DAMAGE BOOK

THE RESTLESS ATMOSPHERE

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GEOGRAPHY

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THE RESTLESS ATMOSPHERE

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PREFACE

THIS brief study was written to fill a gap in the literature of physical geography that caused a good deal of trouble to the author when he was a student. There are many good text-books of meteorology and climatology on the market, and there is no point in adding another which covers the same fields. The author's aim has been to write chiefly in the no-man's-land of dynamic climatology, where the ideas of the modern weather analyst are put to good use in explaining regional climates. The latter have been described a dozen times over by geographical climatologists like W. Köppen, J. Hann, W. G. Kendrew and R. de C. Ward, and the descriptions have often been made partly explanatory, as any good descriptions should. But most of the stock "explanations" of regional climates need a thorough overhauling. They are generally statical in character; if they take in the circulation of the atmosphere at all, they refer only to the "general circulation" or the "centres of action", or other generalizations of a decidedly rough-and-ready sort. The present book tries to go one step further by describing the actual day-to-day weather processes involved in the principal types of world climate. Only one other book, B. Haurwitz' and J. A. Austin's "Climatology" is written in the same vein.

The earlier chapters include a short discussion of the processes whereby weather is made. Particular emphasis is laid upon rain-making processes, on airmass theory and on the history of frontal cyclones, as these are the fields in which text-books of climatology are usually weakest. The treatment is not mathematical: the atmosphere is looked at geographically, though some elementary physics is unavoidable. There are only two differential equations, and if the student cannot understand them he can pass them by; since, however, they express two of the most useful rules in meteorology, they have been included in the hope that they will be comprehended by some.

The second half of the book is a regional treatment of climates on a continental scale. Since many parts of the world have only been opened up to the activities of the weather analyst for a very short time, much of the material in these regional chapters is controversial. Many meteorologists with experience in the regions concerned will differ from the author. Their opinion will be most welcome if they can find time to write. The regions treated in detail are those usually covered by university courses in regional geography for honours students in Britain and North America. Other regions receive only passing mention, as space is limited. One sizable omission is the temperate belt of the southern hemisphere; here the author has nothing to add to what has already been written. Another omission is the sub-tropical desert belt of both hemispheres; changes in day-to-day weather mean so little in desert climates that dynamic climatology is without value except to the pure climatologist. In these later chapters the author acquired most of his ideas while working in the Investigations Branch of the Meteorological Office, Air Ministry. He would like to express his gratitude to Mr. C. S. Durst, head of that Branch, for the searching education he underwent in these years.

The book ends with a review of the general literature of meteorology and climatology, so that the student can read further if he wishes. References are given at points in the text to certain fundamental source materials, which should be followed up if the reader wants a more thorough treatment.

The author's thanks are due to the Editors, to Mr. Richmond W. Longley, Professor G. H. T. Kimble and many other kind friends for reading parts of the book.

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CHAPTER I

THE ENERGY OF THE ATMOSPHERE

ENERGY exists in the atmosphere in various forms. The word is used in a way that may be strange to those who have not met the jargon of physicists. In physics, a body's energy is defined as its capacity to do work. Furthermore, the physicist extends the definition to cover things which do not at first seem to be related to work at all. Light, for example, and radiant heat are as truly forms of energy as is the motion of an aircraft in flight. In the atmosphere we can distinguish energy in four classes:

(1) *Radiant energy*, or simply "radiation". This is energy which travels through space in wave-like disturbances of something it used to be fashionable to call the "ether". Light, radiant heat, the energy of a radio beam, are all differing forms of radiant energy. They can pass through empty space, like the universe outside the atmosphere, and can often pass through material substances like the air itself.

(2) *Kinetic energy*, which is the sense in which we are apt to use the word. It is the obvious energy of objects in motion; the wind, for example, is air possessed of kinetic energy. In the atmosphere much of the kinetic energy is that of the great moving airstreams which traverse the globe, but a substantial fraction is used up in the innumerable turbulent eddies which develop in moving air.

(3) *Potential energy* of gravity. Since the air is permanently subject to the gravitational pull of the earth, it always has the capacity to move downwards unless checked by some opposing force. This bottled-up energy is called "potential" energy.

(4) *Thermal or "Internal" Energy*. Heat is also a form of energy. A hot body possesses a store of energy which

enables it to communicate heat to surrounding cooler bodies. Hot bodies are those in which the molecules are in violent motion.

It is a law of classical physics that all these forms of energy are indestructible, though energy can readily be converted from one form to another; when light falls on the surface of a road, for example, it is readily converted into heat, and the road surface becomes hot. This change of form does not involve any loss of energy; the heat produced is equivalent to the light which has vanished, and could in theory be converted back into the same amount of light. In just the same way heat is constantly being converted into kinetic energy, gain without any loss. It is plain, therefore, that we can prepare a strict budget for the atmosphere's energy, since we know that any loss from one part of the atmosphere must be compensated by an equal gain elsewhere.

The Source of Atmospheric Energy

As far as we are able to measure, the energy of the atmosphere remains constant year after year; there are no obvious signs that its circulation is slowing down or getting more vigorous. Yet we know that every year the earth pours out to space a constant stream of radiant energy which, once outside the atmosphere, is entirely lost to us. This annual drain must plainly be made up from somewhere, or the constancy of energy which we have just laid down could not be maintained.

There are two possible sources of supply:

(i) Terrestrial heat, or energy which escapes from the hot interior of the earth. The crust contains many radioactive minerals which, in their decay, release great quantities of heat. This energy must slowly escape to the atmosphere, or the core of the earth would get hotter than we think it to be. It is certain, however, that this flow is negligible by comparison with that from the sun, and we usually neglect it completely.

(ii) Solar radiation, or insolation. The sun pours out a tremendous quantity of radiant energy at an almost constant rate. The earth intercepts a tiny fraction of this radiation, and it is from this source that the atmosphere gets almost all its energy. Receipts from the sun during the year just balance the earth's export of radiation to space, so that the atmosphere neither gains nor loses in energy over the twelve months. Before a unit of energy escapes back to space, however, it is retained and used by the atmosphere for a fairly long period in the kinetic or thermal form. It is from this reservoir of energy that the atmosphere draws the energy of its circulation.

The Energy-Budget of the Atmosphere as a Whole

The incoming beam of insolation, as we call the radiation from the sun, consists in part of light (which is the part our eyes can detect) and in part of other types of radiation which we call ultra-violet and infra-red. The three types are all radiation of the wave-type; they differ only in wavelength, that is, in the distance from crest to crest of adjoining waves. The solar beam consists of radiation with wavelength between about 0.3 thousandths and 3.0 thousandths of a millimetre, whereas our eyes can see only rays with wavelengths between 0.4 and 0.7 thousandths of a millimetre. The ultra-violet rays are those shorter in wavelength than visible light, and the infra-red are those which are longer. The sun's rays are most intense in the range of wavelengths of visible light, so that we can see more of the incoming energy than these figures would suggest. The ultra-violet rays affect our skin and cause sun-burn. Infra-red radiation has little effect on our bodies, and is familiar to us through its use in photography; film-emulsions can be made which are sensitive to the infra-red, and since these rays will penetrate dust and fog, they permit long-range photography.

After entering the atmosphere, the beam gets a mixed reception. Almost half of it is reflected back to space without change, chiefly from the tops of clouds and from snow, which are good reflectors. This reflected light has no effect on the atmosphere. The rest, however, passes down towards the earth, and we have to trace its course in detail. Some quantitative estimates have been made of the history of the incoming energy, but here our balance sheet can be very rough and ready. The units employed below are based on the assumption that the solar radiation received at the outer boundary of the atmosphere in a year equals 100, and we express all the others in this unitary system; all figures are hence percentages of the total energy reaching the atmosphere from the sun, which is almost a constant quantity.

The chief exchanges of insolation are these:

(1) Of the incoming 100 per cent of solar energy, 33 per cent is reflected directly back to space from clouds, snow or other reflecting surfaces, and is thus at once lost to us. A further 26 per cent is intercepted by the air itself and "scattered"; this is a form of reflection affecting the very short wavelengths in the solar beam. About 10 per cent is sent back to space in this way, but the other 16 per cent continues downwards as the "diffuse" light visible to us as "sky-light".

(2) 14 per cent of the solar beam is intercepted by the ozone, water-vapour, dust and clouds of the atmosphere, and "absorbed", that is converted to heat. Little is thus used to heat the atmosphere directly.

(3) The remaining 27 per cent and the 16 per cent of diffuse daylight referred to in (1) reach the earth, and are used to heat the earth. Plainly, therefore, *the atmosphere is heated from below*, for the absorption of insolation by the earth is three times that absorbed by the atmosphere itself.

The 43 per cent absorbed by the earth is ultimately returned to space, but the process of transfer is very complex. Its main features are summarized as follows:

(1) The earth communicates 2 per cent to the atmosphere by direct conduction of heat, and 18 per cent by the transfer upwards in convection currents of water-vapour evaporated from the earth's surface. Such water-vapour contains much latent heat abstracted from the evaporating surface.

(2) The earth radiates a large quantity of long-wave radiant heat. Unlike solar radiation, such radiation is rapidly absorbed by water and water-vapour, so that very little can

escape through the atmosphere. Only 7 per cent gets out to space direct. The remainder is absorbed by the water-vapour or clouds in the atmosphere, and re-radiated. Of the re-radiated energy, most gets back to the earth, and only a small part escapes to space. Once back at the earth's surface, it is absorbed, converted back to heat and once more radiated. In this manner a remarkable exchange of heat is kept up between the earth and the atmosphere; some 112 percentage units are radiated from earth to air, and 96 back from air to earth. The seeming paradox that the earth radiates more than the total strength of the sun is explained by the fact that most of it comes back again and is drawn to and fro between earth and atmosphere like a shuttlecock. A large quantity of heat is thus retained in the lower atmosphere. This process is sometimes referred to as the *greenhouse effect*, with clouds and water-vapour taking the place of the glass. As a result of these repeated exchanges, a net amount of only 16 per cent (112 less 96) is finally transferred to the atmosphere.

(3) The final exchange necessary is the transfer of radiant heat from the atmosphere to space. This amounts to 50 percentage units, which is added to the much smaller quantity of radiant heat escaping from the earth direct.

This balance sheet is a generalized result which refers to the entire earth for an entire year: at any special place or moment the exchanges may be very different. Moreover, the figures quoted rest on a flimsy theoretical basis, and will certainly be revised; it has already been suggested that the figure of 43 per cent for the loss by reflection and scattering is much too low.

Variation of the Heat Budget with Latitude

Over the year as a whole the poles receive much less heat from the sun than does the equatorial belt: we should therefore expect a difference in temperature between the two zones, which is certainly true of today and recent geological times. It is possible to apply the arguments put forward in the preceding section to each latitude. Thus G. C. Simpson (1928) has computed both the gain from the sun and loss of radiation to space from the equator to the pole. His results are summarized in fig. 3.

Simpson's results indicate that the polar "caps", extending from latitudes 30-35°N. and S. to the pole, have an annual heat deficit of considerable proportions: half the earth's surface actually suffers a net radiative loss of heat over the year, including nearly all the areas of dense settlement. The equatorial belt (between 35°N. and 35°S.) on the other hand, has a net gain, since receipts from the sun exceed the loss by radiation to space. It might appear, therefore, that the equatorial belt should be getting warmer and the polar caps colder every year. These changes are prevented by the circulation of the atmosphere, which every year bodily carries the equatorial excess northward into the area of deficit, and so brings about a balance in all latitudes.

Variations in Temperature with Height

The vertical distribution of temperature is a topic of great importance in meteorology, since upon it depends the extent to which cloud and precipitation can occur. By means of sounding balloons, the use of aircraft and certain other methods, the properties and behaviour of the upper atmosphere (traditionally called the "upper air") have been well explored in recent years. From this exploration has emerged a body of facts and a great weight of theory and experimental methods. The climatologist cannot afford to neglect either the facts, the theory, or the methods, for all of them contribute largely to an understanding of world climates.

Since the atmosphere is heated mainly from below, it is reasonable to suppose that temperatures should be highest at low levels, and should decrease with height. This is observed to be true of the lower atmosphere: it is common knowledge that the air gets colder as one climbs a mountain. Fairly precise measurements of this decrease have been made for many years by sounding balloons, which carry recording thermometers high into the atmosphere, preserving a continuous record of temperature changes. Nowadays the sounding balloon (or *radiosonde*) carries a small radio transmitter which sends back signals enabling receiving stations at the ground to compute the temperature at any height. Such soundings regularly penetrate to 50,000 ft. above the sea, so that by their use the observer is quickly able to get a picture of the thermal condition of a large part of the atmosphere.

The rate of decrease of temperature with height is known as the *lapse rate of temperature*, or more often just the lapse rate. The latter varies very widely from point to point, and also from moment to moment at each point. It is usually expressed in Great Britain in degrees Fahrenheit per thousand feet. It sometimes goes by the more precise title of the environmental lapse rate, to distinguish it from other unrelated lapse rates with which we shall deal later. Since the lapse rate is defined as a fall of temperature with increasing height, a fall of temperature is regarded as being of positive sign; e.g. a lapse rate of +3.5°/1,000 ft. means a fall of 3.5° per thousand ft., and not a rise as the sign seems to imply. In certain cases it is found that temperature actually rises with height; we then speak of an "inverted" lapse rate, to which we assign a minus sign, e.g. —2.0°F/1,000 ft.

Almost all upper air soundings exhibit certain features in common:

(1) The lower half of each curve shows a fall of temperature with height. This lower part of the atmosphere characterized by positive lapse rates is called the *troposphere.*
(2) The upper half of each curve is of fundamentally different form. Temperature is almost constant, and in some cases even shows a very gentle rise with height. The constancy of temperature extends as high as most soundings have been taken. This upper atmosphere of constant temperature is called the *stratosphere.*
(3) The point at which the curve passes from the troposphere to the stratosphere is called the tropopause, which is one of the most important boundaries in the atmosphere. Below it is the troposphere, which contains nearly all the water-vapour, the clouds and the storms of the atmosphere; the stratosphere is clear, cold, dry and virtually cloudless at all times, being totally unaffected by storms or visible disturbances.

This twofold division of the atmosphere into troposphere and stratosphere separated by a clear-cut tropopause is world-wide and virtually permanent. Wherever we take our soundings we find the same fundamental structure; however much the properties of the troposphere may vary, there is never much difficulty in separating the two zones. The tropopause is observed, however, to vary considerably in level, so that the relative thickness of troposphere and stratosphere varies. Over south-east England the tropopause oscillates in level between about 20,000 ft. and 40,000 ft., being in general lowest (i) with low temperatures, (ii) with low pressures, and (iii) in winter. The tropopause is about twice as high over the equator as over the pole. Latitudinal cross-sections also reveal the remarkable fact that the stratosphere is colder over the equatorial belt than over the poles. The lowest temperatures ever recorded on earth were encountered at heights of about nine miles (48,000 ft.) over Batavia, in Java. Above about fifteen miles it is believed that temperatures are uniform all over the earth. At still greater heights it is possible that there are layers warmer than at ground level, and recent tests by rocket in New Mexico suggest that these high-level layers may be very hot indeed.

To what extent conditions in the stratosphere affect the troposphere we cannot say. J. Bjerknes (1937, 1951) has in recent years developed a theory of cyclone-formation which involves disturbances of the tropopause, and presumably of the stratosphere also, but beyond this point the stratosphere remains very much of an enigma. Its very existence has not yet been satisfactorily explained, though it is plainly related in some way to the radiative behaviour of the lower atmosphere.

Inverted lapse rates within the troposphere may develop by several different processes. The commonest are those which develop near the ground on clear, calm nights. The radiative cooling of the ground chills the air above, sometimes until it is 15°F. or 20°F. cooler than at 1,000 or 2,000 ft. above the ground. By damping down convection currents, such "inversions", as they are called, cause the accumulation of smoke in the surface air in cities. These occur on nearly all clear nights all through the year, but tend to be strongest in winter. Inversions may also develop at higher levels when general subsidence of the air is in progress. A third type occurs when warm air overruns cooler air below, as along frontal surfaces.

CHAPTER II

MOISTURE IN THE ATMOSPHERE: CLOUDS AND RAIN

Composition of the air. Though the air is a mixture of gases, its composition is remarkably constant up to great heights in the atmosphere. Most of it consists of the two common gases: nitrogen and oxygen. The latter is the part of the air necessary to most forms of life, whereas the nitrogen is relatively inert, though it plays an active part in plant life when "fixed", that is, abstracted from the air in the form of one of its compounds. Yet curiously enough, one of the much less abundant constituents of the air plays a comparable role in its importance to life. This inconspicuous member of the atmospheric mixture is water-vapour, which never amounts to more than 4 per cent by volume of the air, and is generally much less. Water-vapour is important not only to life but to the workings of the atmosphere itself, for it has very special properties as an interceptor of radiation.

Table I

Composition of the atmosphere, expressed as percentages of the total mass of dry air

	Gas or Vapour	*Mass % of dry air*
Fixed gases	Nitrogen	75.51*
	Oxygen	23.15*
	Argon, etc.	1.28*
	Other gases	Trace
Variable gases	Carbon dioxide	<1
	Water vapour	<3
	Ozone	<0.1

* from a paper by F. A. Paneth (1937)

Table I gives the composition of the atmosphere by mass. The gases of which the air is made up are grouped into two major classes, the fixed and the variable gases. Those in the first class are fixed in their relative proportions, never varying by an appreciable amount in the atmosphere except at great heights. The second class, however, contains gases or vapours which vary widely and rapidly in quantity in the air. Of these we may dismiss carbon dioxide, which is the end product of many organic processes, including our own breathing. As far as we know, it plays little part in atmospheric processes. Ozone is of greater significance, for it is known to be linked in some way with many of the mechanisms operating in the lower atmosphere. It is, however, concentrated largely above about 70,000 ft. above sea-level, and is hence inaccessible to direct observation, which accounts for our lack of information about it. This leaves water-vapour as the sole important variable constituent. Without the presence of these small quantities of invisible water, life in its present form could not exist on the earth, and the entire circulation of the atmosphere would be radically different.

Among the important consequences of the presence of the vapour, we can list precipitation first: water-vapour is one stage in the so-called hydrologic cycle. The vapour also serves, however, as an important store-house of heat. Great quantities of heat are added to the sea-surface before any water can evaporate; this heat is carried away by the vapour in a latent form. It can be transported great distances, and then released in the form of warmth once again when the vapour condenses. Still more significant is the effect the vapour has on radiant energy. It is able to pluck out some of the incoming solar beams to warm up the air, whereas the other gases are transparent to sunlight. The vapour also intercepts a large part of outgoing terrestrial radiation, thus keeping a considerable volume of heat in the lower atmosphere.

Humidity and Saturation. The term "humidity" refers to the concentration of water-vapour in the atmosphere at a given moment. It has, however, nothing to do with wetness, for the vapour is as dry as any other gas.

The amount of water-vapour that can be present in the air is strictly limited. If we attempt to evaporate water into a confined space, we find that after so much has been evaporated, any further evaporation produces an equal amount of condensation, so that the amount of vapour present in the space thereafter remains constant. This maximum value is referred to as "saturation", and the air or confined space is said to be "saturated" with water-vapour. Experiment further shows that this saturation quantity is the same whether the water is evaporated into an empty space or into air occupying the same volume. Strictly speaking, one should perhaps talk of saturated vapour rather than saturated air, for the presence of air has no effect on the behaviour of the vapour.

This highly significant quantity, the saturation vapour content of the air, is controlled absolutely by one thing alone—the temperature of the air (or of the evaporating surface if liquid water is also present). For most temperatures there is one and one only possible value for saturation vapour content, the only possible exception being the range of temperatures between about 32°F. and —40°F., within which limits it is possible for both liquid water and ice to exist in the air. The saturation vapour content has one value for ice and another for water at the same temperature, so that within this range there are two values for saturation, one appropriate for air resting on or containing ice, and the other for water.

If we plot saturation vapour content against temperature, the curves show clearly that cold air must always be very dry, for even when saturated it contains very little moisture. As the temperature rises, however, the capacity of the air for water-vapour increases rapidly: and furthermore this rate of increase itself grows greater as the temperature rises. For example:

At 30°F., saturation vapour content is 4.4 gm/cubic metre
At 40°F., ,, ,, ,, ,, 6.5 gm/cubic metre
Increase in capacity due to increase of 10°F. is 2.1 gm/cubic [metre
Whereas
At 70°F., saturation vapour content is 18.3 gm/cubic metre
At 80°F., ,, ,, ,, ,, 25.0 gm/cubic metre
Increase in capacity due to increase of 10°F. is 6.7 gm/cubic metre.

In other words, a rise of 10°F. in temperature is more than three times as effective at 70°F. as at 30°F. in increasing the amount of water that can be absorbed by the air.

The practical significance of these results is that the warm air of a summer's day can hold much more moisture than can the cooler air of winter; and, what is more, it usually does. It may seem paradoxical to think of summer as the time of high moisture contents, but such is indeed the case. The most humid day in January in London gives moisture contents of about 8 gm/cubic metre, whereas the warm, dry spells of July will ordinarily come with air containing between 12 and 15 gm/cubic metre, or more than half as much again as in humid winter weather. The reasons for the seemingly greater humidity of winter are several. One of them is the slower rate of evaporation characteristic of cold weather, which leaves the streets wet and uncomfortable. Another is that the winter air is closer to saturation than the summer air, even though it contains less moisture.

The Sources of Atmospheric Moisture. From where does all this moisture come? Although it is only a small part by volume of the air, there is still a large mass of water in even quite moderately humid air. In a room 25 ft. square, for example, with a ceiling height 15 ft. from the ground, there are likely to be some seven to ten pounds of water-vapour on a normal summer's day in England, and in the equatorial zone there would be about twice as much even in the drier spells.

The moisture enters the air by evaporation from water or soil surfaces. If unsaturated air rests on a water surface such as a sea or a lake, the water steadily evaporates, thus raising the humidity until the latter approaches the saturation level. There is also equally rapid evaporation if the air rests on a wet soil surface or on green vegetation: in the latter case, the water is brought up from the soil through the roots and stems of the plants, and is then discharged to the air through the plant leaves. Experiments have shown that such plant-covered soils are excellent evaporating surfaces until the plants wither from lack of soil moisture. A plant-covered area of moist soil is almost as effective in moistening the air as is an open water surface, at any rate in the warmer months.

The rate of evaporation does not, however, depend exclusively upon the initial humidity of the air. Even very dry air will cause very little evaporation unless certain other factors are present. These are briefly summarized below:

(1) *Wind Strength.* If there is no wind, the air rests quietly and calmly upon the water surface. Evaporation then proceeds solely by the slow process of molecular diffusion, in which the molecules of water make their own way upwards through the air above. This process alone is so slow that it is usually ignored. If, however, there is a wind, innumerable eddies are formed, which carry bodily upwards the air which is momentarily in contact with the ground: with the air goes the recently evaporated water, and dry air is brought down to the ground. By this process of mechanical stirring the upward flow of water-vapour is maintained as long as the air near the ground remains unsaturated. This "eddy diffusion" is overwhelmingly greater in its effects than the molecular diffusion discussed above. Wind, then, is an essential for appreciable evaporation. In general the stronger the wind, the more rapid the evaporation.

(2) *Temperature.* A supply of heat is necessary to maintain evaporation, so that the temperature of the evaporating surface and the air in contact with it influences the rate of evaporation not a little.

(3) *Dryness of the Air.* The drier the air, the greater the evaporation is a general rule. It is also important that the concentration of water-vapour should decrease upwards.

From these considerations, it is plain that warm, dry air with a strong wind is the ideal condition for rapid evaporation. On such a day in the English summer, upwards of half an inch of water may be lost by ponds, rivers and lakes, and an almost comparable amount may evaporate from plant-covered soils.

The air of the atmosphere is not normally in the saturated state. As long as the flow is level, and no marked cooling occurs, air currents flowing across long stretches of water tend to have humidities (about five feet from the surface) between eighty and ninety per cent of the saturation value. Over plant-covered land, humidities tend to run a little lower, while over very dry surfaces like the plateaus of inner Asia, values as low as ten per cent of saturation are quite common.

The Measurement of Humidity. The measurement of humidity employs the cooling effect of evaporation from water surfaces just examined. The humidity of the air controls exactly the speed of evaporation from a water surface (other factors being equal) and this evaporation induces cooling in the air immediately in contact with the evaporating surface. The degree of cooling is therefore related in a systematic way to the humidity. To measure the latter, we arrange an instrumental system whereby the cooling effect of the evaporation can be measured. At each air temperature there is a definite limit to the cooling that can be caused in this way, so that our method must seek to secure a perfect evaporating surface which will attain the limit in cooling effect.

The instrument employed is the wet-and-dry-bulb thermometer. It consists of two thermometers erected side by side, the one a perfectly normal thermometer measuring air temperature (i.e. the dry-bulb), and the other a similar thermometer (called the wet-bulb) whose bulb is wrapped in a small muslin bag kept permanently saturated with water. Unless the air is entirely saturated, evaporation then takes place from the muslin round the wet-bulb. The cooling effect lowers the temperature of the wet-bulb well below that of the dry-bulb. This "wet-bulb" temperature is the lowest temperature to which a surface can be cooled by evaporation under the conditions of the moment. Knowing these two temperatures, the dry-bulb and the wet-bulb values, we can readily calculate the humidity of the air. The use of the instrument does not in practice involve any long calculation, for there are tables and slide-rules with which to do the routine calculation. The measurement of humidity at ground-level is then a simple operation.

The units in which we express humidity are a little confusing. The unit ordinarily used, the relative humidity, is an unwieldy quantity in the hands of the inexpert. It has been defined as "the meteorological equivalent of absolute humbug". Some of the commoner units are given below for reference purposes:—

(1) *The absolute humidity* (or vapour density) is the mass of vapour in unit volume of air, the units already employed above. In the accepted metric system of measurement, it is expressed in grams per cubic metre. Though it is the simplest and most direct way of expressing the humidity, other units are used more often.

(2) *The specific humidity and humidity mixing ratio* are numerical quantities expressing the mass of vapour present as a fraction of the mass of air. The specific humidity is the ratio of the mass of vapour to the total mass of the damp air sample containing the moisture, viz.,

$$\frac{\text{Mass of water-vapour}}{\text{Mass of dry air plus water-vapour (i.e. the damp air)}}$$

whereas the humidity mixing ratio is the ratio of the mass of vapour to the mass of the dry air remaining in the sample after the vapour has been removed, viz.

$$\frac{\text{Mass of water-vapour}}{\text{Mass of dry air}}$$

Numerically these quantities are almost identical: both are expressed in grams per kilogram.

(3) *The vapour pressure* is a very different quantity. All gases and vapours exert a pressure on objects with which they are in contact. Atmospheric pressure is the sum of the pressures exerted by the gases making up the air. The pressure exerted by the water-vapour is only a small part of the whole, but it may amount to 3-4 per cent. Vapour pressure is expressed in the absolute unit of pressure, the millibar. In England, average vapour pressures vary between about 7 mb. and 15 mb. In the equatorial belt they habitually lie about 30 mb.

(4) *The relative humidity* is the ratio of vapour pressure to saturation vapour pressure, i.e. of the actual pressure at the moment to the pressure that would be exerted if the humidity rose to the saturation point. It is expressed as a percentage; a relative humidity of 50 per cent means that the vapour pressure is half the possible maximum. More generally the relative humidity

represents the degree of saturation of the air. The objections to its use arise from the fact that the denominator—saturation vapour pressure—varies with the temperature. The meaning of the relative humidity therefore changes from day to day. A relative humidity of 80 per cent on a midwinter day means a moisture content of perhaps 5 gm/cubic metre; the same value in July means probably 10-12 gm/cubic metre. In the hands of the inexpert such a ratio can lead to endless nonsense and confusion. The writer can remember one day in the summer of 1946 in his home in Montreal when the temperature rose to 90° and the humid heat was stifling. The newspapers, however, assured him that the humidity was "only" 60 per cent, so that the air was "quite dry". They had overlooked the fact that 60 per cent of a very large quantity—and saturation at 90° is very high indeed—can itself be a large quantity.

The above units can all be used effectively in climatology, but the fourth is subject to the fatal defect of ambiguity. What is more, averages of relative humidity mean very little. It is probable that the best all-round unit for climatological purposes is the *dewpoint temperature*, a unit not so far discussed. It was seen above that saturation humidity depends only on the temperature. It follows logically that for every humidity there is one significant temperature, the temperature at which that humidity is the saturation value. If air is cooled, there comes a point at which any further cooling must lead to condensation. The cooling of the ground by nocturnal radiation, for example, very often leads to dew deposition on the vegetation. The critical temperature at which the dew begins to form is called the dewpoint of the air. The dewpoint may thus be defined as the temperature at which the air would become saturated when cooled.

Condensation

When saturated air is cooled, some of the vapour is converted into liquid water, either by being deposited on the walls of the cooling chamber, or also as tiny droplets which float within the air. If the air is extremely pure and free from dust, it is possible to cool it well below its dewpoint before any condensation occurs, but the atmosphere is never in this laboratory-achieved condition. In the air condensation occurs either as dew or hoar-frost on the ground, vegetation, and buildings, or as fog or clouds, consisting of enormous numbers of suspended water droplets or ice crystals. The formation of ice-crystals goes by the special name of "sublimation".

Experiments have clearly shown that each droplet of water or crystal of snow forms round a tiny particle of some non-aqueous substance These "condensation nuclei", as they are called, occur in great numbers in the air. They vary widely in composition, though it is likely that only certain kinds of nuclei will serve for condensation. Droplets of liquid water appear to condense on nuclei consisting of hygroscopic substances, such as common salt (sodium chloride) and sulphuric acid, both of which occur in the atmosphere in considerable quantities. The salt comes from the evaporation of spray over the sea, and is very widespread. The sulphuric acid, however, is more restricted in occurrence; it appears to form near cities, where large volumes of smoke are poured out into the air. The smoke contains traces of sulphur dioxide, which is oxidized to sulphur trioxide when the sun shines through it. The sulphur trioxide combines with water to form the acid. The enormous numbers of dust particles are believed to be of secondary importance as condensation nuclei, though some are "wettable".

Whatever the type of nucleus employed, however, there appears to be no case where condensation is hindered by the lack of them, except perhaps in the case of sublimation, which may sometimes be restricted. The fact that some of the nuclei are hygroscopic (i.e. have an affinity for water) may even mean that condensation may occur at humidities below the theoretical saturation value, possibly at relative humidities as low as 70 per cent. Lowell (1945, p. 255) has claimed that this pre-saturation condensation is responsible for the hazy, amorphous appearance of the sky before rain.

The Formation of Raindrops and Snowflakes

It is now customary to regard the formation of new cloud particles (i.e. condensation) as a process distinct from the fusion of those particles to form rain-drops or snow-flakes. It might be supposed that rain-drops are formed simply by continued condensation on the tiny cloud-droplets. There is now, however, general agreement that rain-drops develop by the fusion of cloud-droplets rather than by the direct growth principle; the process whereby cloud droplets or particles fuse together in this way is called "coagulation", a word used rather loosely.

Great interest has centred in recent years round the theory of T. Bergeron (1933) who suggested that the chief coagulant process in the atmosphere depended on the co-existence in the same cloud of both water and ice-crystals. Bergeron argued from the fact that the saturation vapour pressure of water is greater over water than over ice. If particles of ice came into intimate association with water-droplets, it might therefore be anticipated that evaporation would begin from the water-droplets and condensation take place on the ice-crystals. Appreciable rain, he argued, always fell from clouds which extended above the height of the freezing level, where such an association of ice and water is possible. It must be remembered that super-cooled water droplets predominate in clouds whose temperature is between +32°F. and —20°F.; only below —40°F. are clouds entirely of ice-crystals. The kind of clouds that met Bergeron's requirements were alto-stratus, where snow very often fell into the layers of super-cooled water-droplets just above the freezing level, and cumulo-nimbus, in which strong vertical currents carry both cloud-droplets and rain-drops far above the freezing level.

It is naturally difficult to put this theory to the test, but there is a great deal of indirect evidence in favour of it. It is a fact of observation that thick clouds are necessary for appreciable rain; thin sheets of stratus or strato-cumulus never give more than slight rain or drizzle, whereas alto-stratus and the thicker nimbo-stratus are the types from which nearly all steady rain falls. The heaviest rain of all falls from the cloud-type with the greatest degree of vertical development, cumulo-nimbus. Aerological observations confirm that in the very great majority of cases of appreciable rain over Britain and North America, the cloud masses extend several thousand feet above the freezing level. There have, however, been numerous observations from intertropical areas of moderate rain from cloud, which, though thick, failed to reach the high freezing levels typical of such latitudes. Plainly other mechanisms cannot be excluded from the solution.

Added confirmation for Bergeron's view has been forthcoming in the post-war period. In the weather-radar experiments carried out by the Defence Research Board of Canada, W. Palmer was able to measure the amount of free rain and snow within a warm front cloud system. His results show that there was only light snow well above the freezing level in the case examined, but there was an enormous increase in the volume of precipitation in the two thousand feet above the freezing level.

Perhaps the most dramatic confirmation of deductive theory ever achieved in physical meteorology, however, has been the outbreak of "rain-making" in the United States. This novel process, which threatens permanently to exclude astrologers and witch doctors from their profession, was conceived by engineers of the General Electric Company at Schenectady, New York. They discovered that by dumping a small quantity of finely divided dry ice (solid carbon-dioxide) into a cumulus cloud, snow could be made to fall from the base of the cloud. During the summer of 1947 many successful snow-making attempts took place in the United States, chiefly in the dry western states. Another interesting and well-documented case comes from Australia, where E. Kraus and R. Squires (1947) were able to convert an innocuous cumulus cloud into a large cumulo-nimbus mass whose anvil eventually reached 45,000 ft. Other experiments have used finely divided silver iodide (released from ground generators) as the seeding substance.

Whether such experiments may have the economic significance claimed for them by some of their sponsors, we cannot yet say. They have made clear, however, that the presence of ice-crystals in clouds is the dominant factor in the production of rain, Whatever the ultimate solution to the problem of coagulation may be, it is unlikely that it can be far away from Bergeron's original deductive theory.

Snow-flake formation presents problems quite unlike those of rain, and it cannot yet be stated that we know how ice-particles grow to the snow stage. The large snow-flakes typical of snow falling from air near the freezing point are aggregates of platy, stud-shaped crystals or crystal-aggregates, and it is the growth of the ice-particle to these constituent crystals which presents most problems. Snow at low temperatures occurs with a wide variety of forms; the commonest type in the great blizzards which affect North America in mid-winter is a small pellet of sub-crystalline ice. The beautiful flower-like crystals usually illustrated in descriptions of snow tend to occur when the circulation is light.

It may be noted that the rain-making process described above is more properly regarded as a snow-making process. Where dry-ice is the "seeding" substance, the effect is merely that each intensely cold dry-ice pellet produces millions of ice crystals by cooling the air below the critical temperature of —40°F. Silver iodide, on the other hand, acts more as a nucleus for sublimation, having a crystalline form very similar to that of ice. In both cases, however, the effect is the same: once created, the ice-crystals grow by something akin to Bergeron's process, and eventually fall as snow-flakes or rain.

Cloud Formation

We have now examined the physical processes of condensation, by which vapour passes to the cloud-droplet or ice-particle state, and coagulation, by which these tiny and almost buoyant cloud droplets are fused together to form rain. It remains to discuss the mode of formation of clouds. All clouds are plainly the result of large-scale condensation taking place through a considerable volume of air, and in almost every instance such condensation results from prolonged cooling of the air concerned. The particular type of cloud which is formed depends on the manner in which this cooling is brought about.

