afternoons. They can only be so produced over land, as the effect of solar heating is slight over the sea, even on the warmest days. Over the sea, however, there appears to be a well-marked tendency for thunderstorms to occur in the late part of the night. It has been suggested that these storms are due to instability produced, not by the heating of the surface-layers, but by the converse process—the cooling of the upper layers by radiation. It is difficult to decide whether this explanation is a reasonable one, as exact information is lacking as to the rate at which cooling by radiation can proceed in the upper air.

(*b*) *Overrunning of a warm current by a cold current.*

When a very cold current sets in above a warm current, the arrangement is unstable, and any disturbance due to irregularity of pressure distribution may set up violent upward movement of the warm air, which produces a thunderstorm in the same manner as the convection considered in (*a*) above. Over England thunderstorms are produced in this manner when a cool south-westerly wind sets in above a warm current blowing from south to south-east. The heating of the surface currents by bright sunshine is not a necessary feature of storms formed in the manner we are now discussing, and they are as likely to occur by night as by day.

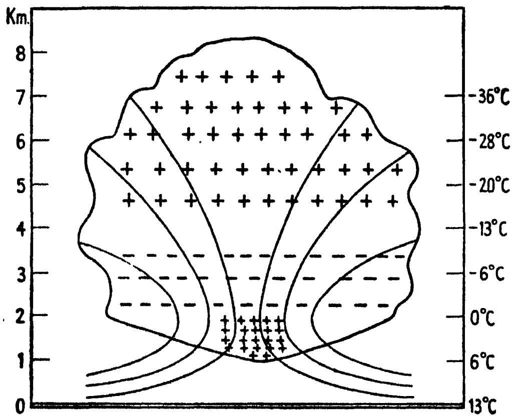
(*c*) *Cold current undercutting a warm current.*

In this case the underrunning cold current forces the lighter warm current to ascend. The effect may only be to produce cloud, or it may in extreme cases yield thunderstorms. The rain associated with the passage of the cold front of a depression is formed in this manner. In late summer and autumn, coastal thunderstorms are frequently produced in this way. They are associated with the passage of the trough or cold front of a depression. Thunderstorms of this class are not usually violent, except occasionally when formed near the centre of a depression. They show no marked preference for day or night, and are as liable to occur in winter as in summer.

Thunderstorms formed by any one of the three processes considered above all have the same fundamental characteristics. They are associated with the formation of heavy cumulus or cumulo-nimbus clouds, with ascending currents in the region of the front of the cloud, and descending currents usually in the rear of the ascending currents. Condensation takes place in the ascending current as soon as the temperature of the ascending air is reduced to the dew-point. The drops of water do not immediately fall out, but are carried upward by the rising air. Small drops of water fall very slowly through still air, and can therefore be carried upward by a relatively slight ascending current, but larger drops require a more intense upward current. The largest drops (of about 4 millimetres in diameter) which can persist in the air, can be kept up by an upward current of 8 metres per second. It has been found that when larger drops than these are formed, they are unstable, and immediately break up into smaller drops. In a thunderstorm the ascending current is not steady, but blows in a succession of gusts and lulls, so that the drops of water in the raincloud rise and fall intermittently, sometimes coalescing into larger drops, and then breaking up again into smaller drops. The drops which get to the edge of the ascending current, or which reach the head of the current and spread out horizontally, fall to the ground, giving the heavy rain of the early passage of the thunderstorm.

The frequent occurrence of hail in thunderstorms affords a striking confirmation of the existence of large ascending currents within the cloud. In severe thunderstorms the base of the cloud is usually from 5,000 to 6,000 feet above the ground, the top of the cloud attaining a height of as much as 20,000 feet in summer and 15,000 feet in winter. In general, therefore, the condensation in the upper part of the cloud will be in the form of snow, and any water-drop which is blown up to the snow-level will be immediately frozen into clear ice, and will gather a coating of snow outside the ice. If in a lull of the ascending current it again falls down to the level where water-drops are found, it will gather a coating of water over the snow, and when it is again carried up into the snow-level the covering of water is frozen to clear ice and a further coating of snow is accumulated. Thus a hailstone which performs a number of journeys from the raindrop-level to the snow-level will for each journey up and down add a covering of clear ice and an outer covering of snow. Actually, hailstones are found to be made up of alternate layers of clear ice and snow, which confirms the suggestion that they are carried up and down in gusts and lulls of the ascending current.