The chief mechanisms leading to atmospheric cooling are listed below:

(i) Radiative cooling, which occurs chiefly at the earth's surface or at the upper surface of clouds;

(ii) Convective cooling, the process whereby vertically rising, buoyant parcels of air cool adiabatically, to be discussed in Chapter III;

(iii) Cooling by mixing of unlike airmasses. A mixture of two samples of damp air of widely different temperatures may induce saturation or super-saturation in the mixture;

(iv) Advective cooling, where a warm airmass passes over a cold land or sea surface, being thus cooled by direct contact;

(v) Cooling by upslope motion; when an airstream encounters a hill barrier or a wedge-shaped mass of colder air (as at a frontal surface) it rises along an obliquely sloping path. This type of uphill motion is distinct from the truly vertical motion associated with convection (No. ii above), though they may occur simultaneously.

It will be necessary to discuss each of these processes separately in order to cover the whole range of cloud-types. There are many excellent accounts of cloud-formations which discuss the appearance and anatomy of clouds admirably, usually with the aid of photographs. The objective here is to discuss processes rather than forms.

Cloud and Fog from Radiative Cooling

In the mind of the uninformed foreigner, the most characteristically British weather type is a thick blanket of radiation fog. Though this judgement is as prejudiced as the British estimate of the worth of the foreigner concerned, there is no gainsaying that such fog is distressingly common in some localities, especially in winter. When fog is mixed with smoke—smog, as the Americans call it—it constitutes a major discomfort to all who experience it, and smog is unfortunately a common weather type in London and the smokier cities of the Midlands or the North.

Radiative cooling within the atmosphere takes place chiefly from the top of clouds and from the ground surface itself. A sheet of cloud radiates more or less as efficiently as the ground surface, and one might suppose that a cloud-layer by night would thereby tend to thicken upwards. The base of the cloud, however, is heated by absorption of long-wave radiation from below, so that the base tends to disperse as the air is warmed above the dewpoint. The effect of nocturnal radiation upon most layer-clouds is hence to cause them slowly to rise, and in many cases to break up or disperse. At ground level, however, radiative cooling can in some circumstances cool the air low enough to produce extensive sheets of fog, which only the sun can disperse in the absence of wind.

It is possible to list the factors which favour fog formation fairly briefly:

(i) Clear skies at night. A cloud cover "blankets" outgoing radiation by sending back to earth almost as much energy as is radiated off. Low cloud may reduce the loss of heat to as little as 15 per cent of the average value with clear skies. Higher clouds are less effective.

(ii) Light winds. Strong winds stir the air so greatly that no one element of air remains in contact with the ground long enough to be much cooled; moreover, there is a flux of heat downwards to the ground, which hence cools less rapidly. On the other hand, a total absence of wind means that only a very shallow layer of air is cooled by contact. This gives a very steep inversion near the ground, below which there may be ground fog. Such fog may be quite dense, though less than six feet deep. The ideal condition for fog is a light wind, say of five miles an hour; there is then enough stirring to spread the cooling process through a thick layer, but not so much as to disperse the fog entirely. The thick fogs of England may be 500 ft. or more in depth.

(iii) Humid air. The more nearly saturated is the air, the more readily does fog form. On the other hand, radiative cooling proceeds more readily in dry air so that once again intermediate conditions are most favourable. Ideally the air should be moist in the bottom two or three thousand feet and dry aloft. These conditions are normally found in Britain within warm anticyclones, which generally have maritime polar air at the surface, overlain by warm, dry, subsiding air.

Such conditions are often realized over Britain, especially in inland districts. The high fog incidence of the autumn and early winter is due to the lag in the fall of sea-temperature behind that of the land. Maritime airmasses entering Britain at this time of year are still relatively warm and humid; they are hence ripe for fog formation when exposed to nocturnal cooling. In the late winter and early spring sea-temperatures are lower, and the initial moisture content of maritime air is correspondingly reduced. At all inland stations there is a marked diminution in fog frequency after mid-February.

The break-up of radiation fog after sunrise is a familiar phenomenon. In the majority of cases thin ground fog sheets are rapidly dispersed by solar heating. The increase in wind which usually accompanies the rise of the sun assists in the dispersal; quite often the fog is lifted off the ground to form a low sheet of stratus cloud, which in turn is broken and later dispersed by the sun. In mid-winter, however, the considerable depth of the fog sheets and the weakness of the sun often prevent dispersal, and the fog continues through the day, sometimes for several days on end. In such conditions the smoke and soot of cities accumulate within the fog itself, leading to the "smogs" referred to above.

Though fogs may occur in many different synoptic situations, it is in the interior regions of warm highs, with the sluggish circulation and moist air typical of such systems, that they are most widespread. The pernicious qualities of winter anticyclonic weather in Britain are twofold; either there is a pall of grey strato-cumulus cloud with hazy, very dull weather beneath, or else the sky is clear and widespread radiation fog develops. The worst of all winter weather comes when a thin sheet of strato-cumulus drifts over a fog-covered surface. The fog is then cut off from the sun, which might have dispersed it, and will persist until the anticyclone or the cloud gives way.

Convection Cloud. The second main cooling process listed on p. 29 is convection, which consists in vertical currents within the atmosphere normally set off by surface heating or turbulence. Convection gives rise to the great suite of clouds called "cumulus", the term applied by Luke Howard to the heaped, cauliflower-like masses of the summer sky. The development of such clouds depends on stability, and the cumuliform suite is hence discussed in the chapter on that broader topic (Chapter III).

Clouds Formed by Mixing. It has been realized for a long time that a mixture of two damp (not necessarily saturated) air-masses of widely different temperatures may itself be saturated or even super-saturated. There are several ways in which mixing of unlike airmasses can take place within the atmosphere. At a frontal surface, for example, conditions are ideal for this process. It is often suggested that the strato-cumulus and alto-cumulus clouds associated with fronts are formed by mixing of the two airmasses concerned. Conditions at frontal surfaces are, however, unfavourable for deep mixing, and it is likely that such cases of cloud are confined to thin sheets. Since the mixing involved is the result of turbulence, clouds formed in this way show signs of turbulent structure, which often takes the form of regular patterns within the cloud, like the mackerel sky of alto-cumulus.

Many of the ubiquitous strato-cumulus sheets of Britain's skies probably develop by this process of airmass mixing. The ever-present stratus and strato-cumulus of the north-east monsoon of South China and Indo-China have also been ascribed to this process.

Clouds and Fog Resulting from Advective Cooling. The low stratus cloud and sea-fog so typical of our south-west coasts in early summer are representatives of a widespread type. Wherever warm moist air streams across a cold surface, there is a tendency for condensation and cloud or fog formation. The cooling induces a temperature inversion, below which the air becomes thickly charged with clouds or fog. With fairly light wind the condensation takes the form of thick, drifting fog, but with stronger winds the fog may lift to stratus cloud. A cold land surface may produce either effect, but advective cloud and fog are both seen best at sea. The cold sea areas are among the gloomiest parts of the world in consequence.

In Britain, marine stratus or sea-fog may occur at any season when unusually warm air crosses the surrounding seas. In the Channel and Irish Sea areas such cloud and fog develops chiefly in maritime tropical or warm maritime polar air moving north-eastwards across them. Stratus is common at all seasons, but sea-fog is especially common in spring and early summer, when the sea is still cool relative to the warmer airmasses. The North Sea also suffers greatly from such cloud and fog, chiefly when warm air from the continent travels westward towards Britain. The "haars" of Eastern Scotland and the less formidable equivalents along England's east coast originate in this way. Whenever warm continental air approaches our east coasts in spring and summer, sailors, airmen and weather forecasters wait expectantly for what rarely fails to come, the clinging wet veil of fog, or at the best low stratus, which blots out sea and land alike.

Elsewhere we may note the great fog belts of the Grand Banks off Newfoundland, where fog is almost continuous in the warmer months. Here the source of the cooling is the cool water of the Labrador Current, and the air is maritime polar or maritime tropical from the warmer seas to the south. There is a similar area in the Far East, around Hokkaido and peninsular Kamchatka. The "cold-water coasts" in the middle latitude belt afford another example; the fogs of the coastlands of California, Morocco, Chile and elsewhere develop when moist air flows landwards across upwelling waters close inshore.

Clouds Associated with Upslope Motion. By this title we comprehend clouds which develop in broad currents of air which are being forced to ascend some sloping surface, either a hill barrier or a frontal surface. Such clouds differ radically in form from convective clouds, where the dominant motion is vertical. With frontal cloud we shall deal at length in Chapter V. Here we must devote attention to orographic cloud and rain formation.

The reluctance of moving airstreams to pass over the summits of hill-barriers in their path is a conspicuous phenomenon. The violent winds like the Mistral and Bora represent canalized flow through gaps in the hill-barrier presented by the Pyrenees, Alps and Dinaric Mountains; to avoid passing over these hills, several thousand feet of approaching air-columns are canalized through the few available gaps. The most conspicuous instances of orographic cloud formation hence tend to occur in one of two situations: over hills like those of Highland Britain, which are too low to dam up the circulation, or in areas where diversion of the flow is impossible, there being no gaps through the barrier concerned. Examples of this second group are the Pacific Ranges of North America and the "Hump" between Assam, Burma and Yunnan. All hills, however small, have an appreciable effect on circulation, especially if they are isolated, but it is impossible to say from height alone how effective a range of hills will be in leading to orographic cloud and precipitation.

There are three main cases which we can examine: the cases of stable, potentially unstable and unstable air respectively; the main features of each are illustrated in fig. 1. Where unstable air is concerned, all that happens is a great intensification in the scale of convection cloud and showers on the windward side and summit of the range, and a rapid clearing of the skies in the lee. All of us are familiar with the showery day in the hills, with heavy rain showers every hour or so, sometimes with thunder, on occasions when over the low ground there was only scattered cumulus. In the Lake District and the western Pennines such weather nearly always occurs in the cooler types of maritime polar air, whereas such air brings to the eastern slopes of the Pennines delightfully clear fine weather. Much more interest, however, attaches to cases A and B in fig. 1, where the air is stable (A) or potentially unstable (B). In such instances the hills are usually blanketed with dense multilayer cloud sheets, often closely resembling frontal cloud. With stable air there may be drizzling rain in the higher parts of the windward slope, though heavy falls are uncommon unless convergence is present. mT air (see Chapter V) gives such weather in most hilly country round Britain. Where the air is potentially unstable, however, the layer clouds are often mixed with convective cumulo-nimbus clouds, due to the release of the instability during uplift.

The effect of relief on cloud and rainfall distribution is extremely important. A rainfall map of Britain usually resembles a contoured relief map, all the areas of heavier rain being hilly country. The same is true of the Pacific coastlands of North America, where rainfall is excessively heavy on the western flanks of the main ranges. As far as Britain is concerned, the position can be generalized as follows:

(a) The proneness of the whole country to frontal passages, instability rainfall and other sources of rainfall independent of relief varies little from place to place. In

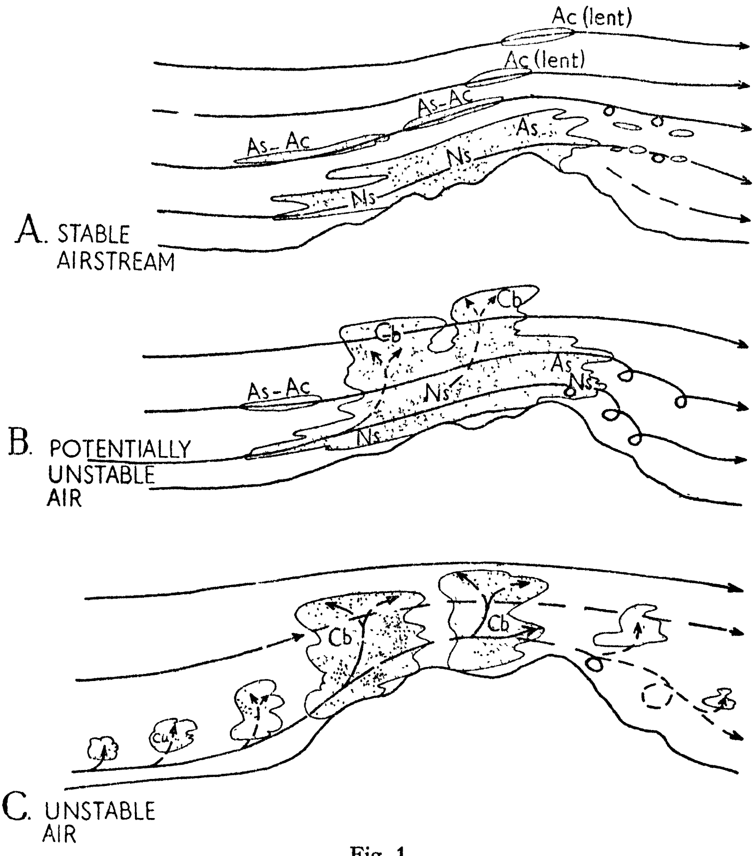


Fig. 1

The formation of cloud during the passage of moist air over mountains. Clouds are indicated by usual abbreviations.

winter, the north and west tend to be closer to the centre of rain-producing disturbances, and so to be rather wetter than the south-east. In summer the reverse may even be true.

(b) The actual rainfall map shows that variations from point to point depend almost wholly on relief, both local and regional. In southern England, for example, the North and South Downs and even 400-ft. high Portsdown are conspicuously wetter than are the broad clay vales, despite the constancy of what might be called "rainfall opportunity", i.e. frequency and intensity of rain-forming disturbances.

So great is the dependence of rainfall upon relief—hills being apparently the most effective way of making air shed its moisture—that geographers are in danger of forgetting that the dynamical control of climate is as significant in hill country as over the plains. Hills have the effect of intensifying rain areas, or, as in case B, fig. 1, of releasing energy already inherent in the air. In hill country as over the plains, the chief source of rainfall outside the Tropics is the frontal or cyclonic passage, to be discussed in Chapter V. Hence we find that most instances of heavy orographic rain occur when lighter rain is also falling in the surrounding plains. The hills cannot pluck rain out of a dry, stable atmosphere.

Rain-shadow is a phenomenon too well known to need detailed treatment in this book. The descending air on the lee side of hill ranges undergoes dynamical warming which disperses all or a large part of the dense cloud found on the windward slopes. The remarkable effects of rain-shadow have been described in many works. Perhaps one that has rarely received attention in Britain is the rain-shadow of the Highlands over the inner shores of Moray Firth between Invergordon, Inverness and Lossiemouth. This tiny area of farmland rarely or never experiences the densely cloudy, wet weather so common on the west coast of Scotland. Even Atlantic depressions rarely give cloud below about 2,000 ft. (down to 300 ft. is common on the west coast), and the total annual rainfall is below 30 in. over a wide area (cf. more than 100 in. thirty miles to the west).

The climate of Inverness is one of the brightest and most cheerful in Britain, thanks to its encircling girdle of mountains. Airfields around the shores of Firth are hardly ever hindered by bad flying weather.

CHAPTER III

STABILITY, CONVECTION AND TURBULENCE

WITHIN the troposphere the predominating motion of the atmosphere is horizontal: vertical currents are both less extensive and less vigorous than the "winds", which consist of air in horizontal motion. These vertical currents, however, are of vital importance, since without them there could be no appreciable precipitation anywhere on earth. It will be seen later that vertical motion in the atmosphere is of two kinds:

(1) general subsidence or uplift of air over wide areas under purely dynamical controls; this type is very slow, though of great moment: and
(2) local up-and-down currents of an altogether smaller type. We call such currents "convection currents"; they may achieve considerable speed. By means of them the troposphere is as it were *stirred* from time to time.

The effects and causes of type (1) vertical motion will be dealt with in Chapter V. Here we are concerned with convection, which is responsible among other things for the formation of cumulus cloud, showers and thunderstorms. The degree of proneness of the atmosphere to convective disturbances is what is called "stability"; the air is stable if convection is hindered or prevented, and unstable if there is larger scale convection. The stability is largely determined by the lapse rate of temperature.

Cooling During Uplift: Adiabatic Processes. When a body of air rises, it expands; that is to say, it does "work" upon the surrounding atmosphere, and acquires more space for its molecules to occupy. Now rising air is largely insulated from the surrounding air: heat can neither penetrate into nor escape from the rising "bubble". To get the energy necessary to carry out its expansion, the air has to draw upon its own supply of heat, part of which is converted into kinetic and potential energy. The air is therefore cooled internally. This form of cooling, brought about by expansion, is an *adiabatic* process; it takes place through internal changes in the rising air and not through the abstraction of the heat by the surrounding atmosphere. A similar process in reverse affects air that sinks through the atmosphere. The latter "does work" on the sinking air, forcing it to contract in volume and to gain in heat content. As the body of air sinks it warms up at the same rate as it cooled at when rising. These processes are what we mean when we talk loosely about rising air cooling and sinking air warming up.

Adiabatic cooling and warming of air take place at fixed rates. If unsaturated air rises, it cools everywhere at the fixed rate of about 5.4°F. per thousand ft., which is known as the *dry adiabatic lapse rate.* If, however, the air is saturated, condensation must take place when further cooling occurs. This condensation releases a considerable quantity of heat hitherto latent in the water vapour, which reduces the rate of adiabatic cooling. Saturated air hence cools at the much smaller rate known as the *saturated adiabatic lapse rate*, which varies from about 2.7°F. to 3.4°F. at ground level.

It is important to distinguish clearly between these adiabatic lapse rates and the environmental lapse rate defined above. The adiabatic lapse rates are the *rates of cooling* affecting rising air. The environmental lapse rate has nothing to do with either cooling or rising air. It is merely a statement of the temperature distribution in the atmosphere at a given moment. To measure the environmental lapse rate, it is only necessary to arrange for instrumental readings at fixed levels, whereas the adiabatic lapse rates could only be measured directly by suspending a thermometer within the rising air currents and allowing it to rise with the air, which is plainly impossible.

Absolute and Conditional Instability

A layer of air is defined as absolutely unstable if the environmental lapse rate exceeds the dry adiabatic lapse rate, i.e. exceeds 5.4°F./1,000 ft. For if a small body of air is displaced upwards or downwards, it becomes lighter or heavier than the air around it, and will continue to move up or down spontaneously. Absolute instability of this sort occurs only near the ground on hot days. It causes violent convection currents, but is not common enough to be of major importance.

Conditional instability develops when the environmental lapse rate lies between the dry and saturated adiabatic values. Such air is stable by the criterion discussed in the foregoing paragraph; air which is displaced from its original level tends to return to that level. If the air is saturated, however, its rate of cooling during uplift (the saturated adiabatic rate) is less than the environmental lapse rate: hence displaced air tends to move spontaneously away from its original level if displaced appreciably. In all such cases instability and active convection is plainly *conditional* upon the presence of abundant water vapour: hence the term "conditional instability".

The air in the lower troposphere is quite often conditionally unstable, especially in the cool, moist currents of oceanic origin that we call maritime polar air. Many showers, thunderstorms or abundant cumulus cloud characterize such air when it is disturbed.

Absolute Stability

If the environmental lapse rate is smaller than the saturated adiabatic lapse rate (i.e. less than about 3.0°F. per 1,000 ft.) the air is plainly stable whether saturated or not. In such an atmosphere all convection currents are damped out, and the air flows smoothly and horizontally, undisturbed by any vertical currents. Cumuliform cloud and showers are hence ruled out, though there may be other forms of cloud. It is clear that inversions are layers of great stability, so that they effectively prevent any movement of air upwards or downwards through them. The smoke and fog which accumulates in cities beneath inversions on clear nights would be dispersed by the turbulent eddies of the wind but for this protecting factor. It is clear also that the stratosphere, with its zero lapse rates, must be a very stable region. Rising convection currents quite often reach the tropopause, but they can never penetrate the stratosphere. Even the huge convection currents set up by the blast of the atomic bomb at Bikini atoll were halted shortly after they penetrated through the tropopause.

Potential Instability

So far we have considered instability in terms of what happens to small bodies of air displaced by some agency from their original level. Potential instability is a condition of a different order. Formerly called convective instability, it was defined in its present form by E. W. Hewson (1936–38) in a series of papers on the use of wet-bulb temperature in weather forecasting.

It frequently happens that large masses of air are slowly forced to rise from one general level in the atmosphere to another. Air which moves towards higher ground, for example, undergoes bodily lifting. Another common case is where an airmass moves upwards over a wedge-shaped mass of cooler air, i.e. at the front. This bodily uplifting may greatly change the stability of the air. If, after lifting, the air has become conditionally unstable, its original condition is referred to as potential instability, i.e. instability which was realized only when the air was bodily lifted.

There is no space here for a theoretical discussion of such changes in stability. It can be stated, however, that potential instability is caused by certain types of distribution of water-vapour in the atmosphere. Where the surface air is moist and the upper troposphere decidedly dry, it is very probable that potential instability will be present.

The types of air which are most prone to potential instability are the hot, moist airmasses of tropical origin such as occasionally reach Britain from the south and south-west. Over Britain the condition is not common, but on the western sides of the great oceans, and especially in eastern North America and the Far East, most of the warmer, moister types of air are potentially unstable to some extent. Such air gives long periods of warm, close but settled weather as long as it flows along a level path, but if it is forced to rise, as over the Appalachians or along a front, heavy rains of a thundery character at once break out. The airmasses of the Mediterranean Sea are very often similar in type. Potential instability is hence an exceedingly important property, even though it is very hard to detect before it has been converted to true conditional instability.

The Trigger Action in Convection

In our discussion of absolute and conditional instability, we assumed a small displacement of the body of air whose history we were examining, and showed that under certain circumstances this small displacement might grow into the great convection currents which cause the cumulus clouds, the showers and the thunderstorms of the atmosphere. This prompts the question: what causes the initial small displacement? The instability itself is not enough. Some trigger action is needed to start the convection off even in the most unstable air. It remains, then, to discuss the various types of trigger action found in the atmosphere. The chief of these are listed here:

(1) uplift over hills and mountain ranges;
(2) heating from below over land or sea;
(3) mechanical and internal turbulence;
(4) convergence in the airflow;
(5) uplift at fronts.

Of these five, the first is most readily understood. Obviously if an unstable airmass passes over a range of hills, the initial displacement will be provided, and convection will ensue. This accounts for the heavy and frequent showers experienced, for example, in the hills of Highland Britain when unstable, moist air comes in from the Atlantic. Quite small and isolated hills are enough to start convection in many cases.

The second type of trigger action is very common, being due to the heating of land surfaces by the sun. After dawn the ground and the air above it heat up rapidly, temperatures reaching a maximum in the early afternoon. The effect of the sun on the air itself, however, is small, so that the air some distance above the ground remains relatively cool. The overall result is greatly to increase the environmental lapse rate to a point at which it may even become absolutely unstable near the ground. Furthermore, the heating of the ground surface is irregular: some parts get much hotter than others. The air resting on these hotter areas becomes buoyant, being lighter than the cooler air around it. It rises, and convection has started: the instability does the rest. At night, when the ground cools down, the convection dies away. It is for this reason that cumulus clouds, showers and thunderstorms are commonest over flat land during the midday and afternoon hours. We are all familiar with the sort of day that dawns cloudless, becomes cloudy around noon, gives us a shower in the afternoon, and passes to another cloudless night.

Heating of a different nature may occur at sea, where cool air may pass over warm water surfaces. Here again the heating steepens the environmental lapse rate until the air nearest the sea becomes very unstable indeed. It is impossible to assume here, however, any differential heating, for the sea temperature is usually uniform over a wide area. We, therefore, have to look elsewhere for a trigger action. Probably this action is provided by turbulence listed above as the third source of trigger action. We shall see in the next chapter that the air does not flow smoothly and evenly over the surface of the earth, but is deformed by a series of eddies and disturbances we collectively call "turbulence". Turbulence arises partly from friction between the moving air and the earth, but also partly from internal forces within the air itself. There are always such eddies in moving air, and in unstable air they are usually very well developed. It is probably the result of such turbulence that unstable air over a uniformly warm sea surface breaks up into convection currents. Even over land it is probable that mechanical turbulence is the dominant factor in starting deep convection in unstable air.

The fourth and fifth factors will be discussed in later chapters.

CHAPTER IV

THE WEIGHT AND MOVEMENT OF THE ATMOSPHERE

THE barometer is one of the oldest and most important of the instruments by which the changing state of the atmosphere is observed. It is a far cry from Torricelli's tube of quicksilver, which he first used in 1643, to the Kew barometer of today, with its vernier scale, built-in thermometer and self-regulating cistern. It is an even further cry from the seventeenth century's beliefs as regards what the barometer actually measured to the elaborate dynamical analyses which the meteorologist of today bases on world-wide pressure readings. Yet though our ideas have changed rapidly in the past century, the barometer has remained the most treasured instrument of the amateur just as much as of the professional meteorologist. Apart from the clock, there is probably no instrument so widely employed by the lay public; and there is certainly none that can exceed its value to the serious student of the atmosphere.

The value of the barometer in meteorological analysis was discovered in the middle of the nineteenth century in England, when with the aid of the newly installed telegraph system, simultaneous readings of the instrument could be collected from a number of stations over the country. It was soon found that the winds were in harmony with the pressure readings, and that they blew roughly along the lines of equal pressure or "isobars". It had been known before, of course, that fluid motion was controlled by its internal pressure distribution, but here was proof that the rule was true of the atmosphere also. Today our knowledge of atmospheric circulation is derived to a large extent from systematic pressure readings over the globe, and only to a much smaller extent from direct measurement of the wind. To the layman rises and falls of the barometer mean good or bad weather. To the professional they mean much more; they are the keys to the movement of the winds, and are hence the starting point in analysing the atmospheric circulation.

It is not too much to say that systematic study of the weather depends primarily on the relationship which exists between wind and pressure, and on the ease with which pressure can be measured.

Atmospheric Pressure

The air is a fluid, and though it is invisible to us, and its pressure is imperceptible as long as it does not change too rapidly, it actually exerts the formidable sea-level pressure of some 15 lb. per square in. on exposed surfaces. The pressure exerted by a fluid at rest is exerted equally in all directions. In whatever position we expose a surface within a fluid, the pressure on it is always the same at the same point. Bodies which are immersed in a fluid are hence unaffected by the force except that it tends to compress them. We ourselves are unable to feel the pressure, since our systems are pervaded by fluids at pressures similar to that of the atmosphere. Such omnidirectional pressure is called *hydrostatic pressure*. The pressure distribution in a moving fluid is not hydrostatic, and objects exposed to it have a higher pressure on their upstream face and a lower pressure on their downstream face.

The atmospheric pressure is almost an exact measure of the weight of the atmosphere above the point of observation. Pressure is defined as the force exerted on unit area; and if we take a column of air of unit cross section, its change with height is given by the equation

$$dp = -g\rho . dz$$

where p is the pressure

g the constant acceleration of gravity (981 cm. per sec^2)

h the height of the air column, and

ρ the density (i.e. mass per unit volume) of the air.

Since, however, the density of the air decreases with height in an irregular and unpredictable manner, this equation means very little. It can, however, be readily converted into what is known as the altimetric formula, which in its simplest form is

$$\log p_o = h - \log p \ (123F - 56{,}400)$$

where p_o = pressure at base of column in millibars (see below)

p = pressure at top of column, at height h (in ft.)

and F = average temperature of the column in degrees F.

logarithms being taken to the base 10.

We see, therefore, that pressure decreases with height, for as one ascends, the weight of air overhead decreases. At 10,000 ft. pressure is barely 70 per cent of what it is at ground level, and at 30,000 ft. it is barely 30 per cent of the surface pressure. Thus, at the level of Mount Everest, some 70 per cent of the mass of the atmosphere lies below the observer; plainly the air must be concentrated at low levels, for the summit of Everest, high as it is, is only at low levels by comparison with the depth of the atmosphere itself.

The fundamental unit of pressure employed by the meteorologist is the millibar, which is the form in which it should be expressed in the altimetric formula given above. Pressures are normally measured by means of the mercury barometer, which balances the weight of a column of mercury against the weight of the atmosphere. The older unit, still sometimes used, was simply the length of the mercury column required to achieve this balance. In the metric system the length was expressed in millimetres, and in British units the inch was standard. One therefore had the seemingly absurd situation that a quantity whose dimensions were force per unit area was expressed in units of length. The millibar is a so-called "absolute" measure of pressure, having the same dimensions as the quantity it is used to express. It is derived from the physical unit of force, the dyne, which is the force required to accelerate one gram of mass one centimetre per second. A "bar" is defined as one million such dynes per square centimetre. The bar is very nearly equal to the normal pressure of the atmosphere, so that it would be an inconvenient unit in practice. Instead, we use the millibar, which is one thousandth part of a bar, and is equal to one thousand dynes per square centimeter.

$$1 \text{ millibar} = \frac{1}{1{,}000\text{th}} \text{ bar} = 1{,}000 \text{ dynes/sq. cm.}$$

The millibar is related to the older pressure units in the following manner:

1000 mb. = 750.1 mm of mercury = 29.53 in. of mercury. Average atmospheric pressure at sea-level is about 1013.2 mb.

Since pressure falls off rapidly with height, we need to define a standard height of observation if we wish to compare pressure at one point with that at another. If we do not take this precaution, differences in simultaneous pressure readings at two points arise very largely from differences in the level of observation. A height difference of 1,000 ft. is equivalent to about 33 mb. of pressure, which is roughly the difference in pressure between the central regions of anticyclones and depressions. The standard height employed is the sea-level surface. If a reading of pressure is made 500 ft. above sea-level, an extra amount equal to the pressure exerted by 500 ft. of air at the ground-level pressure and temperature is added to the barometer reading. In this way, pressure readings from widely separated points are rendered comparable: differences between them reflect genuine differences between the air columns concerned, and not differences of level. The sea-level pressure readings have a definite dynamical significance and application.

In certain high-level areas reduction of pressure readings to sea-level is impracticable. Stations at levels more than about 3,000 ft. above sea-level cannot adequately reduce their pressure readings, which would in any case mean very little when converted. On high-level plateaus it may be practicable to refer all pressure readings to a standard surface closer to the average level of the ground. In the western United States, for example, extensive use is made of charts of pressure reduced to the 5,000 ft. level. In general, it can be said that to draw sea-level isobars in areas where the general ground-level is above 3,000 ft. is undesirable and may lead to serious misconceptions. On several maps of sea-level pressure contained in this book, the isobars terminate at the 3,000 ft. contour.

Pressure is subject to both periodic and non-periodic variations in time. At sea-level it may vary from about 940 mb.[1] to 1075 mb., a variation equal to about 13 per cent of its absolute magnitude. Such large variations are of a non-periodic character. "Periodic" variations are those which occur at regular intervals according to a definite law of variation: the "period" is the length of time between successive maxima or minima. Periodic variations rarely exceed a departure of 2.0 mb. from the mean value, and are hence of trivial significance except in the equatorial belt, where there are no large non-periodic variations to obscure them.

[1]In tropical revolving storms it may go much lower for a brief period.

There are two well-known periodic variations to which attention should be called. One is the semi-diurnal, or twelve-hourly wave, a curious tide-like oscillation of the atmosphere which encircles the globe. If one can eliminate all other factors, such as non-periodic changes or other periodic variations, the barograph traces at all stations show two maxima and two minima during the day, the maxima occurring at 10 a.m. and 10 p.m. local time and the minima at 4 a.m. and 4 p.m. The effect is greatest at equatorial stations, where pressure at the maxima is more than 4 mb. higher than at the minima. Though it closely resembles a tide, the semi-diurnal variation is believed to be a "resonance" wave set up by the regular travel of the heating effect of insolation round the earth. G. C. Simpson (1918) has suggested that the variation is proportional to the cube of the cosine of the latitude, which accounts for its large size near the equator. The second periodic variation is "diurnal", having a period of twenty-four hours. It is less regular than the semi-diurnal pressure wave, since it depends more directly on the solar heating and on the type of area concerned. Over land areas pressure falls to a minimum during the warm afternoon hours and rises to a maximum in the early hours and up to dawn. As a result of these two factors, semi-diurnal and diurnal pressure variations over land are rather complex and obscure. At sea, however, the diurnal variation is much smaller; pressure tends to rise by day and to fall by night, which is the reverse of the relation holding over land.

In middle and high latitudes the non-periodic variations due to the passage of depressions and anticyclones largely conceal the periodic variations.

Pressure Gradients and Pressure Systems

We have seen that pressure varies both with height and with time. It also varies in the horizontal, so that pressure may be high in one locality and low in another. Between these areas of high and low pressure there exist "gradients" of pressure analogous to the gradients of level which separate hills from vales. The gradients of pressure become visible as soon as we plot isobars on our pressure chart—or, following the analogy—we draw contours along the gradient. These pressure gradients are of overwhelming significance in modern meteorological practice, for they determine the behaviour of the wind.

The distribution of pressure at sea-level is often referred to as the "field of pressure", or pressure field. The pressure field is always found on examination to consist of large, circular or elliptical areas of high and low pressure. These "pressure systems" have a high degree of permanence, for though they change in shape and may move bodily across the earth's surface, they often persist for periods of a week or more. Each pressure system represents a vast eddy in the atmosphere, spinning either clockwise or anticlockwise within the neutral air outside. The names applied to these various types of pressure systems are well known, and they will not be discussed in detail here.

Forces Governing the Winds

Outside the equatorial belt the wind at about 1,500–3,000 ft. above ground level is observed to blow at most times almost precisely parallel to the isobars at a steady speed proportional to the pressure gradient. Nearer the ground the wind blows more slowly, often falling to only half the value at 1,500–3,000 ft.; moreover, the surface wind blows across the isobars towards the low pressure at angles varying from about 20° to 50°. The sense in which the wind blows depends on the hemisphere. North of the equator the wind blows in such a way that an observer with his back to it finds the low pressure on his left; south of the equator the lower pressure is to his right. All these rules depend upon a balance of the forces acting on the air in motion, the state of equilibrium christened the "strophic balance" by Sir W. Napier Shaw (below). Both the forces and the balance between them must be reviewed in some detail.

The operative forces are four in number:

(1) The Pressure gradient force. The pressure gradient exerts a force proportional to its own magnitude: the steeper the pressure gradient, that is, the greater the force. The pressure gradient force acts at right angles to the isobars towards the lower pressure, and tends to give the air an acceleration down the pressure gradient, i.e. to make the air move with increasing

speed down gradient. In careless parlance we therefore speak of air tending to flow towards the lower pressure, a thing which it hardly ever does in the atmosphere.

(2) The Coriolis force, or deflecting force of the earth's rotation. Moving objects on the surface of the earth experience an acceleration for which no very obvious force is available as cause. This acceleration results from the rotation of the earth about its axis, which has the effect of deflecting moving bodies to the right in the northern hemisphere and to the left in the southern hemisphere. The apparent force exerted by the moving earth which brings about this acceleration is called the Coriolis force, or sometimes the deflecting force of the earth's rotation. Its magnitude is given by the expression (per unit mass)

$2\omega \sin \phi . V$, acting at right angles to V

where ω is the rate of spin of the earth
ϕ is the latitude
V is the moving body's speed.

Clearly the force is proportional to the speed of the moving object, though it acts at right angles to the direction of motion. Thus air moving north at 25 miles per hour is subject to only one-half the force exerted on air moving at 50 miles per hour, but in both cases the force acts due east. It is not worth while to try to visualize the physical explanation for this curious force: to do so is to run the risk of hopeless confusion, for it is in reality only an apparent force which arises because we choose to employ a co-ordinate system of rotation with the earth. Since we ourselves rotate about the axis of the earth, the rotating axes look real enough, and so does the deflecting force. An object set in motion in either hemisphere actually does describe a circle if no other force acts upon it—the so-called circle of inertia. Stationary objects are not affected in any way. The force also varies with the sine of the latitude. At the equator, since 0°=0.0, the force is zero, and it rises to a maximum at the poles, where sine 90°=1.000.

(3) Centrifugal force. Air which moves in a curved path upon the surface of the earth is subject to a centrifugal force acting radially outwards from the centre of curvature of the path. This force is proportional to the square of the windspeed and inversely proportional to the radius of curvature. It is of

importance only near the centre of depressions, where strong winds and a high degree of path-curvature are normally found.

(4) Friction. Friction between the moving air and the ground retards the lower atmosphere; there is also a very small internal friction within the air itself if the wind varies with height. In practice, this force, which acts in the opposite direction to the wind direction, is of importance only below about 3,000 ft.

The Strophic Balance

Sir Napier Shaw coined the expression "strophic balance" to denote the equilibrium of forces that exists when steady winds blow across the surface of the earth. The forces we have just discussed could never of themselves produce a steady wind. The pressure gradient force would, if acting alone, produce a violent, accelerated wind whose speed increased with distance, and which blew directly down the pressure gradient. The Coriolis force would constantly change the direction of moving air without increasing its speed. Friction and the centrifugal force exist only when motion has already been established. A steady wind system depends upon a balance between the master force, that due to the pressure gradient, and some other force. A classical treatment of the possible types of balance was given by H. Jeffreys (1922) and the following account follows his plan for the most part.

The first and most significant balanced condition is called the *geostrophic wind*, the name dating back to Sir Napier Shaw. This wind results from a balance between the pressure gradient and the deflecting force of the earth's rotation; if the wind is geostrophic, these two forces are equal and opposite, so that they cancel one another. Mathematically, the balance can be expressed as follows:—

$$2\omega \sin \phi . V_G = \frac{1}{\rho} \times \text{pressure gradient},$$

where the symbols have the same meaning as before, and V_G, is defined below.

This simple expression is one of the most important in meteorology. V_G, the speed of the geostrophic wind, is the wind parallel to the isobars, i.e. at right angles to the pressure gradient. The term *left* of the equal sign is hence the deflecting force of the earth's rotation, and the term on the right is the pressure gradient force for unit mass.

There is perhaps more confusion over this point than over any other among non-specialists. The question is asked: why should the air flow along the isobars? Why should this be the solution chosen by the atmosphere? The answer normally given is that the air begins to move down the pressure gradient, but then experiences the deflecting force of the earth's rotation, and turns to right or to left (according to the hemisphere) until it is parallel to the isobars. This is no doubt a fair interpretation of events, but it bears little semblance to the ordinary state of the atmosphere, for within the latter the air is rarely initially at rest; nor can pressure gradients suddenly be created in this manner. The air normally moves along the isobars at very nearly the geostrophic speed. If the pressure gradient changes, the wind also changes until it has achieved a velocity near the new geostrophic value. A perpetual shift of equilibrium rather than catastrophic change is the custom of the atmosphere.

Let us now glance at the top of Fig. 2 The air at P is subject to a pressure gradient force indicated by the arrow PG. It must therefore accelerate towards the low pressure unless the pressure gradient force is balanced by an equal and opposite force, indicated in fig. 2 by the arrow DF (the length of the arrows indicate the magnitude of the forces). In the case of the geostrophic wind, this counterbalance is provided by the deflecting force of the earth's rotation. It remains, then, to find a wind which will experience a deflection equal to DF. It must plainly be a wind at right angles to DF (since the deflecting force acts at right angles to the direction of motion). In fig. 2 the arrow V_G is the appropriate wind. This wind will experience a deflecting force along DF, and if we chose the correct speed for $V_{G,}$ DF can be made equal in magnitude to the pressure gradient force PG. Plainly then, the geostrophic wind must blow along the isobars, and its speed is determined by the pressure gradient. For every value of the pressure

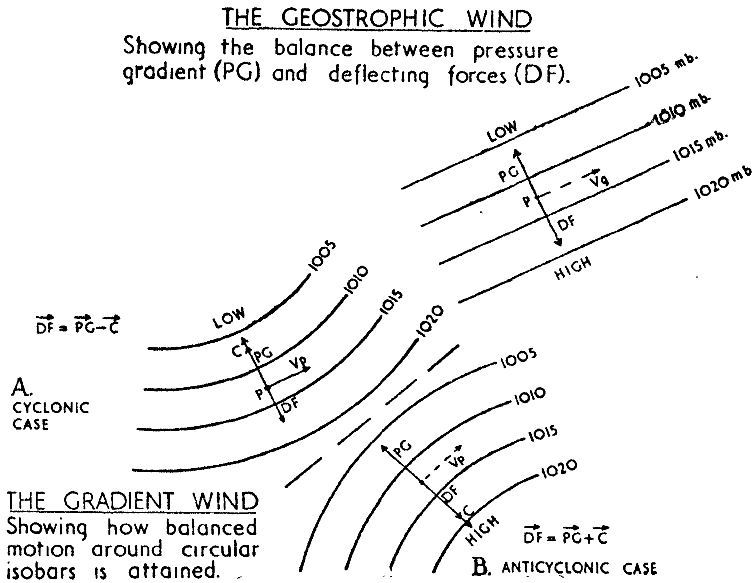


Fig. 2

The geostrophic and gradient winds. The length of arrow is proportional to the magnitude of the forces *only* in the diagram for the geostrophic wind. The other two sketches are only true for direction. The relative magnitudes of the forces are shown by the small equations.

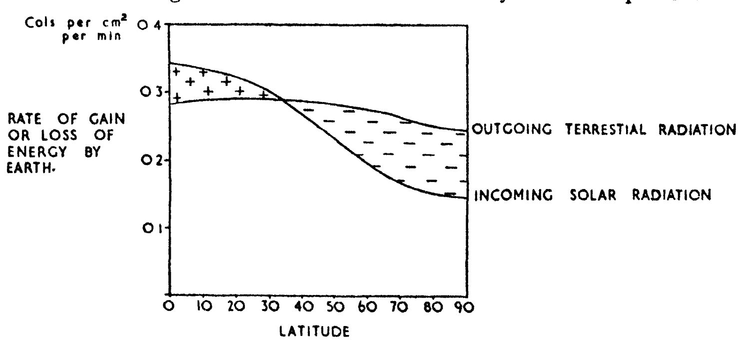


Fig. 3

Latitudinal distribution of solar and terrestrial radiation for earth and year as wholes (following Simpson). These curves are drawn from data computed by very broad assumptions as to reflection and absorption. They show that the earth has a radiative *deficit* poleward from the 35th parallels and a radiative *excess* within them.

gradient there is in the same latitude and at the same density only one possible value for the geostrophic wind.

Provided that the airflow is purely horizontal and straight and no changes of pressure are in progress, the real wind in general is very close to this theoretically derived geostrophic wind. If, however, one or more of these conditions are absent, the wind departs from the geostrophic value. The most readily understood of these departures is that due to curvature of the path of the air particles. If the latter move along curved paths the third force referred to on p. 50 comes into play. A new balance between forces is then reached, and the wind again becomes steady, but with a speed differing from the geostrophic value. This new balanced condition gives rise to what is called the *gradient wind*, which blows when the pressure gradient is balanced jointly by the centrifugal force and by the deflecting force of the earth's rotation.

The term "centrifugal force" usually applied to the effect of curved motion is rather confusing, for it is not a true force at all; it is our clumsy way of trying to express the fact that objects moving along curved paths resume *straight* motion unless we constantly apply a force directed radially inwards towards the centre of curvature. It is perhaps easier to think of the curvature effect as being a constant acceleration of the moving body inwards towards the centre, the so-called *centripetal acceleration*. Shown as arrow C in Fig. 2 (A and B), this acceleration helps to balance the pressure gradient. In case A (cyclonic motion), the pressure gradient force PG is greater than the deflecting force DF by the amount C, the centripetal acceleration; the gradient wind Vp is hence less than the geostrophic wind for straight isobars. In case B (anticyclonic motion), however, the wind is required to accelerate away from the pressure gradient. Hence DF, the deflecting force, must exceed the pressure gradient force by the amount C; thus the gradient wind will have to be greater than the ordinary geostrophic value shown in the upper diagram.

It remains to consider the effect of friction. The latter is confined as an effective force to the bottom layers of the atmosphere, where friction occurs between the wind and the ground. Its chief effect is to slow down the bottom layers of the air. At sea the effect is very small, but over land it is of considerable magnitude. The wind at ground-level is often reduced to as little as one half the gradient or geostrophic wind. Though still blowing in the same general direction as the "free air" 2,000 or 3,000 ft. above the ground, the surface wind makes an angle of about 20° to 50° with the isobars, being directed towards the lower pressure.

Convergence and Divergence

The classes of motion so far considered have all referred to horizontal motion; vertical motion has either been ignored or else specifically excluded, as in the case of the geostrophic wind. The circulation of the atmosphere is in fact overwhelmingly horizontal, so that at first sight this exclusion may seem trivial. But a little consideration will show that this is far from being the case. All appreciable rainfall and snowfall, for example, result from general upward movement of large bodies of moist air. The general circulation of the atmosphere also requires systematic subsidence or uplift over certain parts of the earth. Plainly we cannot dismiss such phenomena as trivial.

In the middle of the nineteenth century, a noted Cambridge mathematician, Stokes, demonstrated that the motion of a fluid (or rather of a small fluid element) was resolvable into four component motions:

(i) a bodily transfer without change of shape or volume;
(ii) a spin about an axis passing through the fluid element;
(iii) a shear, that is, a change of shape without change of volume; and
(iv) an expansion or contraction without change of shape.

This analysis may seem remote from our present purpose, but in fact it is fundamental to an understanding of the circulation of the atmosphere. Component (i) bodily transfer, we have already examined; within the atmosphere it is plainly the dominant component at most times. The transfer of air in the atmosphere we have seen to be carried out chiefly by steady winds which strike a balance between conflicting forces. Component (ii), usually called the "vorticity", is also of importance, but its recognition is too complex for our present purpose. The third component, the shear, is a widespread feature of atmospheric circulation, but it is not of much significance climatologically. As Stokes originally defined it, the fourth and last component, called the divergence, is entirely negligible in the atmosphere. But meteorologists have restricted the idea of divergence to fit a two-dimensional framework: Stokes's divergence was a type of motion similar to that of an expanding or contracting sphere, whereas meteorologists use the word to describe a purely horizontal divergence which is analogous to the expansion or contraction of a circle, a two-dimensional figure. If we use the word in this sense, divergence is the most significant motion of the atmosphere after transfer.

Divergence and its converse, convergence, are radial motions in which speed is zero at the centre and increases outwards. Such simple radial circulation is seen within the atmosphere only on rare occasions in equatorial latitudes. It is, however, very common to see divergence superposed upon transfer. Such areas of divergence and convergence within airstreams are usually areas of pronounced subsidence and uplift respectively. Divergent motion hence tends to give clear, settled weather, whereas convergence creates cloud and rain if the air is damp enough.

CHAPTER V

AIRMASS ANALYSIS: FRONTS AND AIRMASSES

AT the very root of dynamic climatology we find the fertile group of ideas we now call airmass analysis. This analysis consists of the identification of airmasses and fronts, the listing of their properties and the prediction of their movement and modification. The geographer finds the idea of airmasses readily acceptable, for they stand in meteorology much where the region stands in geography. Fronts also have their counterpart in the landscape: the sharp division of Scotland along the Highland Boundary fault results from structural phenomena very like that of the frontal surface. No geographer who has heard of thrust-faulting need fear an acquaintance with a warm front, where atmospheric thrusting is much in evidence. But the parallelism of airmass meteorology and geography rests on more fundamental grounds than crude analogies of this kind. Airmass analysis is an essentially geographical technique. It requires a recognition of regional differentiation and distances, an understanding of the detailed facts of physiography; and above all it has as its essential tool a map, the synoptic chart.[1]

Airmasses are recent additions to the meteorological scene. As early as 1909 W. Napier Shaw and R. G. K. Lempfert (1909) had hinted at their existence, but the idea that they were a major significant feature of the atmosphere as far as static properties were concerned, was developed primarily by a group of Norwegian meteorologists during and after the first world war. This work is outlined in this and the next chapters. We associate it with the name of V. Bjerknes and his son, J. Bjerknes, who, with the aid of men like H. Solberg and E.

[1]It is amusing to note that a problem which required an international meteorological conference to settle was the choice of projection for synoptic charts. The choice—conical orthomorphic with two standard parallels—indicates the prevailing mathematical temper of the delegates!

Palmén, brought about a revolution in meteorological thinking in the years between 1918 and 1930. The present position of airmass theory is based mainly on the work of T. Bergeron (1928), a Swedish scientist who developed the Bergen theories into a practical system of chart analysis. Much is also due to American meteorologists, and especially to H. C. Willett (1944) of the Massachusetts Institute of Technology, whose treatment has been adopted in this chapter. Fronts, on the other hand, have been investigated chiefly by the Norwegians themselves, and a good deal of our present knowledge of frontal behaviour is dependent on the work of S. Petterssen (1940).

Airmasses

An airmass can be defined as *a large body of air in which there are no sharp horizontal variations in temperature and humidity*. An alternative definition is to specify that the air shall be fairly homogeneous horizontally, but the first is better suited to the real facts of the case. We have seen already that temperature and humidity always change with height, so that it can only be in the horizontal plane that we can look for large bodies of homogeneous air. In North America, for example, on a typical summer day both temperature and humidity decrease with height quite rapidly. Temperature at the surface will be in the 80°F.–90°F. zone, but at 12,000–15,000 ft., they are likely to be near the freezing point, 32°F. Yet if we examine a synoptic chart, we find that both the temperature and the humidity are similar all the way from Texas to Quebec, and from Minnesota to Florida. In other words, the south-eastern part of the continent is covered by a body of air which lacks any sharp variations in temperature or humidity over many hundreds of miles, despite the rapid variation of these elements with height.

Airmasses are formed when air rests quietly upon a land or sea surface with a very uniform temperature and (in the case of the land) degree of dampness of surface. An airmass will also be formed if moving air flows for long distances across a uniform surface. Regions where airmasses form are called *source regions*, and it is clear from the definition that they must be situated either in low or high latitudes; in middle latitudes there is usually a rapid decrease in temperature northward. It is also plain that a source region must either be wholly marine or wholly continental, for otherwise the surface would differ in dampness.

Also significant for the formation of airmasses is the type of pressure distribution. Anticyclones are great breeders of airmasses. Their structure tends to create uniformity within the air of their circulation: furthermore they favour slack winds. The ideal source region, then, is an area of uniform surface characteristics where anticyclones are prone to lie.

Airmass Classification

In a well-known monograph, T. Bergeron (1928) showed that it was possible to classify airmasses much as one could classify organisms. He based his classification on source regions, grouping the airmasses according to whether they were of polar or tropical, continental or maritime origin. The classification has won international acceptance, and it is presented here in the slightly revised version now current. The letters in brackets are the labels by which the airmasses are identified on the synoptic chart. It would be tedious to write out the full name of the airmass every time it is marked on the chart, or even mentioned in written discussion. The labels are used without explanation in the text from this point onwards.

The Classification of Airmasses

Major Group	*Sub-group*	*Source Regions*	*Properties at Source*
Polar (P)	Maritime Polar (mP)	Oceans north or south of approx. 50°N.	Cool, rather damp, unstable.
Polar (P)	Continental Polar (cP)	Continents in vicinity of Arctic Circle: Antarctica.	Cold and dry. Very stable.
Tropical (T)	Maritime Tropical (mT)	Trade-wind belt and sub-tropical waters of great oceans.	Moist and warm. Stability variable: stable on east side of great oceans, rather unstable on west.
Tropical (T)	Continental Tropical (cT)	Low-latitude deserts, chiefly Sahara and Australian deserts.	Hot and very dry. Unstable.

Some meteorologists divide the Polar Airmasses into Arctic (A) masses, originating over a snow or ice surface, and

Polar masses originating over cold, bare land or water. Bergeron himself and others have differentiated Equatorial (E) air from the genuine Tropical. These refinements are, however, of little practical value in middle latitudes, which are the main scene of airmass analysis. Another airmass sometimes separately identified is Superior (S) air, the warm and dry air formed chiefly by subsidence above sub-tropical or middle latitude anticyclones. Though this is essentially a high-level mass, it sometimes appears at ground level in North America and eastern Asia. On the whole, however, the fourfold classification presented in the above table is adequate for general purposes.

A moving airmass is often called an airstream. As an airstream moves away from its source region, it undergoes modifications of two main kinds. Further sub-division of the main types are possible on the basis of this modification:

(i) *Thermal modification* arises when an airstream moves over a surface different in temperature from the source region. A polar airmass moving southwards, for example, will be cooler than the surface over which it passes, and will hence be warmed up from below; and a tropical airmass moving northward will usually be cooled at the surface. An airmass which is warmer than the surface over which it passes is given the suffix W—e.g. cTW. Such air is rendered increasingly stable, for the surface cooling leads to the development of an inversion or isothermal layer in the lowest levels. An airmass colder than the surface over which it passes is labelled K (after the German "kalt")—e.g. mPK. Such airmasses tend to become unstable since the addition of heat from below constantly steepens the lapse rate until convective overturning is violent enough to take away the added heat.

A by-product of thermal modification is an increase or decrease in humidity. Any K airmass passing over a warm water surface or even a moist land surface (especially when covered with growing plants) rapidly picks up moisture; on the other hand, precipitation from an airmass lowers its humidity. No special label is used to identify these modifications.

Absence of both the W and K suffices indicates that the temperature of the airmass and the ground surface are similar.

(ii) *Dynamic modification* results from the movement of the airstream, and from its relations with nearby anticyclones and depressions. Convergent motion leads to a general lessening of the airmass stability, so that an airmass which forms part of the circulation of a depression is likely to be less stable than an airmass from the same source outside the depression. Similarly, airmasses which are involved in anticyclonic circulation are normally more stable than the same mass elsewhere. These dynamic effects operate chiefly in the middle and upper troposphere. At lower levels the effect of surface heating is mainly responsible for changes in stability. In the airmass classification these low level changes are looked after by the W and K suffices referring to thermal modification. Willett (1944) has suggested that we need an extra term in the airmass label to allow for the stability aloft. He proposes the suffix "u" after the W or K suffix in cases where there is high level instability, and "s" where there is stability. Though this suggestion has not been widely adopted, it has a real importance in practical work. *mTWu* would in Willett's scheme signify maritime tropical air undergoing surface cooling, but with instability above the surface layers.

Properties of Some Common Airmasses

In the table presented above there is a brief reference in the final column to the characteristic weather of each of the main airmasses. A slightly more detailed account follows.

Continental Polar (*cP*). This air originates over the northern parts of North America, Asia and European Russia. In winter these areas become intensely cold because of prolonged radiative cooling and remoteness from warm seas. Anticyclones of the cold type tend to form very frequently over these cold areas, and the cP airmass is generated within the high-pressure systems. In summer the anticyclones are less frequent, and so, therefore, is cP air; the polar or sub-polar regions of the continents are then more often covered by air from other sources. Furthermore, these regions are quite warm for a brief period in summer.

In winter, the cP air consists of a shallow layer of very cold and dry air which underlies a strong inversion at levels up to 10,000 ft.: hence the great stability of this airmass. Above the inversion, the air is much warmer, though it is also dry owing to subsidence. cP air moving away from its source region takes its cold and dryness with it as long as it moves over land, but over warm seas it rapidly changes to a damp, unstable type closely resembling mP air. Weather is clear, cold and sparkling in this airmass over land. It reaches Britain much modified; usually cP air from Russia and Siberia breaks out across Europe once or twice in a winter, bringing clear skies and sharp frost to Germany, the Low Countries and France. To reach Britain, however, it has to cross the warm North Sea. Strato-cumulus clouds form beneath the inversion, and temperatures are raised to the freezing point or above.

cP air in summer is quite different. Often fairly warm at the surface, it remains cool aloft, and is hence much less stable than in winter. It gives pleasant bright weather with widely scattered cumulus cloud in most areas. Sometimes it becomes involved in the circulation of depressions, and it is then quite common to find cumulo-nimbus cloud and showers or thunderstorms developing in the afternoon. It does not affect Great Britain in summer.

Maritime Polar (*mP*). This is the commonest airmass over the British Isles and the North Atlantic north of the subtropical anticyclone. The name refers to air which has acquired its characteristics over the northern parts of the great oceans of the northern hemisphere or the stormy belt in similar latitudes in the southern. Prolonged quiescence in these areas is rare, and air described as maritime Polar is nearly always (in the northern hemisphere) continental Polar or Arctic air which has had a long sea-track in high latitudes. Such air is rapidly heated and moistened at the surface where in contact with the relatively warm sea, and it is converted into the unstable, rather moist air we call maritime polar.

In western Europe mP air brings a variety of weather, depending mainly on the direction from which the air came. The air behind the cold front or occlusion of Atlantic depressions is mPKu, to give it its full label—cool, moist, unstable air packed with cumulus cloud and rain showers over the sea, and by day over the land as well. This characteristic type of weather—April showers are a good case of it—is a year-round type in western Europe. Often, however, mP air reaches Great Britain, Scandinavia and the Low Countries from south-west and it may then be either neutral or warmer than the sea and land, especially in winter. Such mPW air may be either stable or unstable aloft, but it is always stable at low levels when it crosses the coast. In winter it gives cloudy days with stratus cloud and drizzle, but in summer weather is often clear; surface heating by day may even give thunderstorms over land.

Occasionally fronts will form between two mP airmasses over the North Atlantic, and many of the Atlantic depressions which affect Britain have warm sectors of mPWu air rather than mT (maritime tropical).

Continental Tropical (*cT*) is a rare airmass, formed chiefly over low latitude deserts. cT air from the Sahara flows southward into West Africa throughout the year, and occasionally enters the Mediterranean. It also occurs in summer on the high plateaus of the southwestern U.S.A. and Mexico. Its properties are excessively high temperatures and instability. Weather is, however, cloudless because of the extreme dryness of the air. Dust haze is very common.

Maritime Tropical (*mT*) is the major exported airmass of the tropics. Originating in the trade wind belt and the subtropical anticyclones over the oceans, it penetrates polewards in broad invasions, chiefly in the warm sectors of travelling depressions. In its source region it is typically an airmass of fairly high temperatures and humidities, of clear skies and of stability, with an inversion usually present at some level.

mT air reaching western Europe occurs chiefly in the warm sectors of depressions, but sometimes a broad flow occurs round the north side of an anticyclone centred over France or Spain. It is usually of the mTWs variety: that is, it is undergoing rapid cooling at the surface and is stable aloft. A pronounced inversion is present at quite low levels, beneath which there are continuous sheets of stratus, strato-cumulus or even of sea fog caused by the cooling over cool waters. This stratus cloud penetrates deeply inland in all seasons, though it tends to break up during the day in summer except along the coasts. Above the inversion there may be layers of alto-cumulus, but more often the upper layers are dry and cloudless. mTW air gives some drizzle, but falls of rain are surprisingly small even along fronts due to the great stability at high levels. The mT air affecting the west coast of North America is very similar.

The mT air which enters North America and east Asia round the western sides of the oceanic sub-tropical anticyclones tends to be warmer, damper and less stable. In winter it is of the mTWu variety, bringing wide areas of low stratus to these regions, but heavy rains or snows occur if the air is forced to rise over hills or frontal wedges. In summer the hot sun tends to make mTK the commoner type. This gives the characteristic hot, sunny, thundery weather of the Asiatic monsoon and of the eastern U.S.A. and Canada.

Fronts and Frontogenesis

If we recognize the existence of airmasses, characterized by homogeneity over wide areas, we must also give consideration to the boundaries which separate the airmasses. Such boundaries must plainly be zones of fairly rapid transition from one characteristic set of airmass properties to another. Sometimes the transition is gentle, but more often we find it quite abrupt. Close examination of individual cases shows that two unlike airmasses are often separated by a sloping boundary surface, which we call a *frontal surface.* The line along which this surface intersects with the ground surface is termed a "*front*". Sometimes we find an intermediate condition in which airmasses are separated by a zone too narrow to be regarded as a gradual transition zone, but too wide to be regarded strictly as a front; we then use the term "frontal zone".

Fronts and frontal surfaces have special dynamic properties which need careful weighing. Two liquids of unlike density which do not mix with one another are separated by a horizontal surface of discontinuity analogous to fronts in the atmosphere. Helmholtz (1888) pointed out before the turn of the century that on a rotating earth two fluids in steady motion might be separated by a *sloping* surface which would intersect the ground surface. Margules (1906) gave a simplified relationship for the equilibrium angle of slope of such surfaces which expressed the slope as a function of the velocities and temperatures within the two airmasses and of the sine of the latitude. It is of the form

$$\tan \alpha = \frac{-2\omega \sin \phi}{g} \; \frac{T_2V_1 - T_1V_2}{T_2 - T_1}$$

where α=the angle between the frontal surface and the horizontal plane
ω=angular velocity of the earth's rotation
g=acceleration of gravity
ϕ=latitude
T=temperature
V=windspeed, the suffices referring to the two airmasses 1 and 2 (i.e. cold, 1; warm, 2)

Since sin 0=0, it is clear that at the equator the angle of slope will be zero: only in latitudes some distance away from the equator is a sloping front possible in an equilibrium condition. In middle latitudes the possible values for T and V give slopes varying between about 1 in 30 and 1 in 200. Moving fronts have slopes which may be somewhat different from these equilibrium values. The Margules formula really only applies when the wind flows steadily on both sides of the front and *parallel to it.* It is clear, however, from these results that we should expect the cold air to act as a wedge which underlies the warmer mass.

Frontogenesis is the term we use for the processes which lead to the formation or intensification of fronts. These processes depend on two factors:

(1) *The Geographic Factor.* Other things being equal, the formation of fronts is favoured when there is a close juxtaposition of unlike surfaces. In winter, for example, there is a striking contrast between the cold, often snow-covered surface of North America and the warm Gulf Stream and North Atlantic Drift offshore. Similar conditions apply off eastern Asia, off northern Europe, and in several other localities. The dynamic factor in frontogenesis about to be discussed will only produce fronts if there are initially gradients of temperature within the airmasses involved, and these gradients of

temperature are best developed under the geographical condition described.

(2) *The Dynamic Factor*. Certain types of flow are necessary if fronts are to develop. These are discussed at length below.

The dynamic background of frontogenesis was first comprehensively studied by T. Bergeron (1928), and the implications of his work were more fully developed by Petterssen (1940), to whom we owe much of our modern views on front formation. Broadly, what is required is some type of air motion which will lead to a concentration of isotherms.

The type of flow which most readily leads to frontogenesis is what kinematicists call a deformation field. Fig. 4 shows the combination of two components of deformation, dilatation and contraction, in such a way as to lead to the formation of a saddle-point in the atmosphere. It will readily be seen that this roughly corresponds to the type of flow which occurs round "cols" in the atmosphere, that is, in the neutral ground between two highs and two lows. The model deformation field shown in fig. 4 has two axes at right angles. Parallel to the *axis of outflow* the air flows away from the central point, whereas parallel to the other axis (that of mass inflow) the airflow is towards the centre. It will readily be seen that an isotherm initially along the lines AB or CD will be carried towards the axis of outflow. Along this axis, then, there will be an increasing concentration of isotherms, i.e. a front will be created in the position of the axis. A little thought will show, however, that this will only be true if the isotherms were initially roughly parallel to the axis of outflow. Petterssen (1944) showed that frontogenesis does not occur if the angle between the isotherms and the dilatation axis exceeds 45°.

Fronts may also be formed in other types of air movement. Kobayasi (1923) showed that a cyclonic vortex (viz. a circular depression) which moved rapidly through air containing a temperature gradient would develop a shallow cold front (but no warm front) if the temperature gradient was at right angles to the path of the vortex. Though Kobayasi's arguments were restricted in character and rested on some dubious assump-

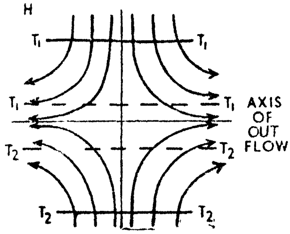


Fig. 4

A frontogenetic deformation field.

Fig. 4. The isotherms T_1 and T_2 approach one another at the axis of outflow, where a front will form.

Fig. 5. The behaviour of isotherms and isobars at fronts.

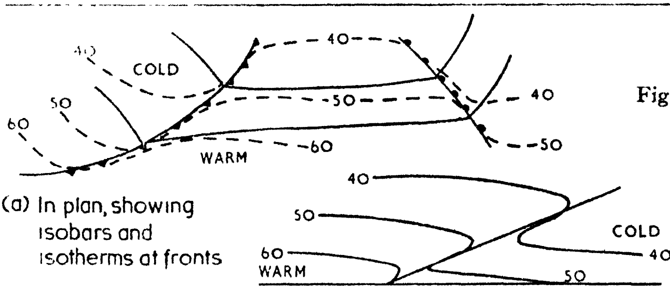
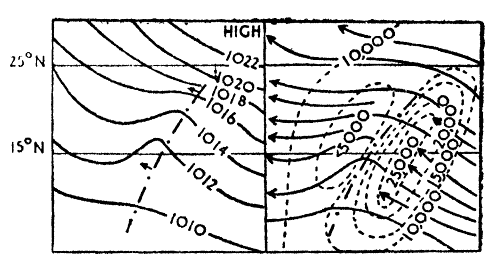


Fig. 5

(a) In plan, showing isobars and isotherms at fronts

(b) In section, showing isotherms at front



A "Wave in the Easterlies" (after Institute of Tropical Meteorology Puerto Rico). Dotted lines are contours (in feet) on top of moist layer.

Fig. 6

tions, it is true that non-frontal depressions which move quickly through air in which there is a temperature gradient usually develop a pronounced front of the cold type in their rear gradient. We shall see later that such fronts may be important in several parts of the world.

Just as there are certain classes of air movement which favour the formation and intensification of fronts, so also there are conditions inimical to their continued life. Space will not be devoted here to a detailed description of such conditions. We may note, however, that divergent motion such as one finds in anticyclones is hostile to either the formation or the maintenance of intensity of fronts. If fronts become involved in the circulation of anticyclones, they progressively decrease in intensity, and in a day or two may entirely disappear, being replaced by broad transition zones.

Some Important Properties of Fronts

There are many properties of frontal surfaces which require our examination. Some of these are briefly listed below:

(1) *Depth.* The type of motion which leads to frontogenesis is quite common at low levels, but becomes increasingly rare with height above the ground. Well-marked frontal surfaces are hence commonest at quite low levels. They can sometimes be traced on the 10,000 ft. chart, but rarely at heights above 15,000 ft. The disturbance of flow extends to much greater heights than this, however.

(2) *The Front as a Discontinuity.* A frontal surface is often referred to as a discontinuity surface, by which is meant that certain properties which are continuously and evenly distributed throughout the airmasses on either side are completely different on the two sides of the front. Elements which always show this discontinuity are temperature, wind direction, pressure gradient (but *not* pressure) and barometric tendency.

The discontinuity in pressure gradient and in wind-direction reveals itself in the behaviour of the isobars as they cross the front. They exhibit a change of direction in such a sense that they turn in the direction of the lower pressure. Fig. 5 illustrates the point. The "kink" or V-shaped trough so produced is *always* concave towards the low pressure. Since the wind blows roughly along the isobars, it follows that the wind ahead of the front must always make an angle with that behind. Before the days of frontal analysis, the kink was often smoothed; many troughs of low-pressure which appeared as rounded troughs on the old charts should have been V-shaped. Not all troughs are, however, of frontal type, and these exceptions are in general of rounded form. It is clear that when a front passes the observer, there must be a sudden shift in wind: in the northern hemisphere it will always *veer*, that is, shift in a clockwise sense.

(3) *Vertical Motion at Fronts.* There is a persistent tendency for the warm air lying over the wedge of cold air beneath to rise up the slope of the wedge, thus adding a vertical component to its motion. This upslope motion of the warm air arises either from the advance of the cold wedge (which then undercuts more warm air) or from a motion of the warm air towards the front. It is this upward thrust of the warm air which causes most of the cloud and precipitation along fronts. This vertical motion is discussed at greater length in the next chapter.

(4) *Fronts in Motion.* So far, we have discussed fronts as though they were stationary lines on the chart. Though we do find stationary fronts on the chart at times, they are more often in motion, sometimes at speeds of 30, 40 or even 50 miles per hour. They move across great areas in the course of twenty-four hours. Since fronts are nothing more than boundaries between airmasses, their motion implies that the airmasses themselves are moving; at all points passed by the moving front there must plainly be a change of airmass and of general weather conditions. If the airmass replacing the other is warmer than its predecessor, we call the front a *warm front*; a *cold front* is one whose passage brings about the arrival of colder air. Note that by the term "warm" and "cold" we do not necessarily imply a striking difference in the properties of the fronts themselves, but merely that they are moving in a particular direction relative to the airmasses concerned. We shall see in the next chapter that the formation of the travelling depressions of middle and high latitudes creates warm and cold fronts.

CHAPTER VI

THE DEPRESSIONS OF MIDDLE LATITUDES

THE large, travelling low-pressure systems called depressions are largely confined to middle and high latitudes; they rarely occur within the 30° parallels of latitude, so that one-half of the entire globe is free from them. In the higher latitude zones, however, they form the main features of the synoptic chart, and are the real arbiters of day-to-day weather changes. Since they bring the rains to the main populated areas of the earth, their geographical significance is enormous. Some geographers have argued that the passage of such storms is markedly stimulating to man, and that their importance is as much psychological as economic. Among those who follow the teaching of Ellsworth Huntington it has even been fashionable to talk of "cyclonic man", an abstract being who is capable of living a civilized life because of prolonged exposure to the passage of depressions. Though such ideas are highly speculative, there can be no argument about the importance of the travelling depression to the geographer.

The existence of the extra-tropical depression has been known for about eighty years, but our present views about its formation and structure date back only to the last war, when the Bergen group of scientists began their systematic study of fronts, which led to the now famous theory of frontal wave-formation.

There was some conservative reluctance on the part of the official forecast services of the world to accept these revolutionary ideas and methods; it was not until 1934, for example, that the British Service began drawing fronts on its working charts. But the obvious superiority of the new methods as tools of explanation began to carry great weight, and they are now universally accepted. The essential truth that many depressions form as waves on frontal surfaces has been put to the test on thousands of synoptic charts. Once the opposition gave way, the general practitioners among meteorologists tended to swing to the opposite extreme; the limitations of the frontal theory were overlooked, and what was perhaps worse, fronts on the charts began to multiply until they were almost as common as isobars. The word "front" itself has been so abused that its essential dynamic implications are often ignored. A "front" is put on the chart to explain a precipitation area, which is a bad case of putting the cart before the horse. The writer can recall hearing no less an authority than J. Bjerknes (at a meeting in the Air Ministry, London, in 1944) deplore the number of fronts which appear on the average synoptic chart.

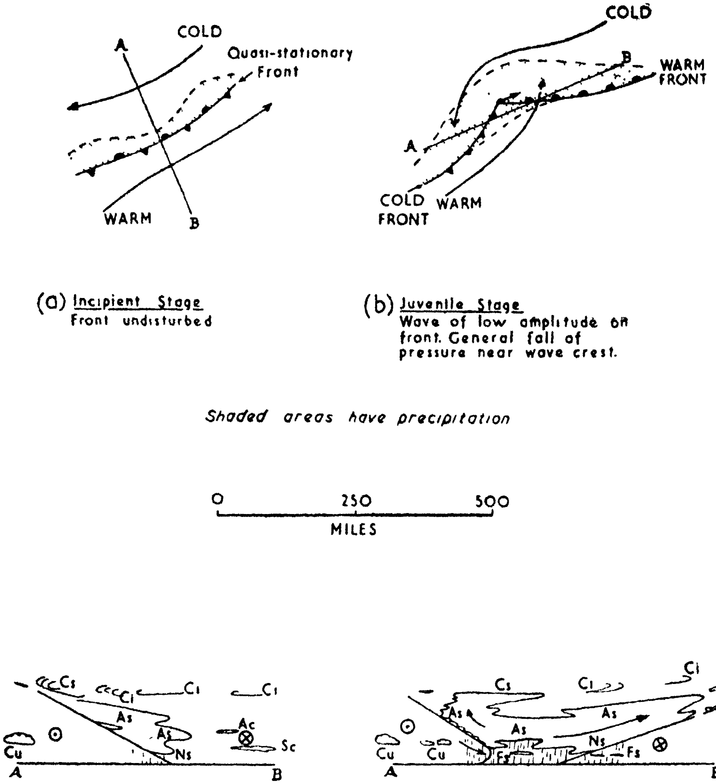
The main limitations of the frontal theory are:

(i) the fact of changes of pressure and of the occlusion process, neither of which is adequately explained, and
(ii) that many depressions do not appear to originate in this manner.