The next point we have to consider is the origin of the electricity in the thunderstorm. In fine weather the electric field in the atmosphere shows that the earth's surface carries a negative charge, and the upper layers of the atmosphere a positive charge. When an electron is removed from a molecule by the action of radiation or some other external agency, the ejected electron and the positively charged residue of the molecule almost immediately attach themselves to one or more uncharged molecules, forming fast-moving 'small ions'. When one of these small ions is captured by a small particle of dust, or by one of the minute droplets of salt solutions which form the nuclei on which condensation of water takes place, the result is the formation of a slow-moving 'large ion'. About two-thirds of the nuclei in the atmosphere are electrically charged, roughly equal num-



18. Distribution of electricity in a thunderstorm

bers carrying positive and negative charges. It will be seen that the intense electric fields in thunderstorms cannot be explained as an intensification of the normal fine weather field.

Fig. 18 shows schematically the distribution of electricity in a typical thunderstorm, as shown by Simpson and Scrase. The curved lines are the lines of flow of the air, and the speed of motion is greatest where these lines are most closely crowded together. The upper part of the cloud, roughly the part above the level at which temperature falls to $-10°$ C., has a positive charge, while the layer of cloud between temperatures of $-10°$ C. and $0°$ C. is negatively charged. Below the negatively charged layer there is, in most thunderstorms, a restricted region of water drops charged with positive electricity, situated above the centre of the ascending current.

Three types of mechanism may be effective in producing the separation of these electric charges. Simpson showed that when a drop of water breaks up into smaller drops, the water drops become positively charged, the air taking a corresponding negative charge. The water drops suspended in the turbulent ascending current in the thunderstorm will undergo frequent collision, with a consequent positive electrification of the water drops. This probably explains the presence of the lower positive charge in the thunderstorm.

Simpson has also put forward a theory, based on experimental work by P. E. Shaw, that when ice particles collide a net negative charge is taken by the ice, while positive ions escape into the air. Thus repeated collisions will result in the ice particles becoming negatively charged, while positive ions escape into the air and are captured or absorbed by the cloud droplets. In a region of generally ascending air the cloud particles, being very small, will be carried upward with the current, while the ice particles fall. Thus across any horizontal plane there will be an upward flow of positive electricity, and a downward flow of negative electricity. Thus the top of the cloud becomes positively charged, while the lower part will contain negatively charged ice crystals. When the ice crystals fall below the level at which the temperature is 0° C., they rapidly melt, and if they fall to the ground unchanged, they will give negatively charged rain. Many of the ice crystals are held up by the ascending current, and so contribute to the negative charge of the lower half of the thundercloud.

The positive charge carried by the water drops has been explained by Wilson as a result of influence of the electric field.

When the normal electric field is reversed, as it is below the negatively charged ice region in a thundercloud, any water drop falling through the field will have a positive charge induced on its upper surface, and a negative charge on the under surface. If the drop remains suspended, or falls more slowly than the downward-moving negative ions, both charges will be neutralized by the ions captured as a result of collisions. If the drop falls more rapidly than the downward-moving negative ions, its under surface, being negatively charged, will repel these ions, but will absorb the upward-moving positive ions, until its charge is neutralized. The upper surface of the drop will retain its charge, since the downward-moving negative ions cannot overtake the drop and neutralize the positive charge. Wilson's theory and Simpson's breaking drop theory both offer an explanation of the lower positive charge in the thundercloud, and it is possible that both mechanisms are active in the production of the positive charge.

Lightning discharges between thundercloud and ground are associated with the formation of a conducting stem of ionized air, which extends downward towards the ground, the field at its tip being sufficiently intense to cause ionization of the air ahead by collisions of electrons and molecules. Flashes generally carry a negative charge to earth.

XI
Some further Notes on the Circulation of the Atmosphere

Motion in the Upper Air.