There can be no doubt of the *fact* of the growth and decay of the low-pressure system associated with the frontal wave, but at present the theory does not completely explain this development. Nor can there be any doubt that there are many active depressions which originate in other ways, though this fact was overlooked in the years of enthusiasm following the general acceptance of the Bergen views. The followers of the theory strained it beyond its limits in seeking to show that it could account for all cyclonic development. Of those types to which it cannot so be extended, we can number the polar air depression, the lee-depression (see below, p. 82) and certain developments at the surface which respond to disturbances in the upper air.

The Frontal Wave Theory of Cyclogenesis

The account given here of frontal wave formation is descriptive: for an account of the mathematical theory of wave formation or of Bjerknes' perturbation theory the reader is referred to papers or books by V. Bjerknes and collaborators (1933) or to excellent summaries in English by B. Haurwitz



a–d. STAGES IN THE DEVELOPMENT OF A FRONTAL DEPRESSION. EACH OF THE FOUR STAGES ARE 24–36 HOURS APART, AND SHOW THE DEVELOPMENT OF THE STORM FROM BIRTH TO FULL MATURITY. A SECTION IS DRAWN ACROSS THE FRONTAL BELT IN EACH STAGE TO SHOW THE CHARACTERISTIC CLOUD FORMATION.

Fig. 7

(1941) and S. Petterssen (1940). The account given here very largely omits any consideration of the upper air: for such materials, reference should be made to studies by J. Namias and collaborators (1940), H. R. Byers (1944), H. C. Willett (1944), and other American writers.

Fig. 7 summarizes the main stages in the development of

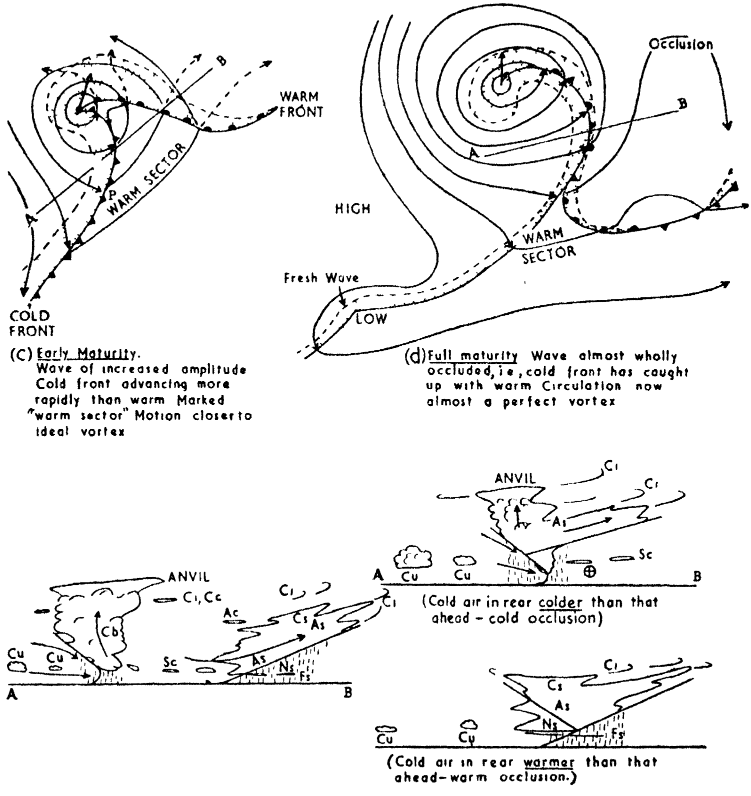


Fig. 7

a frontal depression. Note that these represent *facts*, not theory, for this cycle of events has been observed thousands of times. The four stages illustrated (a), (b), (c) and (d) may be regarded as being about 24–36 hours apart, and the wave crest (or depression centre) will have travelled about 500–1,000 miles in the direction indicated by the arrow at the crest between successive stages. The precipitation area is shaded. The wind arrows are drawn as streamlines, that is, they are at all points parallel to the motion of the air. They roughly correspond to isobars, though in a developing depression the wind does not quite follow the isobars, especially near the direct track of the centre. In some of the diagrams the path of the warm air over the wedge of cold at the warm front is indicated by dashed streamlines.

(a) *The Incipient Stage.* A quasi-stationary front is seen separating cold north-easterly and warm south-westerly currents. The cold air forms a gently sloping wedge beneath the warm. Since neither the cold nor the warm air has any component of motion towards the front, the latter remains stationary. There is as a rule a considerable cloud-mass within the warm air just above the frontal surface. The cloud is usually stratiform, consisting of a layer or layers of alto-stratus, often with alto-cumulus towards the edge of the cloud-mass. Light to moderate precipitation may fall from the clouds, apparently as a result of slow movement of warm air up the surface of the cold wedge.

The condition shown in the diagram, where the cold air is flowing from the opposite direction to the warm, is the typical one, but waves may also form on fronts between two dissimilar westerly currents. In this case, the warm air flows more quickly than the cold. Whether the cold air flows from west or east, however, the existence of such a quasi-stationary front is a sign that wave development is highly probable. Well-marked fronts rarely remain undisturbed in this condition for long.

(b) *The Juvenile Stage.* The earliest sign of wave development is usually a broadening and intensification of the precipitation and cloud areas. It is then seen that a broad, wave-like indentation of the front is taking place, the warm air driving a long, shallow salient into the territory of the cold. The rain and cloud often extend into the warm air near the crest of the salient. A general fall of pressure takes place all round the wave, especially in the cold air just in front of the crest. The front is now strictly speaking divisible into "warm" and "cold" sections, the warm front lying ahead of the invading warm air, the cold front marking the leading edge of the flanking assault of the cold air (see fig. 7). In this stage there is little difference between the warm and cold fronts as to associated cloud and weather. These result from the slow, upsliding motion of the warm air over the cold wedge. At the warm front the motion is gentle but continuous, and is distributed through a considerable depth of the warm air. There is a similar upward movement at the cold front, where the cold wedge is beginning to undercut the warm air ahead. A description of the typical cloud and weather is deferred until the next stage.

The newly formed wave travels rapidly along the front, in the same general direction as the warm air. There are in general two classes of such waves, depending on their size and subsequent development: (i) In cases where the *length* of the wave is less than about 300 miles, the wave is "stable", that is, it travels along the front without getting longer or deeper. Such "small waves", as they are called, cause an intensification of bad weather, but do not result in cyclogenesis. They very often pass over Britain from the south-west along cold fronts which have become stationary over the country. (ii) Larger waves (wave length greater than 300 miles), however, usually increase rapidly in *amplitude*, i.e. the warm invasion into the cold air's territory gets deeper. The fall of pressure intensifies, and a closed centre of low-pressure—a young "depression"—comes into being on the crest of the wave.

Note that the word "wave" is used here in much the same way as we use it in describing disturbances of the sea surface. The travel of the wave along the front resembles in some degree the travel of an ocean wave nearing a shelving sea-shore. Since the frontal surface is a tilted plane, beneath which the cold air appears as a wedge, it is readily seen that the type of disturbance which gives a wave-shaped deformation on the front (the projection of the frontal surface on the ground or sea-surface) is a trough in the surface. It is the movement and deepening of this trough-like disturbance which leads to the movements of the front at ground-level.

(c) *Early Maturity.* In this stage (usually reached in some 36–60 hours from the incipient stage) the wave has increased to much greater amplitude; the warm air now forms a broad and deep "sector"—the warm sector, as it is called—within the rapidly growing depression. The latter has by this time taken up a roughly circular form, though the isobars within the warm sector tend still to remain straight. The dual nature of the frontal system is now very much in evidence; a clearly defined warm-front leads the advancing warm sector and an even sharper cold-front marks the leading edge of the flanking assault of the cold air. For it is now clear that the cold air is undercutting and attacking the warm, the effect being to narrow down the area of the warm sector. As this sector gets deeper, then, its area is diminished by the outburst of cooler air behind the centre of the depression. Note that there are now two distinct areas of upward movement of air; at the warm front, the warm air slides gently over the retreating cold wedge, because the warm air approaches the front more rapidly than the cold retreats; and at the cold front the cold wedge in advancing undercuts and forces up the warm air ahead. A highly characteristic sequence of clouds and weather is associated with both fronts. At the warm front, the upward motion is on the whole slow, continuous and non-turbulent. It takes the form of an upward sliding, and not of violent uplift. The slope of the warm front is usually gentle—of the order of 1 in 100 or less—so that the area of cloud is often quite wide. The first sign of an approaching warm front is wispy cirrus moving from some westerly point. Later the cirrus thickens into a sheet of cirro-stratus, which in turn is followed by a layer or layers of thick, impenetrable alto-stratus from which snow, sleet or rain falls. Near the front itself the base of this stratiform mass is often as low as 2,000 or 3,000 ft., below which there is usually low, ragged fracto-stratus down as low as 500 ft. or less. The formless mass of grey rain-clouds which is what we see of the base of this thick frontal cloud is what we call "nimbostratus". Rain, sleet or snow falls in a band which may extend some 120–200 miles ahead of the front. Near the front itself the rain may give way to drizzle; contrary to an earlier view, frontal precipitation usually starts with fairly big drops, and drizzle is more usual near the front itself.

It used to be thought that warm front cloud formed a fairly solid, wedge-shaped mass extending from the cirrus level down to the lowest layers: it was believed that we saw the lower surface of this mass, which accounted for the sequence of successively lower cloud types that we saw as a warm front approached. We know now, however, that the cloud is more often disposed in horizontal layers. An aircraft ascent through the middle of the rain area normally passes through several layers of nimbo-stratus or alto-stratus, each up to 3,000–5,000 ft. thick. Between the layers there are "lanes" of clear air. The top of the highest of these alto-stratus layers is usually about 15,000 ft., above which there are only thin layers of cirrus or cirro-stratus. Most of the rain or snow comes from the alto-stratus at or above the freezing level.

Sometimes the steady uplift of the warm air releases potential instability, and isolated thunder-clouds (cumulo-nimbus) may form within the general frontal layers. A few claps of thunder and a period of heavier rain occur as these clouds pass over. Such events are common on the eastern side of the continents, where the warm, moist airmasses from the Atlantic or Pacific are apt to be unstable. Quite a severe storm of this character occurred in Montreal on a night in March 1946, though the surface temperature was only 23°F. Similar cases do occur in Europe, however, especially in storms involving air of Mediterranean origin. In Britain the shallow depressions of summer which move northwards from the Bay of Biscay or France often give thunderstorms along their warm fronts. Good examples occurred on August 22nd–24th, 1935, and C. K. M. Douglas (1934) has recorded an instance as late in the year as November 19th.

Within the warm sector of depressions weather is highly variable. Since the air is moist and warm, stratus or strato-cumulus clouds are common, and there may be drizzle. In Europe there is usually little really bad weather, though showers or even thunderstorms (in summer only) may occur just ahead of the cold front. In eastern North America and the Far East, however, showers and thunderstorms are quite common in warm sectors.

At the cold front, uplift of the warm air is more violent. An advancing cold front has a steeper slope (1–50 is typical) than a warm front, and its upward prying of the warm air ahead is often fairly violent. Because of the friction, the surface layers of the cold air lag behind those 1,500–2,000 ft. above, which hence advance ahead of the lower layers. This gives the front an overhanging nose, which periodically breaks up with severe turbulence. It is this which gives a cold front passage its gusty character.

The characteristic clouds of the zone of uplift along a cold front are cumulus and cumulo-nimbus. With these are associated showers or thunderstorms, the latter more particularly in summer. The front should not, however, be regarded as a long line of thunderstorms stretching from the centre of the depression right out to its perimeter. Near the centre itself there are usually extensive sheets of alto-stratus, and steady rain resembling that of the warm front may fall for a fairly long period. Once away from the central zone, however, the characteristic cumuliform, showery weather is found. At a considerable distance from the centre the cold current behind the front tends to adopt anticyclonic curvature (i.e. the streamlines become concave to the right of the motion). On fig. 7c this is true south-west of the point P. In this zone (which is not always present) there is a marked tendency for stratiform cloud to reappear, and for the rain-belt to get wider and steadier once more. The development of such anticyclonic curvature is often a sign of renewed wave activity.

Because they are associated with a long line of storms with gusty winds from a north-westerly point (south-westerly in the southern hemisphere) cold fronts used to be known as "line-squalls", a name which has become little used.

In the cold air behind the depression centre, weather is usually bright, with abundant cumulus cloud and some showers—at least in Europe. North and east of the centre it is more usual to find small cumulus or strato-cumulus layers beneath the cirrus or alto-stratus sheets of the approaching fronts.

In their early mature stages, frontal depressions usually travel rapidly, speeds of 30–40 miles per hour being common in winter, and 15–30 miles per hour in summer. There is a marked tendency for them to acquire an increasing northward component in their direction of motion. In their early stages, such storms usually travel eastwards or north-eastwards (though other directions are not uncommon) but with increasing age they have a marked tendency to turn northwards. Most depressions which disobey this law are either of highly complex structure or are non-frontal.

(d) *Full Maturity and Old Age*. The attack of the cold air against the left flank of the warm sector gathers increased momentum in the later mature stage, and there soon comes a time when the advancing cold front catches the warm front up —that is, the warm air is entirely lifted off the ground. The advancing cold wedge has caught up its own retreating part, and the cold airmass now forms a continuous layer at the surface. The warm air is still present aloft, but as the process continues, it is lifted further and further upwards. This whole process is known as "occlusion", and the frontal type which results from it is known also as an occlusion.

Occlusions vary in structure. The cold air behind the cold front of a depression has usually pursued a widely different path from that which lies ahead of the warm front. There is usually a considerable temperature contrast between the two parts of the airmass, and when they come into contact a front is created between them. In fig. 7d (upper section) the air behind the front is colder than that which lies ahead; consequently it is cutting underneath the cool air ahead as well as under the uplifted warm air. Such *cold occlusions*, as they are called, very soon come to resemble normal cold fronts, and the warm front layer cloud aloft becomes rapidly less extensive after occlusion. Cold occlusions predominate over the eastern sides of the great continents, where the airmass behind the cold fronts of depressions is polar continental, which is usually very cold and dry. Over the eastern half of the Atlantic, north-west Europe and also the Pacific coast of North America one sometimes finds polar continental air *ahead* of the occlusion, and polar maritime behind. In such cases the air behind the front is usually warmer than that ahead, over which it slides just as the air of the warm sector did. *Warm occlusions* of this kind (see fig. 7d) give weather like that of warm fronts, the stratiform rain clouds often extending into the cold air in the rear. It is probably not true, as Bergeron (1928) claimed, that the majority of occlusions affecting north-west Europe are warm occlusions. Most of those which cross the British Isles are definitely of the cold type, though they frequently change in character by the time they have reached the Baltic or the Alpine-Carpathian foreland.

Proof that the warm air is present aloft in an occlusion can be obtained from aircraft ascents or radiosondes, but it is sometimes forthcoming in a much more dramatic manner. In eastern North America, it is common to find occlusions in which the cold air on both sides of the front is well below the freezing point, despite the fact that steady rain is falling. This rain falls from the warm air in the trough of the occlusion into the freezing air below. As it hits the ground, it freezes, coating buildings, telegraph poles and aircraft in flight with a thick film of clear ice. Such ice-storms, as they are called, can do immense damage. Sometimes they occur also at warm fronts, but they are then usually followed by a thaw. They are not unknown in Britain.

In the occluded stage the depression is at its greatest depth and intensity, and may be regarded as having attained full maturity. Its subsequent history is usually one of rapid decline. The occluded depression is roughly circular, forming a fairly complete vortex that may be as much as 1,000–2,000 miles in diameter—though smaller dimensions are more usual. For about twenty-four hours after occlusion the depression continues to move—often in a northerly direction. It then slows down and moves only slowly and erratically in its old age. As a rule, a general rise in pressure begins in the central area, and the depression "fills up", as the jargon of the meteorologist puts it. In perhaps another twenty-four hours the former giant has been almost completely removed from the chart. The old occlusion undergoes frontolysis, and soon there is little trace of bad weather.

As the occluded depression becomes more and more a vortex consisting wholly of polar air, the weather characteristic in the mature stage of its rear parts, tends to spread all through its circulation. The large and deep occluded depressions which become stationary off our north-west coasts often consist exclusively of polar maritime air, the warm sector having been removed long ago. They bring showery, bright weather of the typical polar maritime airmass to large areas, including the whole of the British Isles. Such storms persist much longer over sea areas than over land, and Britain often experiences several consecutive days of showery weather with south-westerly or westerly winds from this circulation type.

Families of Depressions

A fresh wave often forms on the trailing cold front of a depression, and as the new wave matures into a depression, yet another may form in turn on the front in the rear of the second member. A large enough synoptic chart may show four or five storms strung out along a single frontal system; the depression furthest east is usually fully occluded, and that furthest west is a very juvenile wave. Such repeated wave action is often terminated by a great burst through of polar air into the subtropical zone. Behind each wave the front penetrates further south than in the previous case, and behind the last the break-through is complete.

These depressions strung out along a single front were called by J. Bjerknes "cyclone families". He showed that the average number of storms in a family was four or five.

Other Types of Middle-latitude Depressions

Though most travelling depressions originate and evolve in the manner just described, there are important exceptions. Many depressions have a frontal structure so complex or so ill-defined as to defy analysis. Others originate independently of fronts, and it is with these examples that we are concerned here.

"Thermal" Depressions. Areas of prolonged and intense solar heating are normally covered by shallow, complex low-pressure systems. The great heating leads to a general expansion of the air columns and an outward flow at high levels. This causes a fall of pressure at ground level, though the resulting low pressure is rarely very pronounced.

Thermal lows vary in scale from the tiny depressions which sometimes form in quiet, hot weather over East Anglia to the vast monsoonal lows of Asia (about which more will be said later). The latter, however, have certain properties which differentiate them from the more local variety. Thermal lows do not usually move or persist if cooler weather supervenes, but they occasionally develop into travelling disturbances by absorbing fronts. They do not cause widespread bad weather, though if the air is damp there are usually some thunderstorms by day. The lows are deepest in the afternoon, and may fill up altogether at night.

"Polar Air" Depressions. Occasionally active depressions develop wholly within unstable polar air—usually polar maritime. Such developments are especially common in northerly spells in the seas round Britain. They are usually small, short-lived and highly erratic in motion. Sometimes, however, they become large and deep; a memorable snowstorm occurred in Britain on February, 23rd–26th, 1933, as one such storm moved slowly southwards across Ireland. More usually they lead to showery, unsettled, cool weather, without marked rain-areas. A place where they are especially liable to form is south of the centre of an old occluded depression. They are then apt to move very rapidly round the old centre. Some of Britain's worst gales have originated in this manner.

Lee-depressions. A curious group of depressions is associated with high mountain ranges. During the passage of a cold front across such a range as the Appalachians, a wave is often induced, and a frontal depression forms on the lee-side of the range. According to Petterssen (1940) this results from the steepening of the frontal surface as it climbs the hills. There is, however, another type of lee-depression (i.e. a depression forming on the lee-side of a hill range during flow across it). This seems to be a special type of dynamically formed eddy, and fronts are not necessary for its formation. Among those areas where such depressions form, we can distinguish as especially important the following:

(i) The Po Valley and Ligurian Sea areas in the lee of the Alps. These "Genoa" depressions are vitally important to the climate of the Mediterranean Basin. They form when a cold north-westerly current approaches the Alps across France: they accompany the Mistral in the Rhone Valley and Provence.

(ii) The eastern foothills of the Rockies, especially in Colorado and Alberta. The depressions which originate here give North America many of its worst storms.

(iii) The lee of the mountains of South Island, New Zealand.

Other areas where effects of this kind operate include the north Sahara (behind the Atlas) and the Gulf of Tong-king during the S.W. monsoon. All these cases will be discussed in the regional section.

Areas and Seasons of Occurrences

The importance of the travelling depression in dynamic climatology can hardly be exaggerated. Outside the Tropics it controls rainfall distribution and even prevailing weather over a large part of the globe. Only deep in the continental interiors, where there is too little moisture for frontal precipitation, is the frontal storm of little account, and even here it may bring sand or dust storms.

Depressions of the frontal type usually originate in certain broad areas and die out in others, and though the scatter of cyclone tracks is extremely wide, there is a broad regularity underlying their habits which is of the greatest importance to the climatologist. Certain parts of the earth's surface are especially favourable for frontogenesis, and it is in these parts that we normally find the maximum frequency of depression-formation. Other areas are more usually frontolytic, and here the formation of a depression is quite a rarity. Finally, certain other areas seem especially attractive to fully developed depressions; these areas, of which Iceland and the Aleutian Islands are good examples, have been picturesquely dubbed "graveyards" by climatologists, for it is to these areas that great numbers of active storms retire when about to die out.

The world distribution of depressions and their seasonal variation are dealt with at length in the section on regional climatology.

CHAPTER VII

ANTICYCLONES

THE anticyclone has been aptly termed the ugly duckling of modern meteorological research. The great interest that was aroused by the Norwegian methods of chart analysis has kept the attention of meteorologists firmly fixed on the cyclone, to which the new techniques are addressed. We know today little more about the structure and origin of the anticyclone than did the forecasters of the first world war. Indeed, in many ways our uncertainty about them has increased, for many of the older views have been discredited without better views to replace them.

Dynamically, the anticyclone resembles the cyclone in more ways than one. It is a cyclic type of circulation which is capable of movement, growth and decay. Its outline, like that of old frontal depressions, is often roughly circular, though it is less consistent in this respect than the cyclone. In other ways, it is the antithesis of the cyclone; aside from the fact that it is an area of high rather than low pressure, it is associated with fair weather, descending and diverging air and with a uniformity of airmass properties over wide areas. Lastly, it has certain properties which totally differentiate it from the cyclone as an element in the general atmospheric circulation.

Though there are at least two types of anticyclones, all such pressure systems have certain properties in common. All are fairly large: there is no real equivalent among high-pressure systems for the "small" depression, which can cause widespread bad weather despite its small size. Some anticyclones, indeed, may achieve extraordinary size. If we take the 1015 mb. as marking the outer edge of the anticyclonic circulation, the winter anticyclones of Siberia often extend from the Black Sea to the Bering Straits, a distance of over 6,000 miles, and they are almost equally extensive along the direction normal to this long axis. Even the travelling anticyclones of middle latitude often exceed 2,000 miles in long-axial length. Big or relatively small, all these systems are characterized by a central region of light winds (because of slack pressure gradients) and of subsidence (that is, a general sinking towards the ground of the air-mass involved). This subsidence is an essential dynamic property of anticyclonic circulation; it can readily be shown that the creation of such circulation requires divergence in the layers affected, and divergence in turn requires subsidence. The significance of the predominating downward movement of the air is that there is little prospect of cloud- or rain-formation, which require pronounced upward movement. Anticyclones are hence regions of quiet, fair weather, though the latter is sometimes spoiled by fog or strato-cumulus cloud at low-levels.

It was once believed that the high-pressure within anticyclones was due to the coldness of the air above them; since cold air is denser at the same pressure than warm, it was reasonable to suppose that the excess of mass was due to the presence of abnormally cold air aloft. In 1909, however, S. Hanzlik (1909) was able to prove that in many cases the troposphere is *warmer* in anticyclones than in depressions. Hanzlik distinguished two types of anticyclones on the basis of their structure within the troposphere:

(1) *Cold anticyclones*, with abnormally cold air in the lower troposphere. Such anticyclones, he claimed, were quick-moving, short-lived and shallow.
(2) *Warm anticyclones*, where the temperature throughout the troposphere was abnormally high: he regarded these systems as stable and slow moving.

The essential truth of Hanzlik's findings has since been made abundantly clear, and the two classes of anticyclone are now generally recognized. A short account follows of the properties of the two types.

Cold Anticyclones

Cold anticyclones consist of shallow, dome-shaped pools of cold air in the lower troposphere within which there is anticyclonic rotation. The high-pressure at the surface is due to the heavy weight of this dense cold air. Since pressure decreases with height more rapidly in cold air than in warm, the intensity of these cold anticyclones always decreases rapidly as one ascends. As low as 5,000 ft. it is common to find the high reduced to a mere ridge extending northwards from the sub-tropical anticyclone, and at 10,000 ft. there is often little trace of it. If the temperature distribution within the cold air is fairly uniform, there may even be a closed centre of low-pressure above the surface anticyclone. B. Haurwitz and J. R. H. Noble (1938) have described an intense anticyclone centred over the U.S.-Canadian border in the prairies in which anti-clockwise circulation vanished at 5,000 ft.; at 14,000 ft. a deep low-pressure centre lay almost above the surface anticyclone. It is more usual, however, to find strong westerlies above cold anticyclones. Such is the case with the Siberian anticyclones of winter, which give way at 5,000–7,500 ft. to strong and undisturbed westerlies.

Cold anticyclones form chiefly in high latitude zones within continental Polar or Arctic air. Precisely what causes the anti-cyclogenesis is not clear, but at intervals throughout the year such anticyclones rapidly emerge from the polar or sub-polar zone and travel towards some point between south and east. A common occurrence is for them to move south-eastwards in the rear of the last and most intense member of a depression family (see p. 81). Though they can travel along many different paths, two are of especial importance in the Northern hemisphere. One is across North China and Japan, where the highs come from the cold heart of Siberia. The other is across Canada and the United States, which are affected by anticyclones originating over north-western Canada and the Arctic Ocean. Some of these cold anticyclones, however, appear to travel all the way from Siberia, often across Alaska and the Yukon. In both North America and Asia the cold moving highs normally move south-eastwards into sub-tropical latitudes over the ocean to the east, where they coalesce with or replace the warm anticyclones normally found near Hawaii and the Azores. In so doing, they are transformed into the warm type of anti-cyclone. The cold anticyclones which affect Europe form chiefly in the rear of intense cyclones moving eastwards into Northern Europe. They are developed chiefly in maritime polar or arctic air, though a few cases form in polar continental air from Siberia. The European systems are short-lived and relatively infrequent.

Within the moving anticyclone there is usually marked subsidence, the air being slowly warmed and dried. A pronounced inversion separates the cold air from the warmer westerly circulation above. The air nearest the ground may be fairly unstable, though there is usually too little moisture for cloud to form unless the anticyclone passes out over the sea or a large lake. Cold anticyclones normally bring very fine, dry cool weather. Only over the sea or along onshore coasts are clouds and precipitation at all common. The most beautiful weather the writer has experienced occurs in the rear of retreating cold anticyclones in North America; the air, though cool, is sparklingly clear and dry; and visibility is unlimited. Even in as smoky a city as Montreal it is often possible to see the summits of the Green Mountains of Vermont up to 100 miles away from the summit of Mount Royal. In winter, however, the polar continental air involved is too cold for comfort; sub-zero temperatures are brought to a wide area of eastern North America.

Over both eastern Asia and eastern North America, weather is controlled for the greater part of the year by a constant succession of moving frontal depressions and cold anticyclones. The slow-moving warm anticyclone, which may dominate Great Britain for a fortnight or more, is very uncommon in these regions.

Warm Anticyclones

The common type in western Europe is the slow-moving warm anticyclone, in which the air is abnormally warm throughout the troposphere. In a series of papers published during and after the first world war, W. H. Dines (1914) prepared averages of temperature at various levels in depressions and anticyclones over Britain. Since most of the anticyclones were warm in type, his results may be taken as defining typical conditions for these systems. He showed that temperatures in warm anticyclones average 18°F. or more higher than in depressions over the British Isles from 3 km. up to 8 km. In the stratosphere, however, the atmosphere is abnormally cold over the anticyclones, and it is presumably the weight of this cold air that gives the high pressure at ground-level. Warm anticyclones, then, extend up to great heights, quite often into the lower stratosphere. It is quite common to see cirrus clouds moving from the east on the southern side of anticyclones over Britain. The course of the cold air in the stratosphere is not known, though it has been suggested that it may come from the equatorial belt, where the stratosphere is very cold. The plain facts are, however, that we do not yet know why a warm anticyclone forms: because dynamic factors are involved, such highs have been called "dynamic anticyclones", but this term is undesirable since it implies that dynamic factors play no part in the life-history of cold anticyclones, which is untrue. In any event, the slow-moving, long-lived warm anticyclone remains in many ways a meteorological enigma.

A sounding through a warm anticyclone usually shows two distinct layers:

(i) Near the ground there is a shallow pool of relatively cool air, above which there is a pronounced inversion or isothermal layer. In winter this lower layer consists of inert, stagnant air which has been cooled by radiation. In Britain warm anticyclones often form in fairly moist air, and in winter widespread fog occurs as a result of this cooling. The slack winds typical of anticyclones encourage the persistence of the fog. There are also layers of stratus or strato-cumulus cloud in many cases, chiefly at the base of the inversion, which is found as a rule between 1,500 and 5,000 ft. The strato-cumulus typical of the east-wind area south of the centre comes in across England from the North Sea; its deep penetration is made possible by the absence of any real hill barrier to break up the cloud. Warm anticyclones hence tend to bring foggy or gloomy, overcast weather in winter. In summer the greater solar heating gives clear, warm weather, but the lowest layers are still fairly stable, and cumulus clouds are stunted and flattened at quite low levels. Over the sea most warm anticyclones are cloudy at all seasons; the east-wind belt south of the sub-tropical highs is, however, an exception to this rule.

(ii) Above the inversion mentioned above, the air is at all times of year warm and fairly dry. The warmth results from the marked settling which affects the air throughout. The adiabatic warming effectively prevents cloud formation, and since the subsidence extends to great heights, the air is usually completely clear right up to the stratosphere. Precipitation is very rare in warm anticyclones; the most that may fall is slight drizzle from stratiform cloud below the inversion.

Areas of Occurrence of Warm Anticyclones

Warm anticyclones are definitely more common in some areas than in others. Any sort of anticyclonic circulation is impossible (except for very brief periods of an hour or two) near the equator, so that anticyclones are essentially features of the circulation over middle and high latitudes.

Their main area of occurrence is the sub-tropical belt, especially over the oceans. The exact significance of this marked concentration is not fully understood, but it is undoubtedly an essential feature of the general world circulation, under which heading it is discussed in this book. Slow-moving warm highs often emerge from the sub-tropical belt and travel across western Europe or North America. In the southern hemisphere the entire sub-tropical belt of high-pressure consists of moving anticyclones which do not, however, move far outside the sub-tropics.

There are times when warm anticyclogenesis occurs over western Europe, especially over the British Isles; this normally coincides with a decline in intensity of the normal sub-tropical high near the Azores, where it is common to find a curious stationary depression often of considerable intensity. It is clear that the European anticyclone then plays the part of the Azores high in the general world circulation. This condition is to be regarded as an important variant of the normal circulation pattern, for which there is no precise equivalent anywhere else in the world.

The Decay of Anticyclones

Both warm and cold anticyclones may give way quite suddenly, or else may persist for some days with declining intensity. Cold highs usually begin to decrease in intensity as soon as they reach warmer latitudes, and in Europe they may vanish within 24 to 36 hours of crossing the coast. Those which cross North America and Asia normally decrease in intensity after crossing the east coast. Warm highs, however, are quite erratic in their late stages. In western Europe an anticyclone often persists for days with little change of shape, area or central change. It may then begin a slow *in situ* decline, which continues until no trace of it remains. Others begin to move as they decay, usually in an easterly direction. Another common fate is for an old centre to decline in favour of a new one which is forming not far away within cooler air.

The general weather conditions within a declining anticyclone differ very little from those of systems in the active, growing stage. In England some of the finest summer weather comes at these times.

We may note in closing that the distinction drawn here between the cold and warm centred classes of anticyclone is not so marked as to prevent cold highs from warming up until they are warm-centred. This is what happens to most cold anticyclones which cross North America and the Far East. It is not clear what principle is involved in this transformation.

CHAPTER VIII

THE GENERAL CIRCULATION OF THE ATMOSPHERE: CLIMATIC ZONES

Meaning of the General Circulation

For many years geographers have talked about "prevailing winds". As soon as sailors began to navigate the globe, it was realized that many areas experienced winds which persistently blew from one single point of the compass. One might say, with some reservations, that the general circulation is simply the overall story of these prevailing winds, not only at ground level, but in the upper air. We notice, however, that when meteorologists discuss the forces that drive this huge machinery of persistent winds, they talk in terms of the heat-balance, of friction, of the rotation of the earth, and of other subjects which indicate that they are concerned with more far-reaching things than prevailing winds. Perhaps it is best said that the general circulation is a broad, generalized, bird's-eye view of the movement and workings of the atmosphere, which at close quarters is a bewilderingly complex spectacle.

The bird's-eye view at once shows us an important fact. If we take daily weather charts for a prolonged period, we soon learn that the patterns of flow shown upon them fall into two categories:

(1) short-lived, rapidly moving anticyclones and depressions of the sort discussed in Chapters VI and VII. These eddies begin, come to maturity, and then completely vanish from the chart in periods ranging from a day to a month. They are called *secondary circulation features*; and

(2) persistent, long-lived and slow-moving eddies, also of anticyclonic or cyclonic type, but which tend to remain almost stationary for long periods. These are the *primary circulation features*, and it is with these that we are mainly concerned at present.

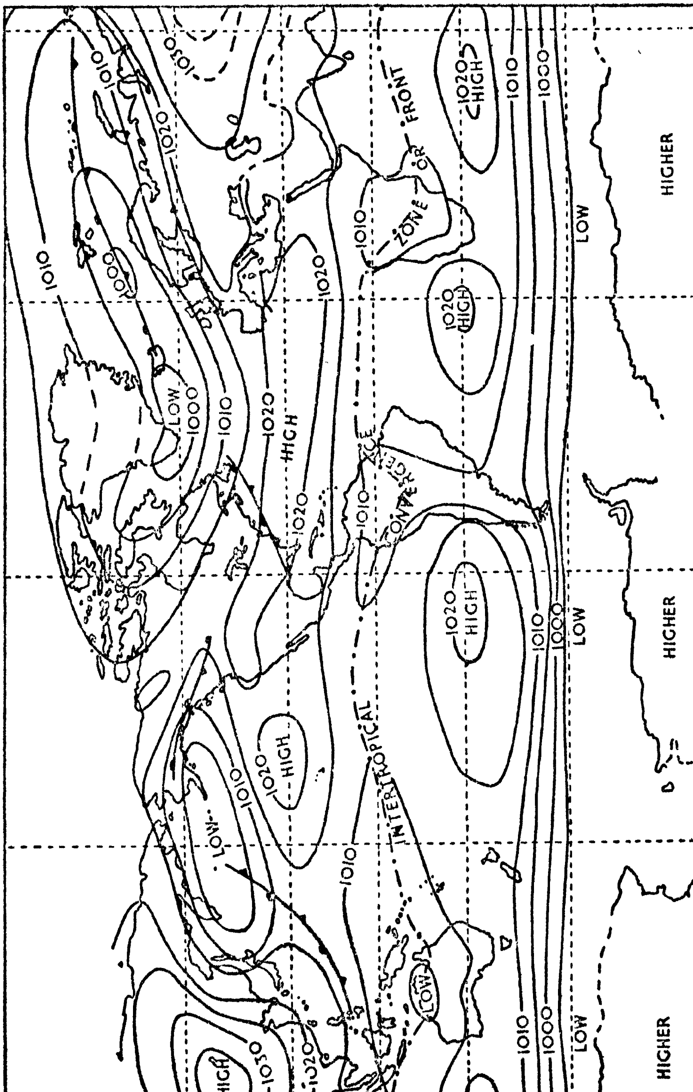


Fig. 8. Distribution of mean pressure at sea-level, January. The heavy alternately pecked and circled lines show zones of frequent frontal activity (following Petterssen).

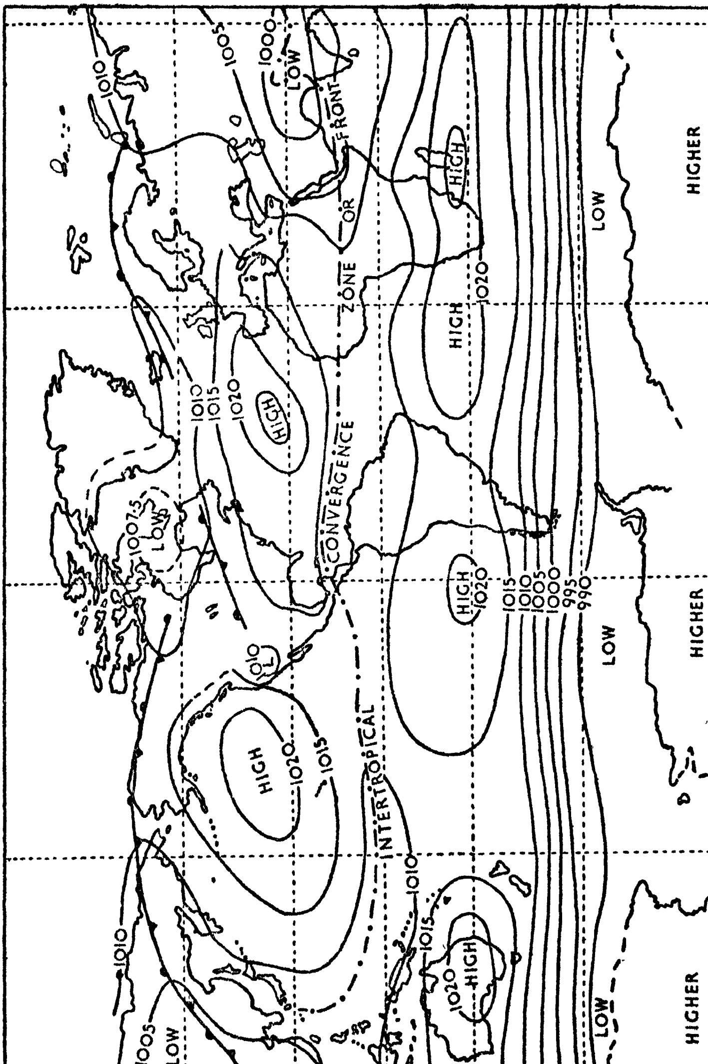


Fig. 9. Distribution of mean pressure at sea-level, July. Frontal zones following Petterssen.

A still more prolonged study might convince us that this seemingly water-tight division is apt to spring leaks, but it will do very well for our present purpose. It is obvious that we must concentrate on describing and accounting for the primary circulation features, which is as far as most theories of the general circulation go; but our picture must leave room for the secondary circulations as well. In very recent years the work of C.-G. Rossby and his collaborators at the University of Chicago has shown that such a perspective view can be derived from existing evidence, though many details remain to be cleared up.

We may notice that the general circulation is in a real sense the factor which determines the climates of the world. Because the circulation tends to arrange itself in latitudinal zones, world climates also occur in zones, a fact recognized as long ago as the days of classical Greece. Where the characteristic zonal character of the circulation breaks down, as in the monsoon areas, there are marked abnormalities of climate.

The Primary Circulation Features

Figs. 8–9 show average sea-level pressure over the earth in January and July. The southern hemisphere is discussed first because it contains less of the abnormalities just mentioned. It is clear that both maps contain certain features in common:

(1) In both months there is a continuous belt of relatively low and very uniform pressure around the Equatorial belt. In January this belt lies rather north of the Equator, but in July it lies either dead on or a little to the south. This low-pressure belt is the belt of calms and hot weather known at sea as the Doldrums. The whole belt can be referred to as the equatorial low-pressure belt.

(2) In sub-tropical latitudes (roughly along 30°S. latitude) there is a broad, continuous belt of high, uniform pressure both in January and July, though in January it is broken up a little by small low-pressure areas over Australia and South Africa. This broad belt is referred to as the sub-tropical high-pressure belt, which is a permanent feature of the general circulation in both hemispheres.

(3) South of the sub-tropical high-pressure belt pressure falls steadily southward to a minimum somewhere over Antarctica. It is not clear whether the fall of pressure continues all the way to the Pole, or whether, as in the northern hemisphere, lowest pressure occurs in about latitude 60°N. In either case, it is plain that we can define a permanent sub-polar low-pressure zone.

If we now examine the northern hemisphere, we see a more complex pattern. On close examination, we find that we can still distinguish these primary pressure zones:

(1) The equatorial low-pressure belt is common to both hemispheres: we note, however, that in July it lies far north of the equator over Africa and Asia.
(2) The sub-tropical high-pressure belt is strikingly different in this hemisphere. In January it occurs over the oceans in about latitudes 25–35°N., but it is linked to much more intense high-pressure systems in higher latitudes over the continents. These latter systems are fundamentally different in character, and it is hence best to regard the sub-tropical high-pressure belt as being confined to the oceans. This is much more strikingly true of July, when large and intense high-pressure areas cover much of the North Atlantic and North Pacific, the high-pressure centres lying approximately in latitudes 30–35°N. On the whole, then, it can be said that this belt in the northern hemisphere is chiefly oceanic, whereas in the southern hemisphere it encircles the globe.
(3) The sub-arctic low-pressure belt is plainly visible on the January chart. It consists of two great low-pressures over Iceland and the Aleutian Islands, covering the northernmost Atlantic and Pacific respectively. These two centres are separated by an area of relatively high pressure over the pole. In summer the low-pressures are much less intense. The Aleutian centre has disappeared, and that near Iceland is quite feeble. Pressure remains relatively high over the poles. In neither season does

the sub-arctic low-pressure extend round the globe as it does in the other hemisphere.

(4) The breaks in the continuity of the sub-tropical high and sub-polar low-pressure belts are caused by new and alien elements for which there is no southern equivalent. In winter there are the great high-pressure belts of North America and Asia, and in summer there is the equally large low-pressure belt centred over southern Asia. The nearest equivalent in the southern hemisphere is the low pressure over Australia and South Africa in January. In each case, in both hemispheres, these abnormalities occur over land masses.

Factors Governing the Circulation

By applying the rules given for the relation between wind and pressure gradient (see p. 49) to their charts of mean pressure, we can get the mean wind strength and direction—i.e. the mean circulation—all over the world. These great wind belts are common knowledge, but it will be useful to note them here:

(1) On the equatorial side of the sub-tropical high-pressure belts are belts of easterly winds with a component towards the equator. These are the *trade winds*—north-easterly in the northern hemisphere, south-easterly in the southern.

(2) On the poleward side of the sub-tropical high-pressure, and extending to 60° latitude are the prevailing westerlies familiar to us in middle latitudes. Notorious for their stormy inconstancy and proneness to disturbances, these westerlies will be called here the *disturbed westerlies*.

(3) Between the centres of lowest pressure in latitude 60° and the two poles, with their relatively high pressure, there are belts in which easterly winds occur over certain areas. The *polar easterlies* as they are called affect only a small part of the globe.

(4) Corresponding to the seasonal high- and low-pressure systems of continental Asia and Australia there are the

great wind systems known as the "monsoons". The similar systems over North America are less well-defined, and no true monsoons are developed.

We can best understand these various parts of the circulation if we examine two of the causes underlying them, viz. the energy budget of the atmosphere, which we have already examined in Chapter I, and the effect of the earth's rotation. These factors act together to create the general circulation. The energy budget enters the story because a large amount of energy is necessary to drive so tremendous a circulation. The energy is made available by the temperature contrasts existing between the equatorial belt and the polar caps. We saw in Chapter I that these temperature contrasts arise because of the widely different annual receipts of radiant energy between high and low latitudes. The circulation of the atmosphere functions in such a way as to transport heat from low latitudes to the cold polar caps, and so prevents the disparity in warmth from becoming even greater than it already is. The need, then, is for a means of bringing about an exchange of energy along the meridians of longitude, i.e. a *meridional* circulation of the atmosphere.

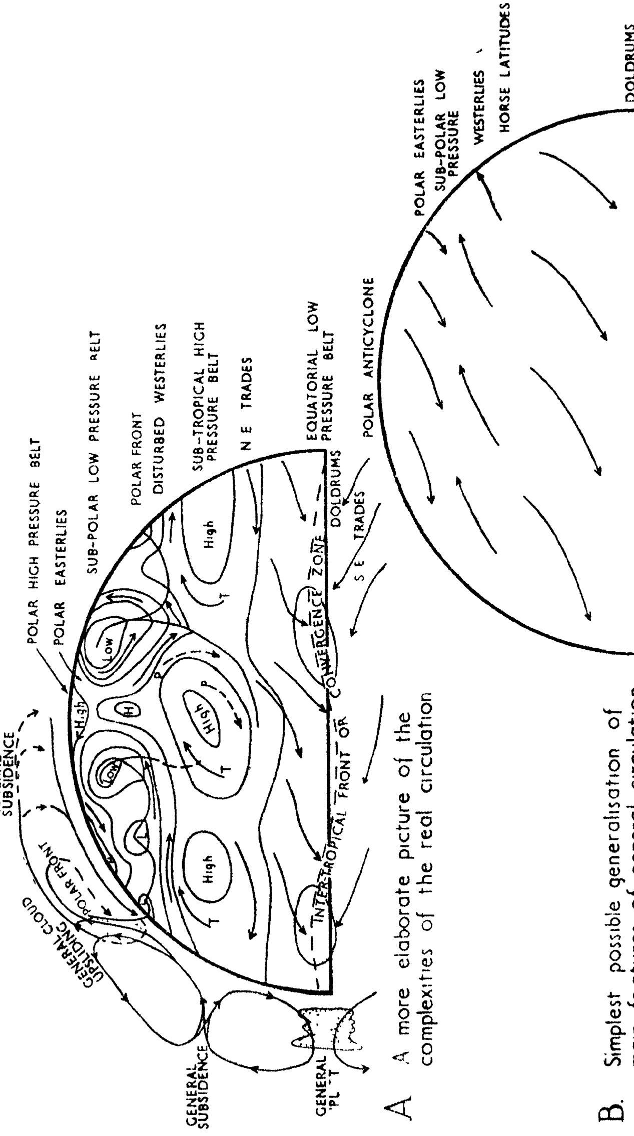
Observation shows us, however, that the prevailing winds of the earth are predominantly *zonal* rather than meridional: that is, they tend to follow parallels of latitude rather than the meridians. Only the monsoons of Asia have a mainly meridional motion. Elsewhere in the atmosphere southerly or northerly winds occur chiefly for limited periods and over limited areas. In the upper atmosphere this becomes even more completely true, for the middle and upper troposphere are occupied by vast circum-polar whirls of westerly winds, except for a belt of calms or variable winds in the equatorial belt. The patterns are basically similar in all seasons, though most of the wind belts are stronger in the winter hemisphere than in the summer. It is clear that the circulation is overwhelmingly zonal in character, even though it is driven, as we saw above, by the contrast in temperature along the meridians.

This transformation is brought about by the rotation of the earth, which, as we saw, deflects all moving objects away from the straight path. On a non-rotating earth, the heated air over the equatorial belt would rise, and would then flow northward at upper levels. Over the poles it would subside slowly to the ground, where it would give up much of its heat to space by radiation. The cooled air would then return to the equator as a surface northerly current. Both the upper southerly and the lower northerly currents would, however, be subject to the deflecting force of the earth's rotation; we might therefore expect the upper current to be deflected into a westerly wind and the lower to an easterly. In reality, the transformations are much more complex than this: the upper westerlies are certainly present over most of the earth, but the surface easterlies are largely confined to the trade wind belt, comprising only about a third to a quarter of the earth's surface. There is no space here for a complete discussion of the breakdown of the meridional circulation under the influence of the rotation of the earth; the interested reader is referred to papers by Rossby (1941) and Barlow (1934). It will be enough for us to note that the very necessary meridional exchanges of air and energy are carried out by inconstant northerly and southerly winds both at high and low levels, whereas the dominating circulation is that along the parallels of latitude, viz. the trade winds, the westerlies and the polar easterlies.

Fig. 10 is an attempt based on a similar diagram by Rossby (1941) to sum up the two parts of the general circulation. It shows a cross-section of the atmosphere along a meridian above a map of typical surface circulation over a hemisphere. The cross-section gives an idea of the net meridional circulation over the hemisphere as a whole, whereas the map shows the dominating zonal circulation belt on which the feeble but vital meridional motion is superposed. The following points are illustrated by the diagram:

(1) The trough of low-pressure in the equatorial belt is shown as a zone of general ascent of air. The north-east and south-east trades flow into the trough, and meet along the axis, which is sometimes called the "Inter-tropical Front". At this axis the moist, warm trades air rises steadily into the upper troposphere; the uplift is

A COMPOSITE VIEW OF THE GENERAL CIRCULATION OF THE ATMOSPHERE.



A. A more elaborate picture of the complexities of the real circulation

B. Simplest possible generalisation of main features of general circulation.

Fig. 10

associated with heavy rains and dense cumuliform cloud, which are sketched in the diagram.

(2) The rising air over the equatorial belt diverges north and south. The current moving northward above the trades can usually be detected above the trade wind by pilot-balloon or observation of cirrus clouds. It has a marked westerly component, and so appears as a south-westerly current in this hemisphere. It is called the "counter-trade". Since the counter-trades on either side of the equator are the only currents leaving the equatorial belt, they must carry away all the excess energy accumulated there by solar radiation.[1]

(3) The sub-tropical high-pressure belt consists of a series of anticyclones set in a belt of generally high pressure. The north-east trades and the disturbed westerlies appear on their southern and northern flanks respectively. Within the main high-pressure belt itself, there are areas of southerly winds on the western flanks of the separate "cells", as the anticyclones within the belt are often called. It is along these western flanks that the invasion of tropical air into the disturbed westerly belt occurs. Three such invasions are shown by the arrows marked T. Northerly winds occur on the eastern flank of each cell. Usually these northerlies are warm air which has travelled right round the anticyclone, supplemented by other air which has subsided from above. At times, however, great outbreaks of polar air occur from high latitudes: usually such outbreaks develop in the rear of the last and deepest member of a cyclone family. A typical case is shown in fig. 10 by the arrows marked P. It is chiefly by such outbreaks that the supply of air in and south of the high-pressure belt is maintained. The return high-level current shown in the cross-section is a very slow, irregular flow.

(4) The sub-polar low-pressure and polar high-pressure belts require to be considered together. In very high

[1]Apart from that removed by Austral-Asiatic monsoons.

latitudes the almost continuous radiative cooling leads to general subsidence of the air above, and so to the relatively high pressure of the polar zone. The easterly winds which occur on the flanks of this high-pressure zone often penetrate southward until they come into contact with the warmer westerlies of middle latitudes. Between the two currents is found a belt of frontal activity—the so-called *Polar Front.* This front is usually to be found somewhere between the polar and sub-tropical high-pressure areas, though in position it fluctuates widely from day to day. Frequent depressions form on the polar front and travel eastward, thus in effect creating the sub-polar low-pressure belt, which is nothing more than the path taken by these storms. At intervals, however, a great outbreak of polar air drives the front southwards into the trade wind belt, as at P in fig. 10, and for a time there is no clear-cut frontal belt in the middle latitudes. The supply of air to maintain the subsidence in the polar belt is brought northward at high altitudes. Much of it is air which has risen upwards over the polar-frontal surface as sketched in fig. 10.

The Effect of Land and Sea

The striking difference between the northern and southern hemispheres in circulation arises from differences in the disposition of land and sea. Since the larger part of the earth's crust is ocean-covered, we may look upon water as the normal surface, and consider the complications introduced by the land. These are of two sorts:

(1) *Thermal modifications,* due to the marked seasonal differences of temperature over the continents. The interior of the continents become very hot in summer and very cold in winter; these differences of temperature radically change the pressure distribution over the continent, and so in turn affect the world circulation.

(2) *Dynamical modifications,* due to the presence of mountains.

The thermal modifications are responsible for the great monsoons. They affect only the largest landmasses of the globe. Asia is the continent most affected, with Australia and Africa the other appreciable cases. North America shows monsoonal tendencies, but the scale of development is much lower than in the other cases.

The Asiatic monsoons are seasonal wind systems established each year with great regularity. In winter the frigidly cold interior of Siberia is covered by almost permanent high-pressure, usually of great intensity. Cold northerly and north-easterly winds flow across eastern Asia, being joined over the Bay of Bengal by a subsidiary stream from India. In summer the position is reversed. Pressure is low over the interior, and huge areas are covered by warm and humid air from the equatorial belt. These monsoons, summer and winter, extend from latitude 60° or 65°N. to the equator and so totally break up the "normal" general circulation in the longitudes of Asia.

Far smaller but similar monsoons affect northern Australia and west Africa. The case of North America is, however, different. As in Asia, the continental interior becomes hot in summer and very cold in winter, and there is a visible tendency for low-pressure over the interior in summer and for high-pressure in winter. But these tendencies do not dominate the circulation, and there is no persistent monsoonal wind or seasonal wind change.

There is so little land in the southern hemisphere that thermal modifications of the circulation are negligible except over Australia and in a lesser degree over south-west Africa.

The dynamic modifications are less clearly understood. It is well known that the larger mountain ranges of the world, projecting as they do into the middle or upper troposphere, cause considerable local disturbance of the flow of air. We have examined the case of lee-depressions, which owe their formation to the barrier action of mountain ranges. It seems likely that the greatest of the barrier ranges may seriously affect the general circulation.

Special significance attaches to the north-to-south ranges and to the great central ranges of Asia. The former include the western cordillera of North America and the Andes. Both these ranges are high enough to project well into the circum-polar westerly belts of the two hemispheres. Attempts have been made to show that a sort of stationary wave is likely to be found downstream when strong winds are forced to cross high ranges, and it is quite likely that the Rockies and other North American ranges, the Andes and the ranges and plateaus of central Asia create semi-permanent disturbances of the high-level westerlies. This would in turn affect surface winds and pressures.

In recent years C.-G. Rossby and his Chicago group of meteorologists (1941) have attached great significance to the strength of the circum-polar westerlies, which appear to fluctuate remarkably in speed. They have shown that these fluctuations significantly affect the behaviour of the lower atmosphere and must also affect the stationary waves referred to above; the waves vary in their form with the strength of the circulation.

Conclusion

We see, then, that the circulation of the atmosphere divides itself up naturally into great zonal belts, which are the background of world climates. In the succeeding chapters we shall examine briefly the climates of the world, in each case following a logical scheme whose purpose will be to relate the observed climate to the dynamic background of the circulation.

CHAPTER IX

THE INTERTROPICAL BELT: THE OCEANS

Limits and General Properties

The adjective "tropical" has become loose in meaning. It now stands for any warm place irrespective of latitude, or even warm weather. As a collective adjective to describe the latitudes between $23\frac{1}{2}$°N. and $23\frac{1}{2}$°S., the tropics, the writer uses "intertropical", and as a more restricted term to cover a narrow belt a few degrees on either side of the equator, he uses "equatorial". The logical limits for the intertropical climate are, of course, the axes of the two sub-tropical high-pressure belts, which divide the warm-climate trade-winds from the cool-climate disturbed westerlies. Even the word "intertropical" is thus imperfect as a means of describing the warm climates of the world, for the high-pressure axes habitually lie just poleward from the tropics. But too great an adherence to latitudinal nomenclature is in any case unwise, for there are some truly intertropical areas whose winter is quite cool, and several extra-tropical areas with high temperatures at all seasons.

This belt between the sub-tropical highs is, nevertheless, to a high degree of approximation the hot climate belt. The whole of it receives annually more solar energy than it can return to space, and it is therefore under the necessity of exporting its surplus to cooler latitudes. It is the belt where day to day weather is most influenced by the sun and least influenced by major dynamical upsets of the circulation of the atmosphere. It would be a mistake, however, to assume that the sun is the only factor to be considered; too much emphasis has been laid in the past upon the diurnal and seasonal rhythm of the sun's influence, and too little upon such matters as the type of circulation and the general dynamic background of the climate.

The intertropical belt is overwhelmingly oceanic: the continents of Africa and South America lie across it, as do

India and northern Australasia, the East and West Indies. But these are small by comparison with the huge expanses of the Pacific, the Atlantic and the Indian Ocean. Over the Atlantic and Pacific the classical sequence of weather-zones reaches its maximum expression: along the sub-tropical belts, the anticyclones divide the westerlies from the trade winds; the latter cover the greater part of the intertropical waters, meeting not far from the equator along the narrow doldrum belt, with its calms and heavy storms. Thus the sequence of becalmed horse latitudes, steady trade winds and equatorial calms is maintained through most of the year, with only small displacements from season to season, and with no disturbances comparable with those affecting middle latitudes. The Indian Ocean is affected by the Asiatic monsoons, and differs from its bigger neighbours.

In this book the sub-tropical belt of high-pressure itself is given separate treatment, as are the Asiatic and Australian monsoons. The account that follows refers chiefly to the trade wind and equatorial belts outside the monsoonal belt.

Weather in Low Latitudes. Many things differentiate the intertropical belt from middle latitudes besides temperature. The geostrophic and gradient-wind relationships, which enable the middle-latitude analyst to determine the circulation without directly observing it, break down near the equator because of the decrease in the deflecting force of the earth's rotation, on which they depend. At the equator itself there are few non-periodic variations in pressure, and the pressure field is useless in determining the flow of the air. The airmass and frontal concepts also diminish in value. The uniformly hot surface inhibits airmass differentiation; frontogenetic wind fields are rarely seen, and sloping discontinuity surfaces are rarely long-lived. Plainly the analytical methods of the meteorologist will find little application in such conditions.

In the war years, the pressing needs of the services for accurate forecasts in this zone led to considerable investigation of the climates. The Australian, New Zealand, American and British meteorological services were all actively engaged in forecast and research work within the tropics. Particular interest attaches to the work of the Institute of Tropical Meteorology at Rio Piedras, Puerto Rico, maintained by the University of Chicago and Puerto Rico, and deriving much of its methods from C. E. Palmer of the New Zealand Meteorological Service. Members of the Institute have brought forward a large volume of new fact and theory which materially affects our views on intertropical conditions, even though some of these results may prove unsound. Much of what follows is based upon the published work of this body.

One general result at least has emerged from this concerted study: the simple picture of straightforward thermal convection in the doldrum trough has had to be abandoned. The real structure of intertropical weather is still to a large extent hidden. The simplicity of earlier views rested chiefly upon the absence of widespread observations, and with the growth of observations there has grown up a realization that the intertropical climates are complex and varied in their mechanism.

The Trade Winds. The trade winds have the deserved reputation of being the most reliable, constant winds of the globe. The high degree of permanence of both the sub-tropical high-pressure and of the equatorial trough ensures that there is always a sea-level pressure gradient for easterly winds over a large part of the low latitude belt. The trades are all-year-round winds, though they undergo significant seasonal changes in strength in some areas. Because of the disrupting influence of continents on the sub-tropical high-pressure belt, the trade winds occur chiefly over the oceans, achieving their highest development on the equatorward flank of the Azores and Hawaiian anticyclones in the northern hemisphere and in the South Atlantic in the southern hemisphere.

The outstanding characteristic of the trade wind belts in both hemispheres is consistently warm, sunny weather, which only occasionally breaks down. The surface airmass is maritime tropical, for which the trade wind belt is the chief source region; this air is moist at the surface, but there is now much evidence for believing that it is almost always dry above. The depth of the moist layer varies somewhat: at the eastern end of each ocean it is usually less than 4,000 ft. deep, but it may attain

8,000 ft. in the western parts. When disturbances occur the layer gets much deeper. There is also an inversion at fairly low levels, especially on the eastern sides of the oceans. In general the inversion is lowest in the central regions of the sub-tropical high-pressure belt and highest and least intense where the trades enter the doldrum belt.

The trade winds consist, then, of warm, superficially moist, fairly stable air. From such air, fair, sunny weather with scattered flattened cumulus cloud and brilliant visibility is the general rule in intertropical latitudes. This is a fair description of prevailing weather in a large part of both trade-wind belts. Near the doldrum trough, however, the stability is markedly lower, and larger cumuliform cloud is common, with thunder-showers by no means unusual. This is also true of the western side of the great oceans, where the depth of the moist layer is greater, and where the air is often potentially unstable.

Disturbances affecting the trade-wind belts are discussed later.

The Doldrums and the Equatorial Trough. The immediate vicinity of the equator is a region of very uniform pressure, with the general level of pressure hovering around 1010 mb., which is well below the sub-tropical values. There is hence a belt of relatively low pressure along or near the equator, which is normally the area of calms or light winds called the "Doldrums". In detail the axis of lowest pressure does not quite coincide with the geographical equator, lying instead over or close to the belt of warmest sea-surface, the so-called "thermal equator". Figs. 8–9 show the average position for the trough in January and July. It is seen that it lies wholly within the northern hemisphere in July, but shifts south of the equator over Africa, the Indian Ocean and the Australasian archipelagos in January. This migration of the equatorial trough "follows the sun", though in an irregular manner.

The equatorial trough, or doldrum belt, is an area prone to frequent squalls and heavy rainstorms, which are generally understood to result from the general uplift of air which occurs in many parts of the belt. This upward movement was formerly explained as being due to heating of the inflowing trades by the warm sea surface. That inflow, however, of itself requires that uplift shall occur; the resultant convergence must be compensated by vertical motion near the axis of convergence. The warmth of the seas ensures high vapour contents in the air involved in the uplift, so that heavy rains are to be expected.

The typical "squalls" of the doldrum belt, however formed, have many curious features. They occur chiefly at night, for reasons not yet adequately explained. There is next to no variation in either air or sea temperature during the day, so that there is no obvious reason why the storms should favour the hours of darkness. A nocturnal maximum of rainfall is experienced in many sea and coastal areas outside the intertropical belt, especially in tropical air. The usual explanation involves a radiative cooling of the upper troposphere by night, a process the writer considers inadequate in itself to account for the magnitude of the nocturnal maxima.

It has sometimes been supposed that the trade winds meet not in a no-man's-land of light winds, but along a fairly definite "front". It was shown statistically that over the Pacific the trade winds from the summer hemisphere were appreciably warmer than those coming from the winter hemisphere, so that there might reasonably be a temperature contrast along their meeting-line. This fundamental atmospheric boundary has become known as the *intertropical front* (abbreviated ITF) or sometimes as the equatorial front. Plainly the intertropical front must generally be the axis of the doldrum belt, at least over the oceans. In monsoonal areas like West Africa, Asia and Australia, it is sometimes displaced far from the equator, in which cases there is no true doldrum belt over the oceans.

It must not be imagined that the intertropical front resembles an extra-tropical front in properties. There is very little difference in density between the two airmasses involved. Furthermore, the equilibrium of a sloping frontal surface depends on the sine of the latitude, which is zero at the equator. There are thus both thermal and dynamical arguments against supposing that the "front" has normal properties. On the other hand, it is clearly the axis of a pronounced field of convergence. The use of the word "front" is hence something of a misnomer, and it has been suggested that the term "*intertropical convergence zone*" should replace the older form. The latter term is not adopted in this book.

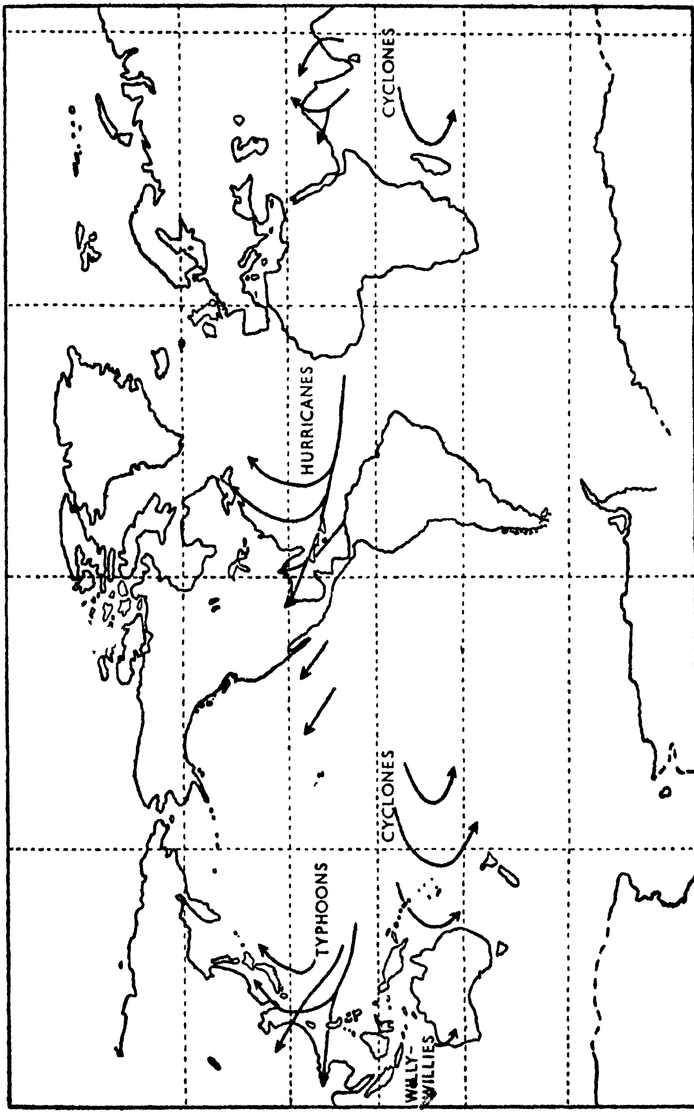


Fig. 11. Areas of occurrence and typical tracks for tropical revolving storms (in part following Tannehill).

Weather at the front over the oceans is highly variable, ranging from clear weather to heavy continuous rain, with dense multi-layer cloud to great heights. The "normal type", if it is possible to describe one, is for frequent squalls of the type described above for the doldrum zone, which is for all practical purposes the same thing.

Disturbances within the Intertropical Belt: Revolving Storms. Though violent storms occur quite frequently in the doldrum belt, they are usually of a localized character. The huge travelling disturbances and anticyclones of middle and high latitudes have no counterpart within the tropics. Certain classes of disturbances do occur, however, and they must be treated here.

Greatest by far in significance are the tropical revolving storms, which are probably the most intense storms affecting the surface of the earth, though they are quite small. They occur in tropical rather than in equatorial latitudes, chiefly over the western sides of the great oceans. In each area affected by the storms there is a special local name for them. In the Atlantic they are "hurricanes"; in the western North Pacific, "typhoons"; in the south Pacific, "hurricanes", "cyclones" or "willy-willies" (the latter north and west of Australia); in the Indian Ocean and the Bay of Bengal, "cyclones". Fig. 11 gives the chief areas of formation with predominating tracks. It should be noted that occasional storms may affect areas far removed from the normal track. During 1944 an aberrant Atlantic hurricane moved east across the Atlantic instead of north-west, crossing the Azores and finally dying out near Portugal.

The typical storm is small but deep, with very low central pressure. An intense cyclonic circulation is set up about the centre, winds attaining 70–80 m.p.h. quite frequently, and over 100 m.p.h. on occasion. Near the centre is an area of calm, the "eye" of the storm. Within the area of violent circulation there is dense nimbo-stratus and alto-stratus cloud, either solid or in close-packed multiple layers, extending to above 20,000 ft. At the outer edge of the storm the cloud thins out to a veil of cirro-stratus and cirrus which may extend some hundreds of miles from the centre. The "eye" of the storm is an area relatively free of cloud: the sky is often clearly discernible. Heavy and continuous rain falls near the central regions (except in the clear eye), with showers or intermittent rain out to the limit of the cyclonic circulation. Thunder may occur, but it is not very common. Areas directly on the track of the storm may very easily get 5 in. of rain in 24 hours, and falls of 10 in. are not uncommon in low latitudes. There is a record of some 46 in. having fallen in one day in the Philippines, and a similar fall was recorded near Tokyo in Japan.

These inconceivable rainfalls are associated with rapid convergence into the centre and consequent large scale-uplift. The uplift appears to be concentrated in a very narrow ring just round the eye, at least at low levels. F. B. Wood and H. Wexler (1945) have described a flight through an Atlantic hurricane on which they recorded marked down-draughts until they were nearing the centre, when they entered violent but not very turbulent rising currents. The central area or "eye" was a neutral area with little cloud. At higher levels the uplift must be distributed over a greater area or else there could not be the broad canopy of dense rain-clouds.

Tropical revolving storms usually move relatively slowly. Originating in low intertropical latitudes, they usually travel westwards or north-westwards.[1] Those which move north-westwards may recurve in latitudes 20°–35°, and travel north-eastwards. Atlantic hurricanes and Pacific typhoons both very often follow this track, and the Australasian willy-willies and Indian Ocean cyclones take the equivalent southern-hemisphere track. As the storms travel poleward, they grow wider but less intense. If they cross a continental coast, they rapidly fill up, but those which do not may develop into normal extra-tropical cyclones by absorbing fronts.

The violent winds and heavy rains of tropical storms may cause untold damage. At sea there is a heavy sea and swell which may endanger shipping. Over land the heavy rains often cause flooding and damage to bridges, embankments and the like, and severe structural damage may be caused by the wind. In 1926 an Atlantic hurricane struck Miami, Florida, effectively

[1]South-westwards in the southern hemisphere.

pricking the "Florida bubble", which had led to fantastic land speculation in what was regarded as a land of perpetual sunshine. In September 1938 an intense hurricane crossed New England and southern Quebec, causing some 600–700 deaths and destroying hundreds of thousands of trees. Perhaps the greatest tragedy of all was the Swatow typhoon of 1927, in which onshore winds ponded up the water in the bay and drowned some 50,000 shore dwellers. Most revolving storms are less intense than this, of course, but even the weak storms are violent by extra-tropical standards.

What mechanism leads to the formation of revolving storms is doubtful. They have a pronounced seasonal variation, being commonest during the late summer or early autumn and least common in late winter and early spring. They are also concentrated chiefly on the western sides of the oceans. Both these facts call for an explanation. The general view is that heavy, massed convection storms, especially if in an area of cyclonic circulation, may provide the necessary energy for their formation. The air involved must be potentially unstable, though the damp surface layer (see p. 106 above) must be deeper than usual. Beyond this point, however, there is little agreement. Some workers have assumed that they form as waves on the intertropical front, and others that they form at "triple points" in the equatorial trough, that is, at points where three unlike airmasses meet. These frontal theories are, however, of uncertain dynamical validity, and it is probable that the true explanation lies elsewhere.

Though the tropical revolving storms are disastrously violent, and are dreaded both by seamen and by landsmen, they are fortunately well scattered. Even in the direct track of the Pacific typhoons, some twenty to thirty of which develop each year, the odds against a destructively violent storm striking a particular point are quite long. Each year the point experiences strong winds and rainfall from typhoons passing fairly close, but it is only in an occasional year that the really destructive part of a storm passes over the station. The reason is that the storms are small by comparison with the area through which they move.

Other Intertropical Disturbances. By comparison with the revolving storms, other types of disturbances are mild, though their cumulative effect on the climate may be greater because of their frequency.