IN an earlier chapter we presented a picture of the surface-motion of the atmosphere as made up of a background, to which we gave the name of general circulation, on which were superposed local circulations in the form of cyclones and anticyclones. But the description given in Chapter IV only refers to the winds in the lower layers of the atmosphere. It was stated that at heights sufficiently great to be removed from the effects of surface friction, say above 1,500 feet, the winds blow round the isobars, with a velocity which can be determined by the distance apart of adjacent isobars on the map. A comparison of winds at 1,500 feet measured by pilot balloons with estimates made from the weather map shows close agreement between the two. We cannot apply the same method to compute the wind at, say, 10,000 feet, since the shape of the isobars at that height may differ appreciably from that of the isobars at mean sea-level. The reason for the change of form of isobars with height is easily understood when we recall that the pressure measures the weight of a column of air from the level we are considering up to the top of the atmosphere. The pressure at any specified height, say 10,000 feet, will be less than the pressure at the ground by an amount equal to the weight of a column of air from the ground up to 10,000 feet, in other words, by an amount proportional to the density of the air from the ground up to 10,000 feet. The colder the air, the denser it is, and the more rapidly will pressure diminish with height. This result can be directly applied to the earth's atmosphere, since the distribution of temperature over the earth can in the main be described as a steady decrease from equator to pole. If, then, we compare the pressure above the equator and above the pole, at the same height, then the higher we go in the troposphere, the more will the pressure at the pole drop below the pressure at the equator, and at no very great height the result will be to yield a distribution of pressure which is lowest at the pole and highest at the equator. In the upper air, then, the main feature of the distribution of pressure will be a colossal cyclone, centred at the pole, and extending to the equator. The corresponding winds will be everywhere westerly, blowing counter-clockwise around the North Pole in the northern hemisphere. This view is borne out by the observations. Further, maps have been drawn of the pressure distribution at 4 kilometres (13,000 feet) and 8 kilometres (26,000 feet) by Sir Napier Shaw, and are reproduced in *The Air and its Ways*, Plates XIV to XVIII, and these show the polar whirl clearly developed to the exclusion of any other marked features. The tendency to yield westerly winds in the upper air is so definite that in general winds which are easterly at the surface are reversed within the troposphere, having westerly winds above them.

The effect of horizontal differences of temperature, of which we considered above only a special case, namely the normal decrease of temperature from equator to pole, can be expressed in a comparatively simple form. The difference between the winds at any two levels can be measured by the difference between the distribution of pressure at those levels, and this in turn depends only on the horizontal distribution of temperature between the two levels. For the sake of definiteness, let us consider the differences of winds at 5,000 feet and 10,000 feet, supposing the mean temperature from 5,000 feet to 10,000 feet to be known at a number of places over a wide area. If we plot these mean temperatures on a chart and draw isotherms, or lines of equal temperature, on this chart, it can be shown that the wind at 10,000 feet is made up of the wind at 5,000 feet with an added component blowing round the isotherms counter-clockwise round low temperature. Two simple cases may help to make this clear. If low pressure and low temperature go together on our chart, the added component due to temperature distribution is added to the wind at lower level, so that the wind will increase steadily with height. If low pressure is associated with high temperature, the wind at all heights removed from the effects of surface friction will decrease with height, and if the horizontal distribution of temperature remains the same to a sufficient height, the wind will eventually be reversed in direction. Two slightly more complex cases may be added. If the wind at a low level blows in the direction from high to low temperature, the wind direction will veer with increasing height; and, conversely, if the wind blows from low temperature to high temperature, the wind direction will back with increasing height. These simple rules are frequently useful in cloudy weather when observations of winds are not available.

The considerations adduced above show that so long as the existing distribution of temperature persists, with a steady increase from pole to equator, the atmosphere cannot possibly remain at rest relative to the earth. For even if all differences of pressure at the surface were wiped out, the effect of the distribution of temperature would be to produce in the upper air a cyclonic whirl centred at the poles, whose intensity would increase steadily with increasing height, and this would everywhere yield westerly winds, steadily increasing with height.

The Function of the Local Circulations.
In Chapter IV we discussed the general circulation as though it were something entirely independent of the local circulations, and regarded the local circulations as being of the nature of accidents in parts of the general circulation. This view requires readjustment, and we must in fact regard the so-called local circulations as essential parts of the general circulation, for reasons we shall now describe.

The more obliquely the sun's rays fall on the earth's surface, the less will be the quantity of heat falling on, and being absorbed by, a given area of the earth's surface. It follows that the heating effect of the sun is greatest near the equator and least at the poles. During the long polar night the polar regions receive no direct heat from the sun, and if these regions derived their heat only by direct radiation from the sun, their temperatures would drop nearly to absolute zero ($-273°$ C.) during the polar nights. The observed polar temperatures are far in excess of this low value, on account of the effect of the winds in carrying heat from low to high latitudes. Winds from the equator carry heat to high latitudes, and winds from polar regions carry a compensating mass of cold air towards the equator, and the winds thus have a powerful equalizing action on the distribution of temperature over the earth. The transport of heat is carried out partly by the wind system which we called the general circulation, and partly by the local circulations—the cyclones and anticyclones—which produce strong winds from high to low latitudes, and from low to high latitudes.

It has been suggested by Jeffreys that the general circulation of the atmosphere could not maintain itself against the effects of friction in the absence of the local circulations, which have the effect of maintaining motion over the surface of the earth. The effect of friction is to impede motion and slow it down, and this upsets the balance between the forces acting upon the air (see p. 64) and allows the air to drift across the isobars into low pressure. Thus friction at the earth's surface tends to destroy differences of pressure at the surface, and so to destroy the whole circulation of the atmosphere. This action is counteracted by the effects of the cyclones and anticyclones in maintaining motion.

We thus find ourselves forced to the conclusion that the local circulations, so far from being accidents in the general circulation, must be regarded as essential parts of the general circulation. Their effect in redistributing heat over the earth's surface is of course familiar to all who have given any thought to the effect of wind direction on temperature, but Jeffreys' contention that they are necessary in order to maintain the general circulation against friction is a novel one.

The Nature of Local Circulations.

We have regarded the motion of the earth's atmosphere as consisting of a general circulation which formed the background to the somewhat irregularly occurring local circulations; but it is now necessary to go back a step farther and to realize that all motion relative to the earth is superposed on a background of rotation with the earth. The earth rotates upon its axis once in a day, and the main motion of the atmosphere is a rotation about the polar axis once in a day. Where the motion of the air deviates from this value we observe the deviation as wind, but so long as the motion of the air does not deviate from the rotational motion we do not think of it as moving. For this reason we tend to overlook the fact that the main store of kinetic energy possessed by the earth's atmosphere is derived from its rotation as a solid. Let us for a moment leave out of consideration all motion relative to the earth, and think only of the solid rotation of the atmosphere about the polar axis of the earth. Our visible horizon may be regarded as a circular disk, which rotates about the vertical in a counter-clockwise sense, giving the sun and stars the appearance of moving around the horizon in a clockwise sense. The air above the disk has the same counter-clockwise spin as the disk itself, and this counter-clockwise spin of the air is in reality the raw material from which all circulations are made. If from the middle of the disk we take away air, removing it vertically, the air which flows in to take its place will preserve its original angular momentum, and will accordingly spin round faster as it approaches the centre in the same way that water going out of a bath spins faster as it approaches the centre of the whirl over the waste-pipe. Air which has in this way approached nearer to the centre will therefore spin round faster than the horizon, though always in the same direction. It will therefore move round counter-clockwise relative to the horizon. If on the other hand we bring in air at a high level and let it descend and spread out over the horizon, it will push outward the original air over the disk, so producing the reverse to what we considered above, a decrease of the rate of spin of the air. In this case the result will be to cause the air to go round counter-clockwise slower than the horizon, so that to an observer on the earth, and consequently going round with his horizon, the air will appear to be continually being left behind, and so will move clockwise relative to the earth. The two cases of removal from and addition to the air originally over the disk to which we assimilated our horizon are respectively the lowering and raising of pressure over the region. In the first case, that of lowering of pressure, we saw that the wind circulation was counter-clockwise over the earth. Moreover, there is no obvious limit to the possible magnitude of this circulation, since it goes on increasing so long as the removal of air is continued. In the second case, that of addition of air to that originally over our horizon, the original counter-clockwise spin with the earth was decreased, so that the air went round slower than the horizon, and the wind circulation, which is the motion of the air relative to the earth, is clockwise about the centre. In this case, however, there is an obvious limit to the intensity of the circulation. For by addition of air at the centre the most we can do is to decrease to zero the original spin of the disk of air, since it is not possible by this means to reverse the direction in space of this spin. In other words, the most we can do is to cause the air to stand still while the horizon slips round below it, and in that case the air would circulate clockwise relative to the horizon with the same angular velocity with which the horizon spins counter-clockwise. This view of things accounts for the observed fact that cyclones sometimes have a rate of rotation relative to the earth of three or four times that of the horizon about the vertical, and tropical cyclones have a rate of rotation far in excess of this, whereas no anticyclone ever has a rate of rotation clockwise exceeding the rate of rotation of the horizon about the vertical. It usually has a rate of rotation of about half the limiting value.

We are therefore forced to the conclusion that the cyclones and anticyclones in the atmosphere are not excrescences on the general circulation so much as excrescences on the solid rotation of the whole atmosphere, deriving their energy from the solid rotation rather than from the winds which prevailed locally before their formation.

The rather lengthy discussion above is the answer to a question which is frequently asked by the curious: 'But why do cyclones go round in the way they do?' And the answer can be boiled down to a very few words: 'Because the earth goes round that way.'

Variations in the Distribution of Pressure, Temperature, and Rainfall.