The commonest is the so-called "wave in the easterlies", a disturbance affecting the trade-wind belt. These waves are simply feeble troughs of low pressure which move slowly westwards through the trade-wind belt in both hemispheres. They have been studied extensively in the Caribbean and North Atlantic areas, and also in West Africa and Australasia. There are differences of opinion as to the reality and significance of these disturbances, but the cumulative evidence that they do indeed affect the climate of some trade-wind areas is now overwhelmingly strong.

Fig. 6 shows an idealized easterly wave, after H. Riehl and others of the Institute of Tropical Meteorology. The wave appears as a trough of low-pressure extending northward from the equatorial trough. As it moves westwards there is general divergence ahead of the trough-line and convergence behind. Immediately behind the trough-line the moist surface layer attains great depth, and there are extensive areas of cumuliform cloud with showers, thunderstorms or areas of rain.

Easterly waves develop chiefly in the trade winds of the summer hemisphere. In winter the depth of the trades is less, and waves if they occur at all, are not usually intense in scale.

The westward-moving disturbances of intertropical North Africa (the haboobs of the Sudan and the line-squalls or "tornadoes" of West Africa) are discussed in the next chapter. It is probable that they are related in character to the waves in the easterlies, though they develop in a shallow layer of south-westerlies.

The remaining types of disturbance sometimes identified in the intertropical belt are of smaller significance. The intrusion of the polar front into the trade-wind belt is often associated with disturbed weather, even when the polar front itself has been rendered almost unrecognizable by modification of the polar air behind it. North[1] of the front, however, the north-easterlies are often much stronger than to the south, the line of the front being replaced by a shear zone. The trough of

[1]South in the southern hemisphere.

low pressure which accompanies the front at high levels may sometimes migrate ahead of the surface front: it may be traced within the trade-wind belt as a shallow trough of low-pressure which moves slowly *eastwards*, that is, in the opposite direction to the easterly waves. Yet another minor disturbance is the so-called "induced trough" within the trades, which develops when a young frontal wave forms north of the sub-tropical high-pressure belt. All these disturbances may give rise to showers or thunderstorms within the trade-wind belt, but none is comparable with either revolving storms nor waves in the easterlies in world-wide importance.

Relation between Revolving Storms and other Disturbances. It was pointed out above that certain broad generalizations can be made about the mode of formation of tropical revolving storms. They appear to form in two main positions:

(1) In the equatorial trough. At certain seasons the equatorial trough is displaced far north or south of the equator. At such times there appears to be a marked tendency for cyclonic disturbances to develop, possibly in the form of waves on the diffuse intertropical front or convergence zone. The disturbances usually move slowly westwards. Some of them develop quite suddenly into very intense low-pressure systems, with all the properties of the tropical revolving storm. Most Pacific typhoons and Atlantic hurricanes develop in this way, though it is believed that these do not achieve the intensity associated with the next group.

(2) In waves in the easterlies. G. E. Dunn (1940, 1944) claimed that many easterly waves approaching the Caribbean developed a complete cyclonic circulation, and that a substantial number of these in a very short time became true revolving storms, embedded wholly within trade-wind (mT) air. This view is favoured by the fact that potential instability is much greater in the western than the eastern part of any oceanic trade-wind belt. Hence if an easterly wave moves into this unstable air, the resultant increased convection may very well provide the necessary energy for cyclone formation.

Conclusion

Any account of the dynamic climatology of a region must unduly stress the disturbances at the expense of the uneventful, settled types of weather. Nowhere is this more truly the case than in the intertropical belt. So much has been learned in recent years of the non-periodic disturbances of intertropical weather that there is a serious danger that the essential calmness of the climate will be forgotten. Dramatic and violent as they sometimes are, the cyclones, typhoons, hurricanes, waves and squalls of the intertropics constitute only a small part of the total story of the hot climates. Greater by far in area, and much more prolonged in time, are the expanses of fine, sunny weather, with only a scattering of cumulus to break the monotony of the blue sky. Within the trade winds themselves such weather is dominant; but even in the equatorial trough, which is authentically an area of predominantly stormy weather, there are days when the sky is clear throughout. It must constantly be remembered, then, that the ordinary fashion of the hot maritime climates is for fine weather except in the doldrums. The disturbances of this settled condition probably gain in apparent stature by contrast.

CHAPTER X

INTERTROPICAL AFRICA

THE climates of the great continents which lie within the intertropical belt differ considerably from those of the oceans. This difference springs from the thermal contrast between land and sea. The great heating of the landmasses at the time of high sun, both in its daily and its seasonal oscillation, leads to greater variations in weather from day to night and from season to season. The slow and not very impressive shift of the intertropical front over the oceans is replaced by major north- and south-displacements in each of the continents. The displacement over Asia and Australasia is so great that the entire zonal circulation pattern breaks down, being replaced by the great Australasiatic monsoons discussed in Chapter XII. Over Africa and South America, however, the displacement of the front is much smaller; it is with the African case that the present chapter is concerned.

The migration of the intertropical front "following the sun" leads to the development of distinctive climatic types, known generally from the prevailing vegetation. The *selva* climates, characterized by dense, evergreen tropical forests, are those which lie near the equator and have no dry season. The *savannah* climates, experienced by areas with open bush and tropical grasslands, have summer rains and winter drought. They are found on either side of the selva belt. Finally beyond the savannahs come the true deserts, with little or no rain at any season. The traditional explanation of these climatic belts involved the movement of the equatorial trough, with which all the heavy rainfall was supposed to be associated. As the belt moved far north or south at the solstices, it brought the summer rains to the savannah belt. In its double crossing of the equator, it was believed to bring a double period of heavy rainfall to the selva belt. This view possesses a large element of truth, but contains two major fallacies. One is that all intertropical rainfall is associated with the trough itself, and the other is that the trough's motion is mathematically regular. It is true that the migration of the trough controls rainfall distribution, but it does so over the continent of Africa because of the influence its motion has on the movement and stability of the various differing airstreams.

The General Circulation. The most comprehensive study of the circulation of the atmosphere over intertropical Africa was made in 1932 by C. E. P. Brooks and S. T. A. Mirrlees (1932) and subsequent work has in no way invalidated their results. Figs. 12–13 show the main airstreams and average rainfall for January and July, modified from this paper.

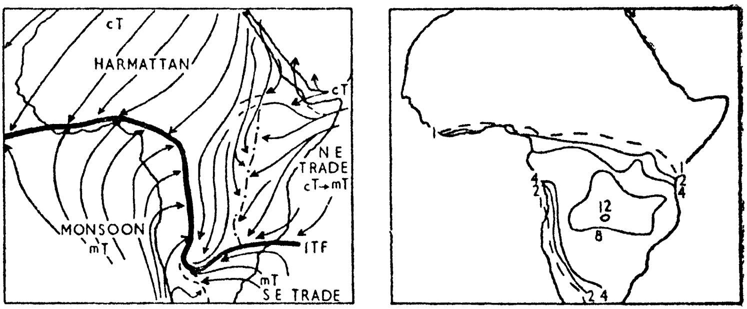


Fig. 12. Circulation patterns and rainfall over Africa, January (following Brooks and Mirrlees).

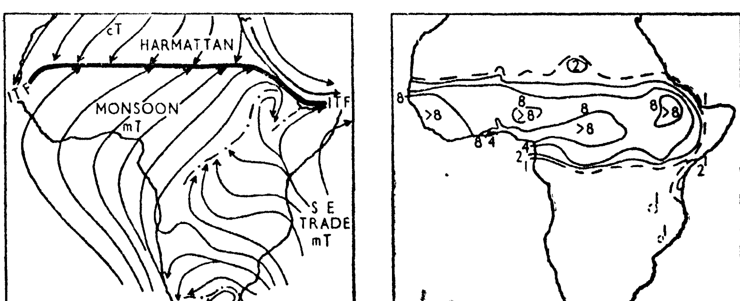


Fig. 13. Circulation patterns and rainfall over Africa, July (following Brooks and Mirrlees).

Brooks and Mirrlees distinguished five primary airstreams:

(1) The combined Atlantic N.E. Trade, Saharan north-easterlies (the Harmattan) and Egyptian northerlies; the part of this joint current which affects the present chapter is at all seasons hot, dry and rather dusty, being derived from the arid tropical deserts of North Africa. It is one of two major currents of cT air anywhere on earth.

(2) The S.E. Trade of the South Atlantic and its extension north of the equator, the so-called "S.W. Monsoon" of West and Central Africa. This air is moist and warm, being of mT type.

(3) The N.E. Trade or N.E. Monsoon of the Arabian Sea and East Africa. This air varies in properties, being dry, warm and stable in north-central Africa, but much more moist further south. In general it is cT air undergoing modification to mT.

(4) The S.E. Trade or S.E. Monsoon from the Indian Ocean. This current penetrates the continent to a varying extent during the year, varying in stability and degree of dampness. It is of mT type.

(5) Air derived from the relatively high pressure over Cape Colony and S.W. Africa is present at certain times of year. Dry as a general rule, it cannot be referred to any one airmass class, being most often maritime polar air from the Southern Ocean undergoing modification over Africa.

Though there are still many unexplained features of the rainfall distribution, this work has come closer to giving a rational, continent-wide picture of the mechanics of the atmosphere than any other study.

The Northern Winter and Southern Summer. At this time of year the sun's heating effect is concentrated over the high plateaus of South Africa, over which thermally induced low pressure develops. Pressure is high over the central and northern Sahara, the Indiar Ocean on both sides of the equator, and in the South Atlantic. Fig. 12 gives the average circulation for January.

North Africa is almost wholly covered by the north-easterlies south of the main high-pressure belt. This tremendous current of tropical continental air is warm and extremely dry, bringing sunny, rainless weather to almost the whole area. In general, it fails to reach the Guinea coast and the Abyssinian Highlands, but a broad arm of the current appears to flow south across the Congo Basin, and penetrates (in the view of Brooks and Mirrlees) right down to South West Africa. The southern part of the area covered gets a heavy rainfall, however, and it seems more likely that the truly Saharan air does not penetrate far beyond the Congo. Weather in this airmass is characteristically hazy, since it comes over long stretches of desert. In West Africa the hazy north-easterly is called the "Harmattan", a name sometimes applied to the whole current.

Abyssinia and the East African coast down to about 17°S. are covered by the north-east trade from the Indian Ocean. The northern part of this current is very dry, giving Abyssinia its dry season, but the southern part is rather more moist, though still warm and fairly stable. It penetrates inland roughly to the East African Highlands, where it meets the supposedly Saharan northerlies. The East African coast south of latitude 17°S. is covered by the south-east trade of the Indian Ocean, derived from the eastward-moving anticyclones of the southern sub-tropical belt. Very warm, moist and potentially unstable, these winds bring appreciable rainfall of the convective shower type to the East African coast. Precipitation is most frequent and heavy within the S.E. trade near the intertropical front. The latter oscillates considerably in position about its mean latitude of about 15°–20°S. so that a wide area gets appreciable rain.

The last current of importance in January is the S.E. trade of the Atlantic, which recurves on to the entire West African coast from about latitude 20°S. to the Ivory Coast. It enters Africa south of the equator across the cold Benguela current offshore, and is at first very cool and stable. A narrow coastal strip south of the Congo hence has a remarkably cool, dry climate. In the interior, however, it becomes warmer, and approaches the intertropical front as a warm, moist, unstable current. In West Africa it makes only a very short invasion inland, meeting the Harmattan quite close to the coast. The latter, being much the warmer current, rises over the wedge-like edge of the "monsoon", as the recurving trade is called, sometimes revealing its presence aloft to inhabitants of the Gold Coast and southern Nigeria by the milky veil of dust. At times it drives south to the coast at the surface, giving short spells of "Harmattan" haze, drought and heat. Only rarely does the wedge of monsoon air achieve great enough depth to give measurable rain even on the coast.

From this picture of the general January circulation, it is possible to draw certain conclusions about the rainfall distribution as a whole. Plainly the complete drought north of 5°N., and of the Horn of Africa arises from the continuous presence of dry north-easterly currents. The drought along the west coast south of the equator may be referred to the stabilizing influence of the Benguela current. But over the great bulk of southern Africa the rainy season is at its peak. In general this rainfall may be referred to the large-scale convergence into the huge low-pressure area centred over the plateaus and basins of the interior; much of it falls in heavy thunderstorms, for which the Transvaal and the High Veldt generally are notorious. Along the east coast, however, the orographic effect during the uplift of the south-east trades on to the plateau must aid the effect of convergence, as must the generally intense solar heating over the whole area.

The April-May Transition Season. With the northward movement of the sun, great changes in the circulation pattern take place. The southern half of the continent cools down, while the Sahara becomes intensely hot. Lowest pressure is transferred to the southern Sahara, roughly in latitude 15°N., and pressure over the high plateaus south of the equator rises considerably. The result is that the airstreams from the southern hemisphere migrate northward. The south-east trade spreads inland to the Abyssinian Highlands and the Congo Basin, while the Atlantic S.E. Trade, or "Monsoon", penetrates towards the Sahara and the Sudan. The Harmattan generally retreats, the intertropical front lying in April along 12°–15°N. latitude. In the south the S.E. trade is ousted from the veldt of the Transvaal and further south by cooler, drier, more stable air derived from anticyclones crossing the Cape.

The rainy area generally shifts northward. Drought sets in in the zone covered by anticyclonic circulation in the far south, and by April appreciable rain is largely confined to the area north of latitude 15°S. As the intertropical convergence zone passes north across East Africa, the heaviest rains of the year fall, chiefly from the south-east trade behind the zone. These "long rains" of April and May are the prelude to the winter drought. In West and Central Africa the rains associated with the S.W. Monsoon and the convergence zone between this current and the S.E. Trade spread rapidly inland, reaching latitude 10°N. by April.

The Northern Summer and Southern Winter. The heating of the Sahara is so intense at the time of high sun that a veritable monsoon is set up about the desert. The axis of lowest pressure reaches 20°N. in July, and the great insweeping currents from the southern oceans penetrate right up to this axis. The south-east trades of both the Atlantic and Southern Indian Ocean are at their coolest, driest and most stable, and the whole continent south of the equator experiences rainless weather (except for a tiny area near the Cape). When the two airstreams reach the very hot area north of the equator, however, their stability is entirely lost, and from the equator up to within about 200 miles of the intertropical front rainfall is heavy almost everywhere. The Arabian Sea N.E. Trades entirely vanish by June, when the Asiatic monsoons begin in earnest. In July, then, the circulation reaches its simplest plan; the three currents, Harmattan, S.W. Monsoon and S.E. Trade, dominate the entire continent within the tropics.

The main rain-belt extends from the coast of West Africa around Sierra Leone and Gambia to the Abyssinian Highlands: only the Horn of Africa and the coast of Eritrea escape. At times the rain-bearing Monsoon may penetrate even to the Tibesti Plateau in the central Sahara. Over the whole vast area the dominant cause of rain is thermally induced convection, producing heavy thunderstorms or showers with great frequency; the generally convergent motion adds to the rainy character of the weather over the Congo Basin and Abyssinian Highlands.

A clear picture of the structure of the atmosphere during this season has been given for West Africa by R. A. Hamilton and J. W. Archbold. The monsoon, they claim, is at all seasons cool and moist relative to the Harmattan. At the intertropical front the Harmattan therefore rises over the sloping wedge of monsoon air and may sometimes be observed aloft by the dust it carries. The entire monsoon wedge is covered by easterlies, the base of the latter increasing in level towards the south. These easterlies are sometimes dust-laden and dry, and are then without doubt an upslope continuation of the Harmattan. More often, however, they are very moist, even when they contain dust. H. Hubert called this moist easterly current the *Equatorial current*. Hamilton and Archbold are undecided whether it is Harmattan air moistened by contact with the monsoon or is derived from East Africa and the Indian Ocean over which high level easterlies are an important feature of the circulation. The final current present is the westerly anti-trade which lies above the northern edge of the monsoon and the Harmattan-Equatorial easterly. The base of the anti-trade rises rapidly southwards, and is rarely or never observed in summer along the Guinea Coast.

The intertropical front lies roughly 20°N. in July, so that the Guinea coast and French Equatorial Africa lie in deep monsoonal air. The character of the weather varies with distance from the Monsoon-Harmattan boundary. In the Harmattan it is always hot, almost cloudless and dry: usually there is haze, and for periods of days the haze may be very dense. In the belt 100–150 miles south of the leading edge of the monsoon weather is also hot but more humid. There is some cumulus cloud by day, and very exceptionally a shower or thunderstorm may develop. The main rainfall, however, takes place in the belt 150–450 miles south of the monsoon edge. The monsoon is here from 3,000 to 10,000 ft. deep, and the overlying easterlies are usually damp. Rain falls chiefly from afternoon and evening thunderstorms, which are often excessively violent; a considerable fraction, however, comes from the remarkable travelling disturbances called "haboobs" in the Sudan, "tornadoes" in West Africa, and "lignes de grain" by the French. These disturbances are discussed below. South of this zone of heavy precipitation, which is characterized by high temperatures and almost intolerable humidity, the structure of the monsoon begins to change; a semi-permanent inversion appears within it, apparently continuous with the trades inversion of the South Atlantic. The rainless weather characteristic of the whole southern part of the continent extends north to the Guinea coast, a narrow strip of which has a period of cloudy and cooler weather with only light showers and drizzle at times. This phase only lasts for the six weeks or so during whch the monsoon has attained its maximum northward penetration.

The travelling disturbances referred to in the previous paragraph are one of Africa's most distinctive climatic features. In form and appearance they resemble the severe line-squalls which accompany summer cold fronts in middle and high latitudes. Though they vary considerably in properties, a typical line squall of the West African variety is 75–150 miles long, usually extending from north to south, trailing off towards the east at the southern end. It moves towards the west at speeds not far from 30 m.p.h. through the monsoon. Dense cumulonimbus cloud and associated alto-stratus form the body of the cloud, which may be thirty miles wide, but there may also be bands of "roll cloud" along its leading edge just as there is on some cold fronts. As the disturbance passes over, there is a squall from the east, sometimes with gusts of over 60 m.p.h., and for an hour or two easterly winds prevail where there were formerly the monsoonal south-westerlies. Very heavy rain usually accompanies the squall, and may last for an hour or two afterwards. As the disturbance passes away the east wind dies away and the monsoonal south-westerly is resumed. Such are the so-called "tornadoes" of West Africa. In the Sudan, where the monsoon air is drier, the roll clouds are absent, their place being taken by a wall of dust or sand. These "haboobs" appear to be otherwise identical with the West African storms.

Though little is known about the mechanism underlying these remarkable and violent storms, it seems highly probable that they are closely related to the "waves in the easterlies" which affect the oceanic trade winds, though in Africa the easterlies are found some way above the ground.

The October-November Transition. As the sun moves south again, and heating is transferred to the southern half of the continent, pressure falls once more over the high southern plateaus and basins, and the circulation begins its second major change. During October the Harmattan begins to extend southward, and in November it is joined by the resurgent north-easterlies from the Arabian Sea. The result is the steady pushing back of both the S.W. Monsoon and the S.E. Trade until they occupy the restricted territory described above for January. In East Africa, some areas experience a minor maximum of rainfall as the north-east trades displace the south-east trades, the "little rains" corresponding to the "long rains" of April and May when the reverse change takes place.

CHAPTER XI

NORTH AMERICA

IN their quality of dramatic changeability, the climates of North America have no rival; nor is there any continent in which greater differences exist between region and region, and in the stimulus offered to human occupance and adaptation. The European settlers in the North American frontier encountered not one climate but many, each with its special hazards, but few without compensations to reward the settler who remained. With the exception of the Pacific coast, however, all these climates are united by their inconstant, sometimes violently erratic character. Even the humid sub-tropical climate of southern Florida is subject to the occasional hurricane passage, and only the sea-bound Keys are immune from the frosts of winter. Both from place to place and from time to time, then, the climates are widely differentiated.

The climates of North America are also unique in the high degree to which they depend upon facts of relief and other geographic factors. All climates on every continent exhibit such a dependence, but nowhere is it so simple, so direct, or so readily comprehensible as in North America.

Geographic Factors

In essence the relief of North America is divisible into three primary units:

(i) *The Western Cordillera*, comprising the broad belt of mountain ranges, plateaus and valleys lying between the Rockies and the Pacific Coast. Since air is always reluctant to traverse high ground, this tremendous barrier effectively divides the Pacific coastlands from the interior in both geographic and climatic senses of the term. Pacific airmasses enter the interior only after losing nearly all of their original maritime characteristics, so that the greatest of all oceans has little effect on the climate of much of the continent. In Europe, by contrast, the absence of a major obstacle to Atlantic airstreams and disturbances allows an appreciable rainfall from this source to extend far into continental Russia and even beyond. The Western Cordillera also radically affect the structure of all fronts and depressions crossing them from the Pacific, a vitally important process discussed at greater length later on.

(ii) *The central lowlands*, including all the land between the Eastern and Western Cordillera, and for climatological purposes including not only the interior lowlands, but also the Canadian Shield and the barren lands of the Canadian Arctic. This vast extent of monotonous land surface presents no barriers of importance to the free flow of airstreams in any direction. Because the Western Cordillera to a large extent cuts off the Pacific, the airstreams which make most use of this even surface are the great outbreaks of cP air from the Canadian Arctic, with northward moving streams of mT air from the Gulf of Mexico their chief rivals. The central U.S.A. and most of Canada are thus completely exposed to both tropical and polar sources, but are largely insulated from the Pacific. It is this ease of access from both north and south that accounts for the storminess and the rapid fluctuations of temperature which characterize the central lowlands.

(iii) *The Eastern Cordillera*, comprising the high ground of the Appalachian system and the related hills to the north-east. These hills are lower than the Western Cordillera, but their climatic influences are far from negligible. Both flanks are open to moist air masses from the tropical Atlantic, and the hills are from one end to the other free from conspicuous rain shadow effects. Their significance is primarily dynamic; they have a considerable effect on the movement of airstreams, though they are less significant than their bigger western counterparts in this capacity.

It remains in this review of geographic factors to discuss the disposition of land and water. There are two major embayments of the continental coast. The Gulf of Mexico extends a broad arm westwards of the tropical Atlantic, and it is from the Gulf that most invasions of maritime tropical air originate.

Hudson's Bay in a similar manner extends southwards of the Arctic Ocean, and many of the cold outbreaks of winter reach the Canadian mainland across it. The Bay does not freeze over until December, so that it considerably modifies continental polar air which crosses it in autumn. In summer it remains very cold from the ice melt, and it is a main reason for the cool and cloudy summers of the eastern Canadian Arctic. The last important water bodies are the Great Lakes, which markedly influence the climate of the areas around them. In winter their largely unfrozen surface heats and moistens the cold air crossing them, and snow or rain showers often fall on the coasts affected by onshore wind; sometimes the associated cloud may be the only dense cloud east of the Rockies. In summer they remain relatively cool, moderating the heat of the continental interior.

Fronts, Storms and Airmasses

North America derives its air supply from three principal sources, the Pacific Ocean, the Arctic and the Atlantic (including the Caribbean). It also experiences disturbances and storms associated with all three sources, and it is a little difficult to know where to start an account of the dynamic climate, especially since all the disturbances are intimately linked together. Perhaps we can get the easiest start by looking at figs. 8–9 which show average pressure distribution over the northern hemisphere.

The continent lies between latitudes 30N. and 70N., and so should extend roughly from the centre of the sub-tropical high-pressure belt to the sub-arctic low. In fact, however, the so-called typical zonal arrangements of the pressure distribution do not work in North America's longitudes. The sub-tropical high belt, for example, appears to be split into two separate cells on the July map, one west of California and the other near the Azores and Bermuda. As a result, the west coast of the continent might be expected to experience a relatively cool flow of west, north-west and north winds from the Pacific, whereas the east coast of the U.S.A. is exposed to a flow of tropical maritime air from the trade wind belt of the Atlantic. This is actually true of a large part of the year; it is quite rare for a true high-pressure belt to cover the whole of the south of the continent. On the January map of average pressure, this effect, still real, is obscured by the presence of high-pressure over the continent itself, the result of frequent anticyclogenesis in cP air.

cP air is the most frequent airmass over the continent as a whole. Originating over northern Canada and the Arctic archipelago, its properties have been described in Chapter V, and no details will be added here. Airstreams of cP origin frequently penetrate to the Caribbean, and at times even surmount the Rockies and Pacific ranges to reach the west coast. The chief rival of cP air over the eastern half of the continent is mT air from the Atlantic, which frequently invades the continent, in summer dominating the eastern U.S. Between the two characteristic airstreams is developed one of the major frontal zones of the Northern Hemisphere, the *Atlantic Polar Front*, on which many of the most intense storms affecting both North America and the North Atlantic originate. The day-to-day fluctuations of this front's position are large: in winter it is most often found along the Atlantic seaboard, where conditions are exceptionally favourable for frontogenesis at this season, the cold land contrasting with the warm waters of the North Atlantic Drift just offshore. Even at this season, however, it may be driven far inland; mT air may reach as far north as Michigan and the St. Lawrence valley, bringing the "January thaws" for which these areas are notorious. In spring and autumn the front usually lies up the Mississippi valley and across to Labrador, though at these times fluctuations in position are very great. In high summer the front is less intense, and tends to lie very far north, not far from the Canadian border. mT air at this season sometimes gets as far north as northern Quebec and Hudson's Bay. The intensity of the Atlantic Polar Front over North America may be hard for an English student to comprehend. The writer recalls one occasion in the winter of 1946–7 when there was a temperature difference of 62°F. within 150 miles, the front lying in between. It must be remembered, however, that the front is not permanent. It may be driven far out into the Atlantic, leaving the whole of eastern North America in polar air. Ultimately, however, another front will take up the characteristic position and properties, so that on most weather maps we can identify an Atlantic Polar Front.

Along the entire Pacific coast the dominating airmass at all seasons is mP from the Pacific itself. This air approaches the coast on the north flank of the sub-tropical high. It varies in properties from the fairly warm and stable type characteristic of south-westerly weather in Britain to cool, cloudy, showery currents from the north-west, again very like that which comes across Britain in the rear of depressions. The west coast hence experiences the same equable, rather monotonous weather that is typical of Britain; there is no parallel for the dramatic events of the eastern climates. Occasionally in winter cP air comes from across the hills to the east, just as in Britain, giving a cold snap reminiscent of that we experienced in the recent hard winter of 1947.

At most times of year, there is a marked contrast between the continental airmasses which predominate east of the hills and the maritime air of the Pacific coast. Normally these airmasses are separated by the hills themselves rather than by a front. If, however, either airmass is forced to cross the barrier, it is brought into direct contact with the other, and frontogenesis may occur. The fronts which may form in winter east of the hills when mP air enters the central lowlands may be quite intense. Since they are first seen over the foothills, the term "Rocky Mountain Front" may be used as a collective term for them. The mP air behind such fronts is warmed and dried by its crossing of the mountains, appearing over the High Plains and foothills as the celebrated *Chinook.* It is not uncommon for the Chinook to appear with remarkable suddenness, taking winter temperatures up from below zero to 40 or 50°F. in an hour or two. As the mP air moves east across the prairies, however, it is rapidly cooled, and is quite cold by the time it reaches Eastern Canada or the U.S. In summer the maritime air is cooler than the cP air over the interior, bringing delightfully cool, dry and brilliant weather as it moves across the continent.

Pacific mT air is a rarity. It sometimes occurs in the warm sectors of depressions forming unusually far south during the winter. The occasional heavy rains and snows of California are chiefly brought in this manner.

In the winter cases where cP air moves across the Western Cordillera to the Pacific coast and out to sea, the fronts which form between this air and mP air are collectively known as the Pacific Arctic front; they are common features of the circulation of depressions which become stationary just off the coast.

The two remaining airmasses are of less importance. mP air of Atlantic origin occasionally enters the north-eastern parts of the continent, bringing cool, cloudy weather with it. During summer, this airmass is common in Newfoundland, Labrador, and the eastern Arctic, where the very cold sea surface leads to general sea fog or low stratus. The other airmass, cT air, is largely confined to the high and arid plateaus of the south-west, where it gives torrid heat in summer and warmth even in midwinter. The extremely high day temperatures in this airmass are to some extent offset by low humidities. The ultimate source of the air is in most cases subsidence from the overlying westerlies.

The storm systems affecting the continent follow readily from this discussion of the frontal and airmass climatology. Each of the main frontal belts is associated with characteristic storm types: in addition, the continent is frequently affected by depressions from the Pacific that originate on frontal systems in the western or central Pacific. In fact, however, the relationship between the different types of storm is so great that we can best discuss them on the basis of area of origin. rather than by classifying them on a frontal basis. The table below summarizes the storm families in this way.

Storm types affecting North America

Type	*Area of origin*	*Fronts and airmasses involved*	*Seasons of occurrence etc.*	*Remarks*
PACIFIC	W. or mid-Pacific.	Western Pacific Polar Front. mP air; some mT in winter only.	All months	Confined largely to Alaska in summer.

Storm types affecting North America

Type	*Area of origin*	*Fronts and air-masses involved*	*Seasons of occurrence etc.*	*Remarks*
ALBERTA	Rocky Mt. foothills north of 50°N. as a rule.	In origin frontal lee-depressions; often involve Rocky Mt. Front. cP and mP air.	All months	The most frequent type; often quick-moving and intense in winter.
ROCKY MT.	ditto	Waves on Rocky Mt. Front. cP and mP air.	Winter half year	Minor in scale and infrequent.
COLORADO	High plains or Rocky foothills south of 50°.	Waves on Atlantic Polar Front during its inland phase.	All months	Though these storms develop as frontal waves, the initial trigger action is probably like that of the Alberta group.
ATLANTIC COAST	Gulf of Mexico, Atlantic Coast belt of U.S., especially near Cape Hatteras.	Waves on Atlantic Polar Front. cP and mT air.	All months, uncommon in summer	Many severe Atlantic storms develop here.
HURRICANES	Caribbean, Tropical Atlantic.	mT only unless Atlantic Polar Front is drawn in later.	Late summer and autumn	Tropical revolving storms.

As in all other parts of the Northern Hemisphere, the dominant direction of motion of all save the tropical hurricanes is from some westerly point, in accordance with the high level circulation. It is therefore appropriate that we should begin with disturbances that come in from the Pacific; a great deal of North American weather is set off in one way or another by high or low level perturbations of Pacific provenance. The surface depressions that approach the Pacific coast form well out to sea—some as far away as the China Seas—and are generally mature and well occluded before they strike the coast. A few form nearer at hand, and still have warm sectors. In general, they closely resemble the storms which come across Britain from the Atlantic; in fact the whole climate of the Pacific coast is remarkably similar to that of western Europe. Periods of showery weather in mP air alternate with rainy spells on occlusions and other fronts coming in from the west. This kind of weather prevails all through the year in Alaska and British Columbia (except in the south) but in the summer the storms are fewer, less intense and smaller in size; what is more, their fronts do not penetrate into southern British Columbia or the U.S. These areas enjoy a settled and largely rainless summer as a result. At all times of year the storms prefer a northern track; their normal approach is towards British Columbia and Alaska. In California, indeed, direct storm passages are unusual; most of their winter rains come from deep southward penetrations of the cold fronts or occlusions passing further south. The occasional storm that does cross the Golden State, however, often involves mT air in its warm sector, and heavy rains and snows result. One such storm was described in George Stewart's masterpiece, *Storm* (1941).

A large fraction of the Pacific storms which affect North America never truly cross the coast; their centres stagnate and eventually fill up over the Aleutians or Alaska, only their fronts crossing the coast further south in the more thickly peopled areas. The Plateau region behind the coast range of British Columbia gets an appreciable snowfall from the eastward-moving fronts of such systems in winter. When the storm itself does succeed in getting across the mountain barrier and reaching the Prairies or the Mackenzie valley, it undergoes considerable rejuvenation; the Rocky Mt. Front is disturbed, and the Chinook sets in south of the centre. Most of the sudden winter thaws over the Alberta foothills are caused in this way; temperatures may rise 60° to 70° as the mP air sweeps down on to the plains. After crossing the plains with its new-born warm sector of Chinook air, the storm usually eventually reaches the Atlantic Coast. In winter it may take a variety of courses across the eastern half of the continent, but in summer the course is usually far to the north. Behind each storm cP air streams southwards, in winter usually penetrating quite deeply into the U.S.A. A feature of the Pacific type of storm is their rapidity of motion, which rests largely with the strength of the upper westerlies, often of abnormal strength at these times.

An important variant of this sequence of events occurs when cyclogenesis occurs over the eastern foothills of the Rockies or the adjacent plains. Instead of crossing the hills, a majority of Pacific disturbances "induce" fresh development beyond them in a manner not yet fully understood. We can recognize two broad types of such disturbances, the so-called "Alberta" lows, which develop east of the northern Rockies, chiefly in Alberta and the Mackenzie valley, and the "Colorado" lows which develop further south. The chief difference between the two types is that the second usually involves the Atlantic Polar Front, and so comes to have a warm sector of mT air, ultimately bringing to the eastern half of the continent heavy and widespread precipitation. The Alberta type exactly resembles the Pacific type just described, and it usually is difficult to tell the one from the other; in fact it may be argued that Pacific lows which appear to cross the Rockies are really reborn over Alberta or the Mackenzie valley.

So far we have been discussing the types of disturbance which either originate over the Pacific, or at least owe their origin to Pacific disturbances. We come now to the group of storms which involve mT air from the Atlantic in their circulation. Such storms are confined to the eastern half of the continent, and are noted for their rainy character. We can distinguish two main types:

(a) those storms that form on the Atlantic Polar Front when the latter lies over the continental interior; such storms usually form over the plains west of the Mississippi—hence the nickname "Colorado depressions"—and travel north-eastwards into Ontario, Quebec and Labrador; and

(b) those forming along or just off the Atlantic coast—hence the term Atlantic coast depression. These storms follow a wide variety of paths; a very common winter track takes them north along the coast itself, a course which brings severe gales to the seaboard states.

The first of these types is conspicuous at all times of the year. In summer such storms are likely to travel very far north;

their wide open warm sectors then bring the worst of summer heat waves, often accompanied by thunderstorms. The cold fronts of such systems undercut moist, dynamically unstable mTK air, and widespread storms of great violence accompany their passage. A typical feature is for a squall line to precede the actual front by 100 or 200 miles, also with heavy thunder storms. The Tornadoes which are so great a hazard in the middle western states develop in this position. Behind the cold front, cool and dry cP air sweeps southwards, to the great relief of the average American. Words can hardly express the sticky discomfort of life in a great city like Chicago with day temperature maxima up to the 100° mark, and dewpoints in the upper 70s. The same systems in winter give heavy rains in the south and snow in the north. There is usually a wide belt of freezing rain just north of the centre's track.

The second class (the Atlantic coast type) is uncommon in summer, but very frequent for the rest of the year. During the fall and spring they give heavy rain to a wide area of the east. In winter, as in the case of the Colorado type, the rain is replaced in the north by snow or freezing rain. Many of the severe blizzards of the eastern states, Quebec and the Maritime Provinces, occur during the passage of such storms. Atlantic coast lows are especially liable to form off Cape Hatteras, often as waves on the warm front of a Colorado-type low moving towards the Great Lakes. Waves are also very likely to form here on cold fronts which have just crossed the Appalachians. After moving up the American coast, nearly all such waves finally die out, fully occluded, in the Davis Strait or Iceland areas.

The only important storm type remaining is the tropical hurricane. These are the small but intensely violent tropical revolving storms already described in detail on p. 110. Essentially confined to late summer and autumn, these destructive storms affect chiefly the coastal states facing the Atlantic and the Gulf of Mexico. Georgia, the Carolinas, Florida and the Gulf States are the areas worst affected, but serious damage has been done as far north as the Canadian maritime provinces. Cities like Miami and New Orleans live through the danger months in a perpetual state of watchfulness;

the hurricane plays the same part in the conversation of the Floridan that Baedeker raids did in England in 1943–4! Whole coastal areas are evacuated if the hurricane approaches, and the floodable area of Lake Okeechobee in Florida (where there was a terrible death-roll in 1928) is evacuated, well in advance of danger.

The chief climatological significance of the hurricanes (which are fortunately not very frequent) lies in the widespread heavy rain they bring. Even though the storm may "blow itself out" shortly after it passes inland, the moist air which is left behind is a fertile source of rain if it becomes involved in the circulation of other depressions. Twice in the fall season of 1947 (as the picturesque American phrase for autumn has it) rain and thunderstorms occurred in Montreal when such air was brought north-eastwards.

Anticyclones

To render complete our study of the dynamic climatology of the continent, we must examine the development and movement of high-pressure systems. The migrant high is as characteristic a feature of the American climates as is the travelling depression. The properties of American anticyclones were described in Chapter VII, and it will only be necessary at this stage to sum up their travel and seasonal incidence. They are of two main classes:

(i) *Cold anticyclones* which form over the Arctic Ocean or the cold north-western part of the continent, afterwards travelling eastwards or south-eastwards towards the Atlantic seaboard. They occur at all seasons, but are commonest and most quick-moving in the winter, when they are accompanied by intensely cold weather. The Arctic Ocean on the Bering Sea side of the Pole is very often the scene of high-pressure at this season, a ridge of high-pressure often linking the Siberian anticyclone to another anticyclone over the Yukon, the Mackenzie Basin or the Arctic archipelago. It is from this belt of high-pressure that many of the migrant highs emerge; sometimes a centre can be traced all the way from north-east Siberia to the St. Lawrence Valley and beyond. Still more, however, appear to begin their life over the American mainland, often between Hudson's Bay and the Rockies. The first signs of their formation are strong rises of pressure, usually in the rear of an Alberta low. The cold front of the storm sweeps southwards to the Gulf of Mexico, and the cold high moves rapidly towards the Atlantic, usually merging in the end with the Bermuda-Azores high. In summer the movement of these systems is less rapid and less predictable, but they are still a real and welcome feature of the circulation, at least in Canada and the northern United States.

(ii) *Warm anticyclones* are more various both in origin and in properties. Warm highs often enter the United States or British Columbia from the Pacific. Once over the interior, however, the northerly circulation ahead of their centres brings cP air down into the southern states, and the superficial circulation rapidly comes to resemble that of a cold high; it may indeed be argued that many of the apparently cold anticyclones of winter originate in this manner. The south-eastern parts of the U.S.A. may also be covered by the westernmost parts of the circulation of the Atlantic sub-tropical anticyclone; in summer this is true for protracted periods.

A Regional Division of North America Based on the Source of Rainfall

The foregoing account of the dynamic climatology of the continent allows us to postulate a regional climatic division with more confidence than is usually possible in such matters. In general the division suggested follows the traditional division of the country into three broad meridional belts, viz. the Pacific coastal rainy area, the arid western plateaus and plains, and the so-called "humid east". We can summarize their extent quite briefly:

(i) *The Pacific Province*, in which precipitation falls exclusively from air of Pacific origin, and in which precipitation during the year is appreciable in amount. Rain and snow fall chiefly in association with the occlusions or cold fronts (warm fronts also in the north) of Pacific depressions, almost always from mP air. In the north there is also an appreciable fall from instability showers in the cooler variety of mP air. The Pacific Province can be divided into two halves, a southern half in which rain falls chiefly in winter, and a northern half in which precipitation occurs all the year round. It is also most noticeable that, whereas in California heavy precipitation is virtually confined to the area west of the Sierra Nevada summits, in Oregon, Washington and British Columbia, winter precipitation extends eastwards of the Cascades and Coast Range to the Rockies, so that all the intervening plateaus get an appreciable snowfall. In Alaska, the penetration is once again confined to the coastal regions south of the first ranges. These facts are explained by the fact that the main track of cyclones penetrating to the interior is across the regions mentioned as receiving winter snowfall.

Rainfall distribution in the Pacific coastal province is subject in a remarkable degree to the local effects of relief, a topic discussed at greater lengths in most of the standard reference works.

(ii) *The Atlantic Province*, in which the ultimate source of most precipitation is the sub-tropical Atlantic. Rains in this area fall chiefly from mT air or from cP air which has been moistened by a deep southward penetration across the Atlantic or the humid south-eastern state of the Union. In general the penetration of the moist airmasses is much deeper in early summer than at other seasons; hence we can distinguish two main sub-provinces, the true humid south-east receiving rain at all seasons, and the peripheral belt getting rains chiefly in the summer half of the year.

The dynamic background of rainfall in this province is complex. Heaviest and most widespread rains or snowfalls occur when mT or moistened cP air is involved in the circulation of Pacific, Alberta, Colorado or Atlantic Coast types of disturbance. In summer, however, there is a considerable fall from surface heating thunderstorms in mT air, chiefly in the hilly eastern cordillera. We may note that the frequency and depth of westward and northward penetration of really moist airmasses varies considerably from year to year. The disastrous droughts of the 1930s in the western plains and prairies, were the result of a prolonged failure of the "American monsoon" as the moisture-laden currents of mT air are sometimes called. On the other hand, the north-eastern quadrant of the continent has a uniquely reliable rainfall; not only are these areas deeply within the zone of penetration of tropical airmasses, but they are also the region where most of the characteristic storm tracks converge. A high frequency of cyclonic depressions at their most active stage of development combines with an abundance of moist airmasses to give such areas as Quebec, the Maritime Provinces and New England one of the world's most reliable sources of rainfall.

(iii) The arid belt lying between these chief rainy provinces has been divided into a *Northern Dry Belt* and a *Southern Arid Belt*. Both belts have much in common: both are beyond the reach of Atlantic mT air with its attendant moisture; both are also beyond the reach of humid mP air. The latter is the predominant airmass apart from cP, which is commonest in the north. But the mP air arrives over these regions largely free of precipitable moisture, much of which is left behind during its crossing of the hills to the west.

In the northern dry belt winter precipitation is light, since the extreme cold of the prevailing cP air precludes moisture. Very deep frost penetration results from the absence of a deep snow-cover. Spring and autumn snowfalls and light summer rains come chiefly from Pacific mP air or modified cP air involved in Pacific or Alberta type depressions. The southernmost regions may get an occasional heavier spring or early summer rainfall from an unusually deep north-westward invasion of Atlantic mT air. But throughout the belt, rainfall and snowfall are both scanty and unreliable. This lack of reliability has been a major factor inhibiting economic stability in the Canadian prairies, the western parts of which lie within this belt.

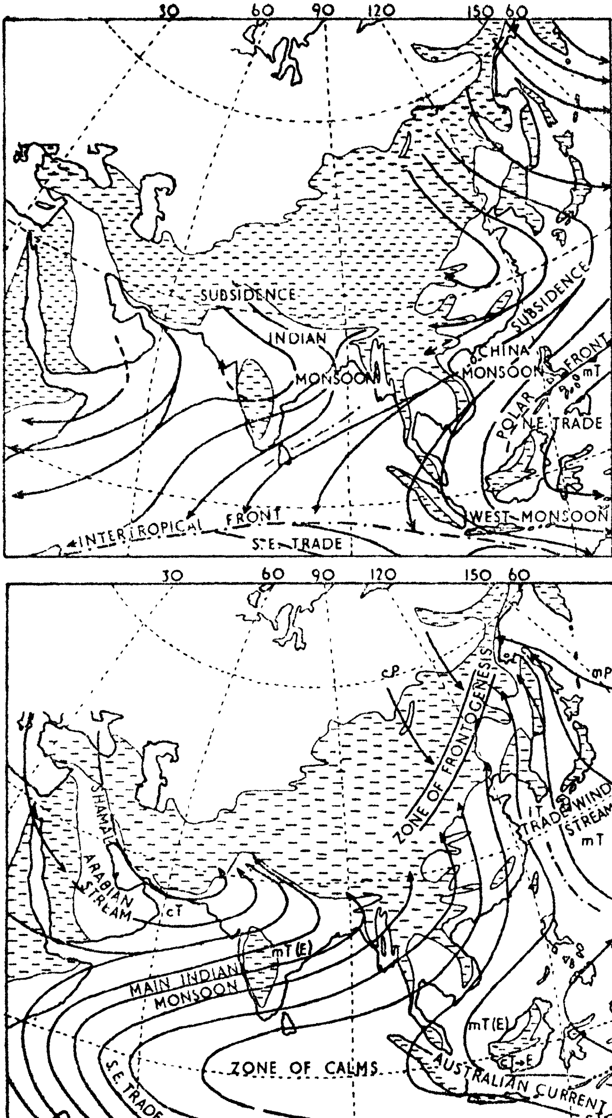
The southern arid belt is in general drier and hotter than its northern equivalent. For long periods in summer the region is covered by very hot and dry cT air. To the general lack of rainfall is hence added excessive evaporating power in the air, with a resultant intensification of the drought.

CHAPTER XII

THE MONSOONAL CLIMATES OF THE EAST

LIKE most words which have both a technical and a popular usage, "Monsoon" has long ago ceased to have an explicit meaning. To some, the word implies a season characterized by a distinct type of weather differing from that of the rest of the year. To others, "monsoon" means a wind-system occupying a definite span of time and space, and bringing a peculiar suite of weather phenomena in its train. The usage of the meteorologist inclines towards the latter sense. The terms S.W. Monsoon and N.E. Monsoon are very common jargon, showing clearly that the professional has the wind-system uppermost in his mind. All of us know, however we define the word, that the climates of east and south Asia are monsoonal in character, and are thus differentiated from those of similar latitudes elsewhere. The whole of this gigantic area, which stretches across Indonesia and the Philippines to include northern Australasia, is set apart from the rest of the world in the circulation of its winds and the behaviour of its weather. Though monsoonal tendencies are visible elsewhere (see for example Chapter X) they are nowhere comparable with those which affect the Far East.

The monsoon climates are characterized by clearly marked seasons, each with its specific type of wind and weather. In most areas there is a complete reversal of wind between summer and winter, accompanied by remarkable changes in prevailing weather. Over southern China, Indo-China, the Philippines and the surrounding seas, for example, the summer is the season of the S.W. Monsoon, a broad stream of hot, humid equatorial air which brings to all areas a prolonged period of sultry, wet weather. The winter on the other hand brings the N.E. Monsoon, which imports very cool, cloudy weather, largely free of rain over much of the area. Such changes are



Figs. 14 and 15

Sketch-maps of typical circulation patterns at climax of monsoons, in January (Fig. 14, above) and July (Fig. 15, below). High ground is shaded.

characteristic of the entire area, though the detailed pattern varies greatly from point to point.

The Dynamical Background of the Monsoons. The Monsoons arise from the thermal contrasts between land and sea; they are the dynamical expression of the condition normally called "continentality". In middle and high latitudes it is well known that the land surfaces of the continental heartlands cool down rapidly in winter and warm up in summer, whereas the oceans and the west coasts of the continents have a far more conservative temperature régime. This phenomenon is largely confined to Asia and North America, for the other continents lie in lower latitudes, where seasonal inequalities in the radiative balance are less noticeable. In North America, where the relief favours north-south (meridional) circulation, the effects of the winter cooling are to some extent mitigated by northward penetrations of tropical air; similarly the heat of summer is lessened by outbreaks of polar continental air from northern Canada. In Asia, however, the grain of the country is entirely hostile to meridional circulation, and this, coupled with the greater size of the land mass, makes the seasonal contrasts much more dramatic. The hill-valleys of north-east Siberia are perhaps the coldest part of the earth in mid-winter, yet in summer these same localities enjoy weather as warm as that of Northern England. Even as far south as the Yangtse basins, winter ice may impede navigation in the same localities which suffer the world's most oppressive climate in the summer months. Small wonder, then, that the monsoons are an Asiatic rather than an American phenomenon.

The Winter Monsoons

The winter cooling over the heart of Asia leads to fairly general subsidence over truly enormous areas of the continent. A more or less permanent pool of intensely cold air accumulates over the lower ground, and pressure rises to high levels, sometimes attaining sea-level values of 1070 mb. The weight of this pool creates what is in effect a cold anticyclone, which remains stationary over eastern Siberia through most of the winter. At its full development, the Siberian winter anticyclone may extend to the Arctic Ocean, to south-east of Japan, and west to the Black Sea; over the whole of this area circulation is under its control. The anticyclone is very shallow, and at no great height the circumpolar westerlies are found, being very little disturbed by the cold weather below them. So shallow is the cold pool that many of the higher plateaus of inner Asia project up into the westerlies, and see nothing at all of the monsoons which affect the lower ground. The Siberian high is also sharply limited on the south, being shut off from India by the Himalayas, which extend far up into the westerlies. India has an independent circulation of its own.

The whole of eastern Asia comes under the influence of the Siberian system. A great outflow of bitterly cold polar continental air streams southwards out of north-eastern Siberia across China, Japan and the China Seas to reach the equator between Borneo and Ceylon. This current is the *North-East Monsoon*, so called from its dominant direction of flow. Its characteristic flow pattern is illustrated in fig. 14. For at least two-thirds of the winter synoptic charts of the Far East show circulation closely resembling this idealized picture. It will be noted that the circulation over India is obviously distinct from that of the Far East proper.

The onset of the winter monsoons is a welcome relief after the heat of summer. In the Far East the first monsoonal outbreaks occur early in September and the monsoon is fully established by mid-October. In India the onset is more gradual, a spell of hot, dry autumnal weather interposing between the end of the summer monsoon and the winter monsoon. The monsoon begins to give way in the Far East in late March or early April, but in India the breakdown comes rather earlier. When fully developed the north-east monsoon extends far out into the Pacific, the South China Sea and the Indian Ocean. Along its south-eastern edge the monsoon is in contact with maritime tropical air derived from the trade winds of the North Pacific. Conditions are often frontogenetic along this frontier zone; the name "Western Pacific Polar Front" is given to the front often found in this position. Frontogenesis is favoured by two factors; one is the extremely strong temperature gradient between the cold land and the warm Pacific, and the other is the deformation field existing between the Siberian and North

Pacific sub-tropical highs. Most of the intense extra-tropical depressions affecting the North Pacific originate as waves on this front. They travel north-eastward from the China Seas towards the Aleutians, where they are at maximum intensity.

Over India the mid-winter period is one of quiet, calm weather, outstandingly cool by comparison with the other seasons. The monsoon, sketched in fig. 14, is a cool, dry stream of air which moves from the north-west down the Ganges valley, passes out over the Bay of Bengal, and then recurves to pass across southern India and Ceylon from the north-east. Along its southern margin over the Bay it comes into contact with the stream of north-easterlies which come across the Malay peninsula from the South China Sea. Usually the stream is merely the extension of the China N.E. Monsoon, but occasionally it is moist, warm maritime tropical air from the Pacific. On these latter occasions there may well be a clearly marked front across the Bay between the Indian Monsoon and the north-easterly from the Pacific. Winter weather in the monsoon lands depends very strongly on the strength, persistency and characteristics of the winter-monsoon. Both in India and in China we can distinguish (i) the weather associated with the full monsoon circulation and (ii) weather which results from breakdowns in the normal circulation. The latter class is less frequent than the former but contains most of the bad weather. Before we discuss regional climates we must hence dispose of the lulls or breaks in the monsoon; these interruptions are far more common than is sometimes supposed. We can classify monsoon lulls as follows:

(a) *Waves on the Western Pacific Polar Front.* Such waves occasionally give rise to active depressions which bring bad weather to the extreme south of China, the south-east coast and Formosa, Japan and the Eastern Sea. The normal position for the front is sketched in fig. 14. When quiescent, it has a narrow belt of low stratus and stratocumulus with drizzle or slight rain on its northern side. The front oscillates in position, at times even penetrating as far as the Si'-Kiang Valley, and at other times lying entirely south of the Philippines. Waves are most likely to develop on it when it lies just north of Luzon and south of Formosa. Such storms may give gales and heavy rain over the Eastern Sea and southern Japan (where snow is also common), but they rarely affect the mainland apart from the extreme south-east coastal districts of China. Depressions of this sort quite frequently cross the Pacific, reaching Alaska or British Columbia. The development of such storms does not really disturb the monsoon over most of the Far East.

(b) *Disturbances from the West.* The true lulls in the monsoon, both in India and the Far East, are traceable to disturbances coming from the west. As far as the Far East is concerned, these disturbances appear to travel across Siberia and the plateaus of central Asia from Europe. Most of them are probably travelling trough-like disturbances in the upper, circumpolar westerlies which are known from most parts of the northern hemisphere. In some cases the high-level disturbance is accompanied by an old, occluded surface depression which crosses Siberia. Though they invariably weaken as they cross the continent, not a few of these European or Atlantic storms actually penetrate to Manchuria or North China every winter. At such times the Siberian high, usually regarded as permanent, is displaced far to the north-east, or may even be absent. As the anticyclone gives way, the supply of cold air is cut off, and temperatures rise throughout the Far East. Sometimes the diminishing Siberian system sends off a travelling cold high, which moves south-eastwards across Manchuria, North China and Japan, giving a brief spell of settled weather.

Ultimately the lull in the monsoon is terminated by a renewed intensification of the Siberian anticyclone, and a new burst of the monsoon sweeps southwards across the Far East behind a strongly developed cold front. Before the monsoon is fully re-established, however, there are usually some stormy episodes. During the lull, it is common for warm, moist air (either mT or modified cP) to spread far northwards across China and Japan, giving a welcome relief from the cold of the monsoon. Interaction between this warm air and the advancing cold front of the fresh monsoonal current usually gives rise to one or more of the winter cyclones which are a conspicuous feature of the lulls. These storms develop as true frontal waves, usually over China. The commonest place for them to form is over the middle Yangtse, where the configuration of the ground tends to deform southward-moving fronts into a wave-like form. The "Yangtse depressions", as they are called, give periods of light rain or snow to central China several times in a normal winter month. They travel east towards Japan where, after drawing in moister air, they may give fairly heavy precipitation.

The western disturbances of India are of like origin. The majority of them are surface expressions of disturbances in the high-level westerlies which here, as elsewhere in the east, are permanent features of the winter circulation. Some are old depressions which travel from the Mediterranean across the Levant, Iraq and Iran to cross Northern India. Such storms give a light winter rainfall or snowfall to northern India (especially in the hilly north-west) but rarely affect central and southern India.

The Full Monsoon. But the true stamp of winter in the East is imposed by the monsoon itself, important as the lulls are to the climatologist. The weather brought by the monsoon is fairly constant in any one locality, but differs remarkably from place to place.

In North China, Manchuria, Korea and Japan, the monsoon is a deep, cold, unstable and little modified cP airstream. On the mainland it brings clear, rather blustery weather with severe frost; even as far south as Peiping (lat. 40°N. approx.), the river freezes over thickly enough for a month or more to impede navigation. Since even the travelling cyclones of monsoon lulls bring little precipitation to these areas, North China and Manchuria have a clear, cold, largely snow-free winter, permitting the very deep penetration of soil frost, to the detriment of spring cultivation. As the monsoon passes out across the Japan and Eastern Seas, however, it is warmed and moistened rapidly at the surface. As it approaches the Japanese coast, it becomes heavily charged with cumulus cloud, from which wintry showers fall. The entire north-west coast of Japan, right up to the summits of the country's mountainous backbone, is blanketed by this cloud whenever the monsoon blows, and snow or rain falls intermittently throughout the day. The average cloudiness of this coast is over 9/10 (i.e. sky averages 90 per cent covered) in January over wide areas, and some districts have average precipitation as high as 20 in. in this month, nearly all of it is in the form of snow. In December 1927, during a spell of the monsoon, the town of Takata had 95 in. of fresh snow in the last three days of the month. These figures are all the more remarkable when it is recalled that this precipitation is almost wholly convective or orographic; little frontal precipitation is involved. Beyond the mountains the south-east coast enjoys sunny and warmer weather during the monsoon, the clouds having been dispersed by the descent of the air. They re-form over the warm waters of the Kuro-Shio just offshore.

In about latitude 30°N., the monsoon current invariably begins to acquire an anticyclonic circulation; it turns to the right, becoming first a northerly and then a north-easterly current. It enters South China, Indo-China and the Philippines after a prolonged journey over the warm Pacific, so that it is fundamentally different in properties from the monsoon of the north. The difference is accentuated by changes in its vertical structure; the development of anticyclonic "spin" in the current is accompanied by subsidence, divergence (see Chapter IV) and stabilization. The airstream which crossed Japan deep and unstable becomes a shallow superficial current in latitudes 30°N. At Hong Kong it averages only 3,000 ft. in depth. Above the monsoon, and separated from it by a very stable mixing layer, are the warm, dry, circumpolar westerlies, which are here derived from the Thibetan plateaus. Although the monsoon itself continues unstable, then, there is a very stable layer just above it. The convection currents set up by the warm sea create abundant clouds, just as they do over the Japan Sea, but the rising air is unable to penetrate through the stable layer above the monsoon. Instead the clouds spread out into an even, unbroken sheet of strato-cumulus covering thousands of square miles. Light rain showers or drizzle fall at intervals. Dull weather of this kind spreads all over the Eastern Sea and China south of the Tsingling Shan throughout the spells of monsoon. The contrast between North and South

China is hence remarkable: the north, clear, dry and bitingly cold, the south, overcast (with fog in the hills), cool and damp, with frequent light rain.

The southernmost parts of the monsoon area have a very different climate. The Philippines, Indo-China, Thailand and northern Malaya are affected by alternate periods of trade-wind and monsoon, the former bringing warm and humid weather, the latter cooler and less oppressive conditions. In most areas weather is fair or fine in both airmasses, both of which blow from N.E. Furthermore in these latitudes there is less contrast between the airmasses than further north, and the Western Pacific Polar Front is at most times diffuse. There are, however, certain important exceptions to this generalized account. There is very heavy winter precipitation in the east-facing hill ranges of the Philippines, Annam, southern Thailand and Malaya, especially in the early days of the winter monsoon. This rain falls as heavy convectional storms, chiefly from mT air either just ahead of or behind the Western Pacific Polar Front during one of its southward migrations. Though this rain is primarily orographic, the hill-crossing is not of itself sufficient to give rains on the observed scale, and the extra impulse given by the arrival of the true monsoon current is certainly the real trigger action. Such rains become less common in mid and late winter, when the monsoon current dominates the region more continuously.

We may note in passing that the N.E. Monsoon of these southern regions is a composite current, in part polar, in part tropical in origin. Quite small oscillations of the front between the two north-easterly currents of monsoon and trade wind can radically change the airmass distribution over large areas. The inconstant weather of south-east Asia largely springs from the oscillation of this frontal system, even though it may be very hard to detect by normal surface analysis.

Over Indonesia the combined current turns southward and then eastward as the westerly monsoon of Java, Timor and northern Australia. Throughout this vast equatorial belt rainfall and cloudiness are both high. Rainfall is convective in type, the trigger action being in part orographic, in part of a dynamical origin so far inadequately explored. Ultimately the monsoon current enters the vast, overheated interior of Australia, where a permanent low pressure-system exists during the hot months of the southern summer.

It is tempting to say in summary, that the N.E. Monsoon of the western Pacific-China Seas province is a vast stream of air originating in Siberia and terminating after many vicissitudes in the Australian desert. Such a judgement would be superficial and entirely ill-considered. In every part of the great circulation system air is constantly being taken away from or added to the monsoon current; subsiding air from the westerlies, for example, is added over a wide area, and in subtropical latitudes there are substantial accretions of maritime tropical air. The grand total mass-flow into the equatorial belt across latitude 15°N. is several times greater than the mass transported southwards across 30°N., even if one computes the figures between the same two streamlines. The reader can readily confirm this for himself by a glance at a chart of mean winds. The only possible source for the additional air is from above, i.e. it is derived by subsidence from higher levels. Such subsidence is very general south of latitude 30°N., both during spells of the monsoon and during lulls. In brief, the monsoon must never be regarded as a single, homogeneous current making its way as a unit across a third of the earth's circumference. It is more proper to regard it as a specific flow-pattern imposed on the general world circulation by the maldistribution of land and sea.

The Summer Monsoons

In both India and the Far East the spring months are a time of rapid rise of temperature and humidity. In India circulation in April and May is very sluggish, the chief result being the slow penetration of the superficially humid air which gives rise to the notorious hot season. Both months are characterized by extremely high temperatures with moderately high humidities. The heat results from the largely cloud-free skies; the only break in the fine weather is caused by occasional thunderstorms, which are often of a squally character. Such storms give little rain in the arid interior, where they often cause dust-storms. In the Far East, on the other hand, the rise in temperature is accompanied by heavy rains in all areas except the north of China, Mongolia and Manchuria.

The breakdown of the winter monsoon in the Far East is a gradual business. In March and April there is an increasing tendency for "western" depressions to cross Siberia and Manchuria to reach Japan or the Sea of Okhotsk. Behind each storm there is a reassertion of the monsoon, but the intensity of the outbreaks lessens towards the end of April. In effect, what happens is that the lulls in the monsoon increase in frequency and duration. The disturbed cyclonic weather typical of such lulls in winter become the predominant weather type in spring. The heavy rainfall affecting all areas except the extreme north is predominantly frontal in type, the heavy character of the falls being due to the progressive rise in humidity which takes place in the warm, northward spreading air.

In May the heaviest belt of frontal rains lies across South China. By June it has moved northward to the Yangtse Valley and by the end of the month lies across southern Japan. By this time the air south of the main frontal systems is usually maritime tropical or equatorial, i.e. the true summer monsoon itself. The resulting rains in central China and Japan are hence very heavy, and mark the onset of the monsoon. The special combination of heavy rain, low cloud, high temperature and humidities occurs every year in June and early July in these regions. To the Chinese these rains are the Mai-U (Bai-U in Japan), the plum rains. At their conclusion the monsoonal current sweeps northwards across the whole of China and Japan, and the summer monsoon has truly begun.

The summer monsoon of the Far East is a vast current of very warm and humid maritime air flowing northwards on the right flank of the great low-pressure systems which develop over Inner Asia as the heat of summer increases. Its northward spread is a gradual business, as has been described above, and it is not until the end of June or the beginning of July that it can be said to attain full development. By this date it has spread northward to a line running approximately across the Sino-Mongolian border and northern Manchuria. Along its northern edge it is in contact with cooler and drier continental polar air, and cyclonic developments along the frontal zone are common, bringing a brief rainy spell to these arid regions. Most of the Far East, however, lies fully within the monsoon, and sees no other airmass until the first outbreaks of the winter monsoon early in September. Over almost the whole area this season is one of quiet calm, oppressive weather of sunny days and occasional heavy thunderstorms. The only truly disturbed weather results from typhoons.

Three broad currents can be differentiated within the general mass of the monsoon (see fig. 15). All are hot, damp and prone to heavy convective rainfalls, but they differ enough to deserve separate description:

(a) *The Indian Monsoon* and its extension into China. The monsoon of India is established in a sudden and dramatic manner that has been the subject of countless numbers of qualitative descriptions. Its source is the south Indian Ocean trade winds, which in early June begin to flow across the equator, chiefly west of longitude 80°E., turning towards India some 5°N. Arriving over the west coast in mid-June, the monsoon crosses the Deccan and rapidly extends to all parts of India, Burma and the Bay of Bengal. The highly characteristic flow pattern is sketched in fig. 15. Until September there is little departure from this pattern. From the centre of lowest pressure over the intensely hot Sind and Punjab, a trough of low-pressure extends east-south-eastwards towards Bengal. Heavy rains occur in all areas throughout the monsoon period, except in the southern Deccan, where the rain-shadow of the Western Ghats eliminates the opportunity. The rain is overwhelmingly of convective origin, the trigger actions being chiefly surface-heating and orographic uplift. The effect of the latter is well seen in the extraordinarily heavy falls experienced in the Western Ghats, the Khasi Hills, the Arakan ranges and the Himalayas, all of which face the monsoon current at a high angle. The trough along the Ganges valley, however, is sometimes the scene of more general, prolonged rainfall, often in association with shallow depressions which travel slowly westwards towards the main monsoonal low.

The Indian monsoon varies considerably in strength. When it is strong, rainfall is well above average in most parts of India, while weak monsoon brings a general lack of rain. Attempts have been made to demonstrate a connection between these surges in the monsoon and penetrations of polar air into the trade winds of the southern hemisphere. Cold fronts have even been traced all the way from the south Indian Ocean to the Bombay coast, and it has been claimed that their arrival is marked by a considerable strengthening in the monsoon itself. Similar claims are made for East Africa.

The northward penetration of the monsoon is effectively limited by the Himalayas, but there is fairly continuous flow across Burma and Thailand into Indo-China and South China. The high hills running down the axis of this peninsula (the notorious "Hump" on the India-Yunnan air route) are covered by heavy convection cloud and rain is frequent. Most of this large area gets weather very like that of peninsular India. Over the Gulf of Tong-king, however, the form of the ground deforms the current, leading to the formation of a small, semi-permanent lee-depression (see p. 82). The convergence associated with this system leads to markedly heavier rains around the shore of the Gulf. In China itself the Indian Ocean air competes with air from other sources; in general it brings fine, hot weather with only scattered thundershowers, chiefly in the hills.

(b) *The Australian Monsoon.* The second general current contributing to the monsoon is the stream of tropical continental air emerging from the Australian deserts. The passage of sub-tropical highs across Australia is resumed as temperatures fall in the southern autumn, and south-east winds become general over the north coast. This S.E. monsoon travels north-westwards across Indonesia, bringing a pronounced dry season to the southern and south-eastern islands. After travelling across the warm equatorial seas, however, it becomes moist at low levels and unstable, leading to a considerable rainfall over the rest of the archipelago and over Malaya. Turning north-eastwards, it penetrates into China, over which it is the dominant airmass during most of the summer. Moister and less stable than the Indian Ocean current, it gives a heavy orographic rainfall to the coastal mountains, and an appreciable fall in all central and southern China. Weather over the Eastern Sea is similar.

(c) *The Pacific S.E. Monsoon.* The third and last member of the monsoonal current is the S.E. Monsoon of the Pacific, which affects the China Seas and Japan rather than the Asiatic mainland. During the northern summer the sub-tropical anticyclone over the Pacific achieves a high degree of permanence both in time and in position, being centred in about latitude 35°–40°N., 150°W. Around its western extremity a broad stream of maritime tropical air moves northwards. This re-curving trade wind constitutes the right-hand member of the summer monsoon over the China Seas. Though in surface properties this mT air closely resembles the other two currents, it is usually dry aloft and is quite often fairly stable. G. S. P. Heywood (1941) states that it gives clear, fair weather on its rare westward penetrations to Hong Kong, and there is much reason to believe that this is true of most other areas. Like the equatorial currents, however, the air is potentially unstable, so that it loses its fine weather characteristics when forced to rise. In hilly Japan, for example, it often gives heavy thunderstorms, most of Japan's summer rains being derived from the airmass.

The Intertropical Front and Typhoons. The foregoing account must have given the impression that summer weather over the whole of the eastern monsoon lands was made up of a long succession of fair, sunny days with high humidities and temperatures, varied only by an occasional thunderstorm. This is in fact a fair summary of typical conditions in most areas. But it neglects the drastic changes which may occur if tropical revolving storms develop within the warm, moist air of summer. Such storms occur both in Indian and Far Eastern waters, and in the latter area constitute an extremely important element in the summer and autumn climate.

Since two of the three main airstreams of the summer monsoon come from the trade-wind belt of the southern hemisphere, it is obvious that we cannot in this region distinguish a true doldrum belt; furthermore, the intertropical front must plainly lie well within the northern hemisphere. In India the northern limit of penetration of southern hemisphere air is the Himalayas, but in China and the eastern seas such air streams much further north; the extra-tropical frontal zone stretching across Manchuria is in fact in many instances its northern limit.

There is one intertropical boundary within the monsoon itself. The Australian monsoon and the Pacific S.E. Monsoon meet along a boundary zone running roughly from the Marianas towards Formosa, and thence towards the Japan Sea. Moreover, this boundary zone is a clearly marked trough of low-pressure on many occasions, at any rate from Formosa to the Marianas. Weather within this trough resembles in many ways that of the Doldrums, into which it merges east of the Marianas.

This trough is notorious as the breeding ground for most of the typhoons affecting the Far East. Some 70 per cent of all typhoons originate in the trough within 500 miles of the Marianas. A much smaller number forms over the south China Sea between the Philippines and the coast of Indo-China, usually when the trough is displaced well south of its characteristic position. The typhoons of the Marianas group travel north-westwards towards the Philippines, S.E. China, the Ryukyu Islands and Formosa, all of which experience occasional direct passages. A majority of the storms turn northwards and then north-eastwards towards Japan, those which do not affect south China (chiefly coastal areas). The south China Sea group affects Indo-China and south-west China. The total number of typhoons per annum is 20–30. Very few storms penetrate far inland in any region, and the following account refers mainly to coastal China and the islands off east Asia.

Typhoons are quite typical tropical revolving storms, and as such they have been fully described in Chapter IX. Despite their great intensity, their smallness reduces the menace they offer to the eastern lands and seas. Climatologically their significance is, however, considerable, for widespread rain occurs not only within the typhoon itself but over a wide area in the broad low-pressure area surrounding the storm. In the subtropical and extra-tropical belts the typhoon habitually occurs as the intense core of a much larger and less violent non-frontal depression, within whose circulation rain-showers and thunderstorms are common. Within the belts covered by typhoon tracks, it is not very often that the destructively violent typhoon centre will pass across any given point; yet rain will fall on many occasions from nearby storms, and a large fraction of the total summer rainfall originates in this way. As a random case, we may choose 1929 in Naha, the capital town of Okinawa, notorious as the scene of much bitter fighting in the recent war. In that year, rain fell in Naha on nine occasions from typhoons or the associated larger depressions, the first case occurring on June 28th–30th, and the last on October 23rd–25th. Only one storm passed over Naha itself; on September 28th the centre passed directly across the town, causing damage and casualties, which resulted from the strong winds (gusts to over 80 m.p.h.) and heavy rain (7.4 in. in about fourteen hours). It was four years before another storm came as close. The nine storms gave 72 per cent of the total rainfall of the period June 15th–November 1st.

The typhoon season extends from June into November, though no month is entirely free from them. September is the peak season, and is the wettest month of the year in southern Japan, which suffers greatly from these storms.

The cyclones of the Arabian Sea and the Bay of Bengal, though sometimes severe, are less frequent and climatologically significant than the typhoons.

Conclusion. We cannot leave this rather lengthy account without adding a few generalities.

Because of the extreme contrasts between their seasons, monsoonal climates offer a very specialized stimulus to human settlement; in most areas the existence of a pronounced dry season imposes on the population the need for intensive use of the abundant rains of the summer monsoon. So many studies have been made of the human geography of the monsoon lands that the character of monsoonal climates has tended to be over-simplified. The classical picture of prolonged drought, followed by the husbanding of the summer rains for the purposes of irrigation, is true of a large part of India and of the extreme north of China, but it is untrue of the rest of China and of Japan. In these areas the heaviest rains come not with the summer monsoon but ahead of it in the Mai-U, the belt of frontal rains spoken of in the previous pages. Similarly, Tong-King and Annam get their heaviest rains in the autumn, when the first bursts of the winter monsoon arrive from the south China Sea. While seasonal contrasts are the invariable rule, then, the character of these contrasts varies widely from place to place.

CHAPTER XIII

EUROPE (EXCEPT FOR THE MEDITERRANEAN LANDS)

CLIMATICALLY Europe can be divided into two sharply contrasted and very unequal parts. South of the main Alpine divide—the line of the Cantabrians, Pyrenees, Alps, Dinarics and Balkan Mts.—the climates are all of the familiar Mediterranean type, with the universal heat and drought of summer which place so emphatic a stamp upon the landscape. The rest of the continent as far east as the Urals lies wholly within a single dynamic province, the region of predominantly Atlantic airmasses and frontal disturbances. Within this large region there are, of course, some very obvious variations in temperature, rainfall and the other elements of climate; the same is true of all climatic regions based upon dynamical as distinct from static definitions.