A record of pressure at any one place shows fluctuations through a wide range, some of the variations having the appearance of periodic variations over a short interval of time, then suddenly losing all semblance of periodicity. The most thorough investigation of the variations of pressure, temperature, and rainfall in different parts of the world was carried out by G. T. Walker in an endeavour to discover arithmetical relationships between the rainfall of the Indian summer monsoon and other factors for different parts of the world. Walker used the method of 'correlation', and computed some thousands of coefficients of correlation between various factors of weather over different parts of the world. If we regard the phenomena of weather as being eventually due to the effect of solar radiation, we should obviously first consider the possibility of accounting for variations in world weather by variations in the solar constant. Walker showed that the observed variations in the 'solar constant' were insufficient to account for the variations of monsoon rainfall, and so came to the conclusion that the causes of variation of the monsoon rainfall should be sought in the earth's atmosphere. An extreme example helps to show the insufficiency of the solar variation as a determining cause. The monsoon rainfall in 1917 was the highest ever recorded, while that of 1918 was far below normal, but there was no abnormal variation in solar radiation in those years.

The main result which Walker's labours bring out is the existence of a swaying of pressure backwards and forwards between the Indian and Pacific Oceans, and of smaller-scale swayings between the Azores and Iceland and between the areas of high and low pressures in the North Pacific. The formula which Walker gave for forecasting the monsoon rainfall from the variations of pressure, temperature, and rainfall at other places distributed over the earth has, however, met with only a modified success.

Whether the variations of weather at different places can be regarded as due to oscillations of the general circulation of the atmosphere is doubtful, particularly if we use the word oscillation in its normal sense. It is of course possible that the variations are of the nature of oscillations which after a time die away, and are later brought back into existence with a different *phase*, i.e. with times of maximum effect not fitting into the series of times of maximum of the earlier existence.

A vast amount of labour has been devoted by different workers to an effort to elucidate the possible oscillations present in the variations of pressure, temperature, and rainfall. The results, at least in so far as the British Isles are concerned, are of no use for forecasting the future course of the weather. The oscillations of temperature, pressure, or rainfall at stations in the British Isles are all so small in magnitude that their effect is negligible by comparison with the apparently casual day-to-day variations. Indeed many writers have maintained that the apparent periodic oscillations have no physical reality. We need not enter here into the question of the reality of the apparent oscillations, since it is unquestionable that they are of no use for forecasting the weather for the future. By far the greater part of the variations in temperature, pressure, and rainfall over the British Isles is of an apparently casual nature, following no known law. The claims of various writers that they can forecast the weather for long distances ahead are not justified by our present knowledge of the nature of the variations of weather.

The Variation of Climate.

During geological times there have been considerable variations of climate over widely distributed regions of the earth. To take the example nearest to us, the climate of the region which is now the British Isles must have been very much milder than it now is at the time when the coal measures were laid down, since the plants which formed these measures are of types which only grow in mild or even tropical climates. There is also well-authenticated evidence that in India and Northern Australia, within 20° of the equator, there was once glacier ice formed at sea-level. Thus we are faced with evidence of much milder climates in high latitudes at one stage of the earth's history, and at another stage an ice-age in the tropics. We have not the space available here to enter into detailed discussion of these problems, and they are only mentioned here because they are to some extent bound up with our views of the general circulation of the atmosphere.

The intensity of the general circulation of the atmosphere may be regarded as being roughly proportional to the difference between the mean temperatures at the equator and the poles. It has, however, been shown above that the temperatures in high latitudes would be much lower were it not for the effect of the wind in transmitting heat from low to high latitudes. It thus appears that the existing state of things represents a balance between the temperature distribution and the general circulation. If the general circulation were intensified, its effect in transporting an increased amount of heat to high latitudes would diminish the temperature difference between equator and pole, which in turn would cause the intensity of the general circulation to drop back and eventually to regain its balance. The actual point at which the balance is attained must depend upon a variety of factors, in particular upon the intensity of solar radiation, the nature of the earth's surface, the distribution of water-vapour in the atmosphere, and the distribution of ocean currents. Any factor which transports heat from the tropics to high latitudes must affect the exact balance between wind and temperature. Next to the winds, the ocean currents are the main carriers of heat, and we have a familiar example in the Gulf Stream which gives to the Scandinavian Peninsula a temperature 10° C. higher than the mean temperature of the same latitudes over the remainder of the earth. But the main ocean currents of the globe owe their origin to the winds, so that the amount of heat transported by the ocean currents is at least approximately proportional to the winds. The effects of ocean currents are of a lower order than the effects of heat-transport by the winds, which are mainly responsible for the occurrence of polar temperatures far in excess of those which should occur if the distribution of heat over the earth depended upon direct solar radiation only.

The problem which we have raised here, of the variation of climate, is one which has produced a vast amount of literature, much of which is of little permanent value. We have not the space to discuss the problem here, and it is merely stated in order to show the reader that there still remain unsolved questions which involve the consideration of all the physical factors pertaining to the earth, the ocean, the air, and solar radiation.

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