The Distribution of Land and Sea

The map of Europe is presumably familiar to all readers of this book, and there can be no point in listing details of shorelines, hill masses and vegetation zones. We shall be content here with a few comments on the special significance of some of the more important relief features:

(i) The effect of distance from the sea. Europe is poised between the largest continental land mass and a great ocean; hence her climates exhibit a steady increase in "continentality", that is, of contrast between seasons, towards the east. In midwinter mean temperatures decrease eastward rather than northward; the outer Hebrides, for example, are substantially warmer than Paris, Berlin and Venice in January. Summer temperatures similarly increase south-eastwards rather than southwards: temperature rises slowly eastward as one travels along each parallel of latitude. We may note that "maritime" influences in inland Europe are largely Atlantic influences; the Mediterranean has singularly little effect on winter temperatures beyond the line of Alpine ranges. The Balkans, for example, have a very bleak winter in spite of their closeness to the Mediterranean. The dynamical background of the continentality is dual in character. Atlantic airmasses—mP of various types—undergo progressive winter refrigeration and summer heating as they travel eastwards; moreover, they are often replaced in eastern Europe by cP air of Siberian origin. The degree of continentality hence depends both on the degree of modification of mP air and the relative frequency of mP and cP air.

(ii) The effect of the Alpine mountain barrier. It is impossible to exaggerate the importance of the high mountain barriers of the Alpine fold belt. Attention has already been drawn to the fact that these ranges separate the Mediterranean from the European climate proper. They also have a striking effect on the climates themselves. It is by no means common for moist, warm airmasses from the Mediterranean to penetrate into Europe north of the divide; it takes a major disturbance to draw such air across the mountains, so that invasions are normally brief and are abruptly terminated as the disturbance passes away. The consequences of these rare incursions are discussed later in the chapter. Another significant effect of the hills is in arresting the flow of cool air south-eastwards in winter. Invasions of mP air behind the cold fronts of Atlantic depressions are brought to an abrupt halt at the foot of the Pyrenees and Alps, with remarkable results on the climate of both central Europe and the Mediterranean. These effects are more fully treated later. But the outstanding result of the general east to west run of Europe's major climatic divide is that Atlantic airmasses, fronts and depressions can penetrate with little hindrance right into the heart of Russia. The contrast with North America is hence complete: in that continent Pacific influences are reduced to small proportions by the presence of the Western Cordillera, whereas warm and moist mT air from the Gulf of Mexico is free to spread deeply into the continental interior; Europe by contrast is wide open to the ocean to the west, but largely sealed off from the warm sea to the south.

The Airmasses of Europe

The internationally accepted airmass classification of T. Bergeron (see pp. 57-64) was conceived against the background of the European climates, so that it might be expected that it would prove highly satisfactory in description. The most recent detailed study of the properties of airmasses affecting the British Isles was prepared by J. E. Belasco (1945) for the winter months, and much of the succeeding account is based on his invaluable work.

Maritime Polar (mP) air is by far the commonest type in north-west Europe at all seasons, and it is common in all parts of the continent for most of the year. In practice we need to subdivide the type considerably, for its properties vary widely. We can identify the following broad types:

(i) mP air of direct northerly track is the coldest variety. This is the airmass which sweeps southward across the country during northerly spells, when pressure is very high from Azores to Iceland or even further north, and low over the Baltic. This weather type is common in winter and spring. The source region for such air is usually the Greenland or Barents Sea, but the author can remember spells when it came from the polar basin itself, beyond Spitsbergen—hence the term "Arctic" Air which is sometimes applied to this type. This airmass is outstandingly unstable: lapse rates may be almost equal to the dry adiabatic up to remarkable heights. Over the warm seas around Britain heavy cumulo-nimbus and ragged scud cloud develops, and there are frequent wintry showers. This cloudy, showery weather also affects coastlands exposed to the north. North Scotland, East Anglia and the Low Countries are especially prone to such bad weather; indeed the dense cumuliform cloud from the North Sea may penetrate into central Germany if the winds are strong. Inland over Britain this type of mP air is inclined to be clear and frosty at night, but showery by day during the period of solar heating.

(ii) mP air of north-westerly track is a familiar perennial in all parts of Europe. This is the air which follows the cold fronts and occlusions of Atlantic depressions, giving us the April shower type of weather, typical of many months besides April. Such air is very often old cP air from Canada or Greenland which has totally changed its properties in its long ocean journey. According to the complete Bergeron classification, this air is mPK, fairly moist at the surface, and moderately unstable at all heights, with lapse rates near the saturated adiabatic. As it moves eastwards in winter towards cold eastern Europe, the air changes to mPW in character. Over the seas there is abundant cumuliform cloud with showers of rain or hail, and this kind of weather is intensified over the hills of Scandinavia, highland Britain and north-west Spain, all of which have quite a considerable amount of rain from such air. In the lower lands, however, inland weather is clear by night, and bright, cool and showery by day. As the air passes into central and eastern Europe slowly diminishing humidities reduce the showers to daytime cumulus cloud. There is little seasonal variation in the properties of this airmass, which in Britain usually gives temperatures a little below average.

(iii) mPW air from west or south-west is the most variable of these sub-types. This is air which has made a fairly deep incursion southward into the Atlantic, arriving over western Europe warm and superficially stable, with considerable moisture at low levels. The majority of depressions which cross Britain or Scandinavia have mPW air in their warm sectors; the storms originally formed on fronts separating such air from mPK air. mPW air is stable relative to the sea temperatures around Britain, and thus appears over our coasts laden with stratus or strato-cumulus layers, giving the grey skies and drizzle so typical of the cooler season. Considerable rain may fall from this airmass if it is compelled to rise over large hill masses, but as a rain-bringer mPW air is at its best along the warm and cold fronts of eastward-moving Atlantic depressions. In summer the heat of the sun may disperse the surface stability and the accompanying stratiform cloud in southern and eastern parts of Britain, and fairly generally over inland Europe. Widely scattered showers and thunderstorms are not uncommon by day at such times.

Maritime tropical (mT) air is much less common than mP over Britain and western Europe though some of the mPW airmasses mentioned above may have originated as mT air which curved round the western end of the Azores-Bermuda high. True mT air reaches western Europe chiefly in the warm sectors of depressions forming much further south than usual, often near or south of the Azores. It gives warm, humid weather, usually cloudy and drizzly over coastal areas because of the cooling effect of its passage across cool waters. Though very moist, mT air is second to mPW air as a rain-bringer, for it is usually fairly stable. In this respect it contrasts strikingly with mT air over the eastern U.S.A. and Canada, where it is usually potentially unstable; European mT air is mTWs in Willett's complete classification (p. 61).

A significant variant of Atlantic mT air is warm, moist air from the Mediterranean, which occasionally penetrates northwards into Europe. Such air is warm and moist at the surface, but tends to be cool aloft, and is hence inclined towards instability. Such invasions take place along two main paths: (i) northwards across Provence and Languedoc into France and Britain, and (ii) northwards and north-eastwards into Danubia and central Europe across the lower northern Dinarics. In both cases the air always forms part of the warm sector of travelling depressions, and widespread heavy rain (or snow in winter) is usual. Quite often in winter the warm air will override cold cP air over the Danube lands, giving overcast skies and precipitation without any appreciable rise of temperature at low levels. This is the source of much of the winter precipitation of S.E. Europe. It must be borne in mind that these airstreams usually consist of Atlantic mP air that has been warmed up over the Mediterranean. Sometimes, however, cT air from the Sahara is drawn into Europe. Though very dry aloft, such air is moist below after its sea crossing, and precipitation may occur from it. The instances of "blood rain" which have been described from many parts of Europe, including Great Britain, are due to Saharan dust which is washed down by the rain.

Though penetrations of Mediterranean air chiefly follow the paths indicated, it is not uncommon for such air to be drawn across the Alps and to descend on the Foreland of Switzerland very much warmed and dried. This warm dry current is the celebrated Föhn; the name is now accepted internationally for warm dry currents descending from mountain ranges. In the Alpine region, a distinction is drawn between the Südföhn, the southerly wind affecting the north flank and Foreland, and the Nordföhn, which affects the Italian Alps, the Ticino and the Po Valley. The latter is discussed under Mediterranean climates in the next chapter. Like the American Chinook, the Südföhn produces some striking rises of temperature when it arrives. Dense cloud and rain occur on the south flank, but on the north side skies are brilliantly clear, and temperatures even in midwinter may rise into the 50–60° range after having hovered about the 20° mark for days. In the Alpine valleys the föhn may achieve destructively violent speeds; it also produces rapid thaws and floods, so is by no means as welcome as the good weather may suggest. North of the Alps it brings clear weather to the Foreland and a part of south Germany, but rarely retains its warmth much beyond the Danube. The Südföhn occurs at all times of year when depressions move north-east along the north-west seaboard of Europe, but especially when the centres pass over England, France or the Rhineland. There is a pronounced spring maximum.

Radically different in type from the airmasses so far described are airmasses reaching Europe from the east, over Russia. Such airstreams include the vast extrusions of cP air from Siberia which bring the year's coldest weather to western and central Europe, but they also include very warm and dry air currents in the summer which approach, though do not truly attain, the properties of cT air.

Continental Polar air (cPK) sweeps westwards across Europe at intervals in every winter and spring. Eastern Europe sees a great deal of this air, but it occasionally sweeps out into the Azores region of the Atlantic across France and Britain. The circulation type favouring such invasions is an anticyclone over northern Europe, the Baltic or the Russian Taiga. Very often such highs are warm anticyclones which form as north-eastward extensions of the semi-permanent Azores anticyclone; the latter then gives way, and pressure becomes low in the

Azores region, creating a gradient for easterly winds over the whole of western Europe. The properties of this air depend largely on its ultimate source. Where the air is derived from the Atlantic and adjacent seas north-west and north of Scandinavia, there is widespread strato-cumulus over much of Europe, as the easterlies tend to be stable and superficially rather moist. If the anticyclone is merely a great westward ridge of the central Siberian high, however, the easterly current is deep-seated cPK air from the Russian steppes, and it is bitterly cold, dry and cloudless over the European mainland. Britain, however, gets such air after a crossing of the relatively warm North Sea and the Straits of Dover. Strato-cumulus cloud is usually unbroken over all eastern and southern parts of the country, and light wintry showers are general. The cold of February 1947 with its unbroken, gloomy pall of cloud was of this type.

A winter variant which needs attention occurs when a ridge of high-pressure develops across central Europe in mP air behind an eastward-moving cold front, a very common winter development. Within such ridges conditions are quiet and very stable. Surface temperatures are low, the lowest layers of air lying inertly over the various hollows and basins of central Europe. In such conditions there is usually stratus or strato-cumulus cloud in most parts of the continent; fog occurs in areas where the cloud is broken.

In summer conditions are entirely different; central and eastern Europe are then very warm. It is still quite common for anticyclones to be over the Baltic or north Germany, allowing a broad stream of easterlies to move across central Europe to France and Great Britain. In such circumstances, weather is hot and usually nearly cloudless over wide areas. The air tends to be dry and moderately stable. Even in England, after the crossing of the North Sea the air is warm and dry: except along the east coast temperatures rise high into the 80°s, and relative humidity has been known to fall below 20 per cent. Over the North Sea, however, such air is chilled and moistened from below, and advection fog or stratus forms, drifting across our east coast and penetrating deeply inland overnight. Next day the inland cloud disperses, but the coast itself remains blanketed on many occasions, especially north of the Wash.

Main Features of the Circulation

Europe lies wholly within the Atlantic province from the point of view of its circulation; both long-term and day-to-day changes in its weather respond to dynamical controls traceable back to the Atlantic area. Certain things about the general circulation must hence be made clear:

(i) The average pressure distribution shows much the same pattern throughout the year. A large anticyclone appears between the Azores and Bermuda, the centre lying a little further north in summer than in winter. A ridge of high-pressure extends north-eastwards across France into central Europe. In winter this extends to the Siberian high, but in summer terminates over Germany. Pressure is low from the Davis Strait across southern Greenland to Iceland, lowest pressure being usually a little south-west of Iceland itself. A trough of low-pressure extends east-north-eastwards beyond northern Norway to Novaya Zemlya and beyond. This great low-pressure belt, the "Icelandic Low" of the meteorologist, varies in intensity from a maximum in midwinter to a decided minimum in July. Pressure is also relatively low in the winter half-year over the Mediterranean and high over the western Sahara.

(ii) The mean circulation derivable from these distributions also has well-defined characteristics. Between the axis of the Azores—central Europe high-pressure ridge and the long low-pressure trough running from Iceland to Novaya Zemlya, the general flow is from a little south of west, so that moist Atlantic air is very often carried deep into the continent at all seasons. This current is heterogeneous in properties, varying from mPK to mT in character. Numerous fronts are carried eastwards in the current, which is hence characterized by changeable weather. North of the Icelandic trough winds are north-easterly, often coming from the pole and beyond. Finally, south of the central European ridge there is a tendency for north-easterly circulation along the north shore of the Mediterranean.

It must be stressed that this average pattern, though it resembles many actual day-to-day circulation types, is very often broken up. So complex are these changes it is almost impossible to summarize them rationally. An attempt is made below under two headings: (a) displacements of the Azores anticyclone, and (b) the travel of cyclonic depressions.

Displacement of the Azores Anticyclone. The large area of high-pressure appearing in the sub-tropical Atlantic between the Azores and Bermuda reflects the fact that stationary warm anticyclones do actually lie over this belt for a large part of the year. The westward extension of the high-pressure which appears on the winter maps represent the paths of migratory cold highs moving south-eastwards from North America into this belt. So nearly permanent is the high-pressure that meteorologists speak of the "Azores anticyclone" in the full confidence that their listeners will know what they mean without further qualification. But this high degree of permanence must not blind us to the occasional displacement of the system, for these have a profound influence on the European climate. We can summarize these effects in this way:

(i) Azores anticyclone more intense and further north than usual—especially common in summer. The chief effect of such a displacement is to drive Atlantic storms very far north, so that western Europe experiences only the trailing southern ends of their fronts. This leads to cloudy, mild weather in winter, largely free of rain and to fairly warm, settled weather in summer.

(ii) Azores anticyclone further south or south-west than usual—especially common in autumn and winter. This permits southern track for Atlantic depressions, which may penetrate deeply into Europe, giving spells of very disturbed weather.

(iii) Azores anticyclone extending northwards to the Iceland and Greenland region. This displacement gives the "northerly type" over Europe, bringing a continuous stream of mPK air down from north-east of Iceland. Cool and showery weather is general over most of Europe. This type is uncommon in summer.

(iv) Azores anticyclone displaced north-eastwards. It is very common for large warm anticyclones to be centred over some part of Europe, very often over Great Britain or Scandinavia. At such times there is often relatively low-pressure over the sub-tropical Atlantic, and it is plain that the European system is effectively replacing the Azores high in the general circulation. Sometimes the Azores system actually migrates north-eastwards, but more often there is fresh anticyclogenesis over Europe, usually in mP air. Such anticyclonic weather is familiar to all of us. Foggy and gloomy in their central regions in winter, brilliantly fine and warm in summer, the slow-moving, persistent highs effectively break the monotony of the westerly spells with which they alternate. When they lie over Scandinavia, they give rise to the easterly type of circulation described above under cP air.

In short, we can say that displacements or abnormal behaviour of the Azores anticyclone give us our "spells" of weather—the periods of days or weeks in which the weather tends persistently to follow certain types of development. This is the phenomenon called "persistency of type".

Travelling Depressions: Stormy Weather. Just as displacements of the chief anticyclonic belt control the general weather type experienced over Europe in a given period, so are daily changes in weather—and particularly of stormy weather—largely determined by the movement of depressions and associated frontal systems across the continent. We can distinguish three broad groups of disturbance: Atlantic depressions, Mediterranean depressions, and the shallow but often rainy disturbances forming over the continent, usually in summer.

Atlantic depressions are hardy perennials of our own climate, and at intervals they affect every part of Europe. Even when the centre of a travelling storm passes wholly north-west or north of the continent, its fronts (especially the cold front) may penetrate deeply into the interior, sometimes even reaching the Ukraine and the Urals.

These oceanic storms are frontal developments associated with what was called the Atlantic Polar Front in Chapter XI.

When the latter lies off the American coast, the young frontal waves involve mT air in their warm sectors. They travel north-eastwards towards Iceland, which they normally reach in their most intense state, fully occluded, and covering a vast area of the north Atlantic with strong winds or gales. In the Iceland region they tend to slow up, but they very often maintain their intensity for several days. The occlusion travels eastwards across Britain and western Europe as a cold-type occlusion, giving a period of cloudy, rainy weather, followed by a veer of wind and a clearing of the sky as fresh mP air sweeps in. Occasionally both warm and cold fronts are recognizable, but in the majority of instances this class of cyclone occludes fully before its fronts strike the European coast.

This Icelandic type is by far the commonest source of strong winds and of rainy weather in the British Isles, and the rest of western Europe. It occurs at all seasons, but is most intense and most frequent in the cooler months. It is to be regarded as the "normal" storm type of the ocean, if normals can be identified in so complex a series of patterns. The remaining Atlantic storms can best be treated under three headings:

(a) When the occlusion or trailing cold front of the Icelandic type of Atlantic storm has penetrated to western Europe, it very often tends to become stationary along a line running south-westwards towards Bermuda or the Azores. Fresh waves may form on this quasi-stationary front, and travel rapidly north-eastwards towards Europe. These are the storms which actually penetrate the continent, or pass immediately along the north-west seaboard. They are more youthful than the foregoing "normal" type; when they pass or cross Great Britain, both warm and cold fronts are often identifiable, a warm sector of mPW or mT air travelling north-eastwards across both Britain, France and the Low Countries. They normally occlude as they approach Scandinavia or Germany, according to the path they choose. Such storms, though less frequent than the Icelandic type, are often quite intense, and may give widespread precipitation to all parts of Europe, for their fronts penetrate deeply inland even if their centres pass north-eastwards along the coast. C. K. M. Douglas and A. J. Glasspoole (1947) have recently shown that nearly all the heavy orographic rainfall so characteristic of the hilly areas of Britain (and presumably of other N.W. European countries) falls in the warm sectors of these storms, within 300–400 miles of the centres. It cannot be too strongly emphasized that the hills alone do not produce heavy orographic rainfall in moist airmasses; they merely intensify frontal rainfall, or lead to heavy rain in warm sectors. At other times rain in the hills rarely exceeds a thick drizzle (Scotch mist) or convection showers, however moist the air.

(b) Either Icelandic storms or the type just described may travel far enough north-eastwards to enter the zone of the so-called "Arctic front"—more properly termed the Arctic frontogenetic zone. This is the N.E.-S.W. belt extending from Novaya Zemlya and Spitzbergen towards Iceland. This area is very favourable for frontogenesis when warm south-westerly winds blow across Scandinavia and come into contact with truly Arctic air from the permanent ice-pack of the Arctic Ocean. Hence when depressions move north-eastwards past North Cape they are often able to revive themselves by absorbing an entirely new frontal system. In this way rejuvenated Atlantic storms may penetrate far into the Siberian Arctic, and very often into northern Russia west of the Urals. Considerable rain (or snow in the cool seasons) accompanies these storms in northern Europe. In autumn and spring it is not uncommon for cyclones involving the Arctic front to penetrate to the Pacific over eastern Siberia in much reduced guise.

(c) The third type of storm is much less common, and may be regarded as essentially a type of late winter and spring; it accompanies the so-called "northerly type" over western Europe, in which pressure is high between Iceland and the Azores. In such cases there is a frontogenetic field over northern and eastern Europe between the occasional penetrations of relatively warm air coming round the northern end of the Atlantic high and the stream of cold cP air which habitually covers Russia at such times. Wave cyclones forming over the Greenland, Norwegian or even the Barents Sea travel south-eastwards, typically over the Baltic. Though these storms do not in general give very heavy precipitation they bring a spell of unusually stormy weather to the eastern part of the continent.

Mention has already been made (p. 82) of the so-called "polar air" depressions of non-frontal type which occasionally affect northern Europe. No further notice is necessary here.

The second broad general class of cyclonic disturbance is of much less significance, though not of interest. *Depressions of Mediterranean origin* are very much less common than the Atlantic type over most of Europe. They occur chiefly in two areas, the Gulf of Lions and the Adriatic. The former develop as waves on the cold fronts of Atlantic depressions that have reached the Gulf, and then been retarded. Such waves—which are by no means common—may move north across France, Britain or the Low Countries. Rain, sometimes of a thundery type, is widespread west of their path, but as a rule they do not attain any great size and intensity. The second type occurs several times a year, though never in summer. When there is an anticyclone over the Balkans warm southerly airstreams—the scirocco of the Italians—pour towards the head of the Adriatic. At such times there may be cyclogenesis over the Alpine foreland north of the Alps, or over northern Italy. Sometimes a depression may even move northwards from the Mediterranean. Whatever the origin, the effect is the same; the slow northward movement across central Europe and Germany of a shallow cyclonic centre with warm, moist Mediterranean air on its eastern flanks.

The third and final class of storm is the type normally referred to as "continental depressions". These are of various types. In summer it is common for shallow, slow-moving, complex disturbances to drift across the continent, accompanied by widespread thundery rains. The origin of these storms is often obscure, as are their frontal characteristics; at times they may move northwards or north-westwards across Britain, giving us our spells of sultry, thundery weather. Much of central and eastern Europe's summer rainfall maximum comes from such storms. More significant in the cooler seasons are the wave developments on Atlantic cold fronts or occlusions. The latter tend to move quite slowly across Europe, and it is very common for stable waves to develop upon them, especially as they move across the low hill masses and plateaus which diversify the surface of central Europe. Fairly prolonged precipitation and cloudiness occur in the path of these feeble systems, which rarely or never develop into mature cyclones. They are none the less the dominant source of cool-season precipitation in all but the north-western parts of Europe.

The Seasons

We are now able to attempt a synthesis on a seasonal basis. The seasons of Europe's climates are very far from resembling the clear-cut periods of the monsoon lands or of North America, but they are none the less distinct to us. To some extent, of course, our assurance that autumn is a season distinct from spring rests upon non-climatic differences: obviously it is difficult to compare a sunny spring day, with its carpet of brilliant green, its daffodils and bluebells, the young leaves on the trees, with its equivalent in the autumn, when the ground is drab in hue and the faint, pungent odour of dying vegetation is all around us. Yet climatically there is very little difference between the seasons: a greater raininess in autumn, it is true, but falling from the same type of frontal storm which affects us in spring. In short, we have to be careful, in our assessment of climatic seasons, that our differentiation really does rest on climatic factors. The brief summary that follows tries to bring out the truly climatic differences between the seasons.

The winter, which we can perhaps identify with the months December, January and February, is a season of great contrast between east and west. The east is bitterly cold, and the ground is covered thinly by snow. The rivers freeze over, and life is generally adapted to severe and continuous cold. This cold arises from the frequency of the cP airmass, or of clear skies in old, dry mP air from the Atlantic, which allows rapid nocturnal cooling. The gales and stormy weather characteristic of the west penetrate only rarely, and thaws are rare. The west, on the other hand, comprising the Atlantic seaboard states, experiences the turbulent weather characteristic of the Atlantic. There is considerable rainfall, much low cloud, and weather is kept mild by the frequency of maritime airmasses. Between these two extremes there is a central transition zone stretching from the Alps and Carpathians to the Arctic Ocean. Over Germany this is an area in which the eastward fall of temperature is gradual and continuous, but in Scandinavia the mountainous backbone of the country sharply separates maritime from continental climate.

Spring (March, April and May) in the west develops slowly out of winter with no pronounced changes in the behaviour of the weather; the Atlantic depressions decrease in vigour, each airmass slowly warms up, and almost imperceptibly spring—the visible spring of field, garden and hedgerow—arrives. Yet spring is a season in which the types of circulation giving unusually cold weather over western Europe tend to recur all too often. Prolonged spells of cold east winds are common throughout; in parts of southern England they are as common as the westerlies in April and May. Northerly type circulation is also common: snow has fallen in central and southern England as late as mid-May in these situations. All in all, then, spring climatically falls far short of its own natural beauties in the landscape. In the east, too, it is the season of the agonizingly slow, slushy thaw, the period of raw cold, of fogs and of liquid mud that annually halted military operations in the past war. May, however, brings a hot, strong sun, and the east rapidly warms up until it is even warmer than the more slowly changing west.

Summer (June, July, August) reverses the winter story. It is now the east that enjoys the greater warmth. All through the central and eastern parts of Europe summer is a reliable season of pleasantly warm, sunny weather. The occasional spells of rain may be fairly heavy—for this is the continental rainy season—but they are habitually of the erratic thundery type which dampens the spirit much less effectively than the clinging rains of the west. The west of Europe, on the other hand, never knows what to expect in this most erratic of all the seasons. Sometimes an unbroken succession of Atlantic depressions moves past or across the seaboard lands, and the summer is cool, wet and cloudy: years go by in which there is no real summer, as did 1946 over so much of western Europe. In other years persistent high-pressure over the continent may repel all Atlantic disturbances, and fine warm weather persists for weeks. Even though 1946 dampened our spirits so effectively, 1947 gave us a golden summer and early autumn almost without parallel. The moodiness of summer rests upon the persistence of type referred to earlier in the chapter: displacement of the normal position of persistent high-pressure may radically vary the prevailing weather.

Lastly, the *autumn* (September to November) develops out of summer, usually by the arrival early in September of an unusually vigorous outbreak of mP air behind the cold front of an Atlantic storm. Over the ocean itself there is a remarkable increase in energy of circulation during September: the equinoctial gales are no myth. Autumn is hence a season of increasing storminess in the west: it is the wettest season in many lowland parts. It is also traditionally the season in which radiation fogs appear again, and in November they can be as widespread and as thick as in midwinter. Further east the fall of temperature is much more rapid, and by the end of the season the winter frost has begun. Then, once again, the rhythm is repeated, and another year goes by.

CHAPTER XIV

THE MEDITERRANEAN LANDS

FROM secondary school days onwards, every student of geography has ceaselessly had drummed into him the main features of the Mediterranean climate. This climate fascinates the geographer because of the extent to which it has called for specialized economies and specialized ways of living, both today and in the past. This was the environment of Periclean Athens, of Augustan Rome, of the Israelites and later of early Christendom. Today it is associated in our minds with certain types of sub-tropical fruits and with a poverty-stricken peasantry living on eroded hillsides in the perpetual fear of drought. To what extent the beauty, warmth and tranquillity of so much of the year's weather stimulated the early growth of civilization along the shores of the eastern Mediterranean is a matter for controversy. Some of the most extreme determinist claims seem absurd, though we cannot on this ground dismiss the idea entirely. But we can be sure that the present-day character of occupance of the Mediterranean rural landscape rests fairly and squarely on climatic control; there are few environments which so severely restrict the livelihood of peasant communities, or call so loudly for the skill of the soil conservationist and the irrigation engineer.

The Mediterranean climates are by no means restricted to Europe. In that continent they are confined to the narrow strip of coast south of the Alpine-Dinaric divide, to the peninsulas of Iberia, Italy and Greece, and to the islands scattered along the sea from the Balearics to Crete and Cyprus. Asia Minor and the Levant belong in this province, which grades into the central Asiatic deserts in the plateaus of Iran. In Africa, the Atlas province and Cyrenaica have a thoroughly Mediterranean climate, but in Tripolitania and Egypt a very narrow coastal strip alone has the appreciable winter rainfall characteristic of the type.

This climate is characterized by a dry and wet season régime as emphatic as anything in the Monsoon lands. The summer is everywhere hot, sunny and almost rainless, the winter mild and rather wet. Similar rainfall régimes occur in restricted parts of other continents, notably in California (see pp. 136-137), Chile, West and South Australia and South Africa.

General Characteristics of the Circulation. The peculiar type of rainfall régime which more or less defines the Mediterranean climatic belt rests upon two major dynamical considerations:

(i) Lying as it does between latitudes 30° and 45°N., the Mediterranean is in a position in which it might be expected to be within the westerly belt in winter and along the axis of the sub-tropical high-pressure belt in summer. This is the stock explanation of the climate, and it does contain a large measure of truth. Though rarely the scene of actual anticyclonic development in summer, the Mediterranean is usually covered by fairly uniform pressure a little above the average value. Moreover the air, though moist at the surface, is dry and stable aloft. Over the eastern Mediterranean, however, a gentle pressure gradient towards the east is common, giving the persistent northerlies and north-westerlies—the "Etesian" winds—so characteristic of the region. In winter, on the other hand, the Basin is covered by a succession of depressions travelling from the west, and the prevailing winds over much of the Basin are westerly, pressure being on the average high over the Sahara. Qualitatively, then, it is true to say that the Mediterranean enjoys a sub-tropical desert climate in summer and a disturbed westerly type climate in winter.

(ii) To this first factor must be added the fact that in the cooler seasons conditions are often favourable for frontogenesis over the Basin. Most authorities refer to it as the zone of "Mediterranean frontogenesis" running from near Genoa south-eastwards towards the Levant. The geographical factor favouring such frontogenesis is the temperature contrasts between cP and mPK airmasses over Europe proper and warm mPW air over the Mediterranean. Dynamically conditions are rarely very favourable for simple frontogenesis of the deformation type described on p. 65. If, however, there is cyclogenesis from other causes over the Basin, the frontogenetic temperature field help to give the young cyclone energy. Most cool season cyclones are formed orographically, and are not wave-cyclones in the ordinary sense.

With this general background we can look more closely at the two seasons.

The Summer Drought

Though the dates of beginning and ending of the summer drought vary from place to place, the three months June, July and August are almost rainless in nearly all parts of the Basin. In the south-eastern Mediterranean the months of May and September are also arid, but the circulation, nevertheless, has more affinities with winter than with summer.

What is the background of this remarkable spell of dry, hot sunny weather? On the face of it we might expect considerable rainfall of the thunderstorm type; the air is very moist at low levels, and the sun is very strong. The answer appears to lie in the dryness of the upper atmosphere, and in the stability of the air-column, both of which are very general. Conditions resemble those over Britain when an anticyclone lies over the country in moist, warm air. Even more significant, however, is the absence of any real source for cold air which might penetrate into the Basin and start off cyclogenesis. Europe to the north is warm in the summer months, and the chances of a polar outbreak reaching the Mediterranean in a tolerably cool condition are remote. Genoa, for example, has never recorded a temperature below 58°F. in July. The temperature contrasts associated with strong cyclonic development are hence lacking.

The outstanding characteristic of the summer circulation is in fact its sluggishness and formlessness. Strong and persistent winds are rare up to great heights. Only in the eastern Mediterranean is there to be seen any persistent circulation; here the circulation of northerly and north-westerly winds forms part of the great monsoonal circulation of Asia. The low wind speeds and clear skies gives great scope for solar heating, and heat is unbroken. The Levant particularly is extremely warm, much of Palestine, Syria and the Iraq lowlands having average July temperatures of 80°F. and over. The effect of the heat in most of the Mediterranean lands is heightened by the oppressively high humidities engendered by the great expanses of warm sea surface. The absence of cool spells adds to the strain upon man and beast.

The consequences of the summer drought are well known to all students of geography. All vegetation withers or becomes dormant; drought-resisting species alone can persist from year to year. Furthermore, overstocking of the hilly landscape, coupled with the depredations of the ubiquitous goats, has led to disastrous soil erosion; many of the hillsides become veritable deserts during the summer. Summer life is adjusted to the oppressive heat: the peoples reduce activity to a minimum, and until autumn comes there is little heavy work undertaken in rural districts.

The picture painted above is true of most parts of the Basin, but needs qualification in others. The drought is most continuous and the heat most intense in southern and eastern parts of the Basin, and most of the qualifications refer to the north and west. In these areas, nearly all of which are hilly, there are periods in which the stability of the air column is at times reduced, and the daytime sun is able to generate considerable convection. Small cumulus is very general in the north-west in the afternoons, and at times thunderstorms may break out in the hills. This is particularly true in the Pyrenees, the coastlands of France, the Apennines and Alpine foothills. Indeed the Po Valley and the lowlands round the head of the Adriatic get a very considerable summer rainfall from such storms. They occur as a rule when relatively cool air is advected to the region from the Atlantic at considerable heights. The approach of an Atlantic cold front is usually heralded over the Alps and northern Italy by pronounced cooling of the upper air, and it is this reduction in stability which sets off thunderstorm development. Even here, however, the predominant weather is the same hot, sunny type typical of the Basin at large.

The Rainy Months

The onset of the disturbed weather of the cooler season is an event almost as welcome and dramatic as the onset of the summer monsoon in India. Some time during September the cold front of an Atlantic depression will sweep south-eastwards across France and Germany with a vigour not seen since the previous May. This cold wave penetrates through the gap of Provence and sweeps out into the western Mediterranean, usually reaching the coast of Algeria in the south and Dalmatia in the east before it is rendered unidentifiable by the heating of the sea. The sudden injection of cold air into an area which is covered by warm and moist air releases enough energy for cyclone formation, and heavy rains occur over a wide area of the western basin. The summer drought and sluggish circulation are banished, and the rainy season begins. As autumn goes on, the invasions of polar air southward into the basin become more vigorous and more frequent, and the resulting cyclogenesis is correspondingly increased. By mid-October depressions from the western basin are penetrating right into the Levant, and the summer drought is everywhere at an end. From then until the following May the climate retains this rather stormy, cyclonic aspect, so strikingly different from the quiet of summer.

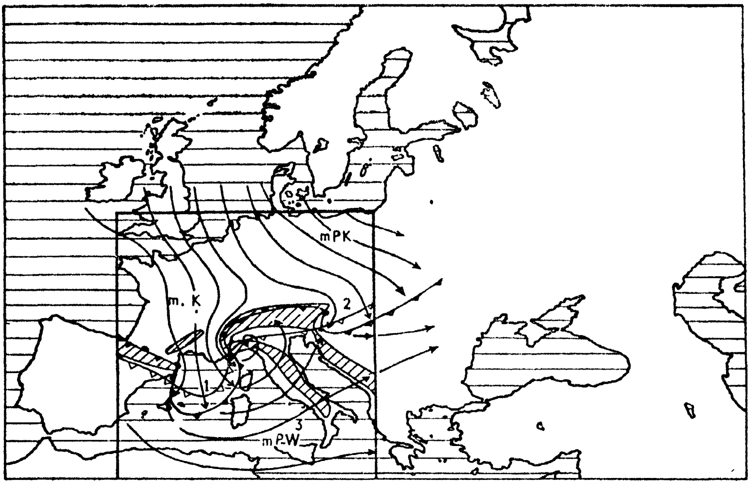
There is a great deal of confusion in climatological literature about these Mediterranean cyclones. Generally they are pictured as Atlantic storms which enter the basin through the Provence gap, across the Spanish Meseta, or through the Straits of Gibraltar. An alternative view presents them as waves on fronts formed in the frontogenetic zone over the Basin itself. Though both such sources do exist, the really important source of storms is lee-cyclogenesis behind the Alps, already discussed on p. 82. A large majority of the bigger depressions affecting the Mediterranean form over northern Italy, the Ligurian Sea or the Gulf of Lions. They form when cold fronts or occlusions of Atlantic storms are arrested by the Alps and Pyrenees. The key to the winter rainfall distribution hence rests upon two primary factors, the incidence of cold waves of Atlantic origin, and the peculiar relief of the ground along the north shore of the Basin.

A glance at fig. 16 will remind us that the Mediterranean is cut off from Europe and the Atlantic by a formidable succession of hill barriers. In the west the Spanish Meseta is a considerable obstacle, but the largest and most significant barriers are those which lie athwart cold waves from the north-west, the Cantabrian-Pyrenees line and the Alps. Between these two barriers lies the gap of Provence, split into the Rhône Valley and the Gate of Carcassonne by the Cévennes. The relatively shallow cold waves from the Atlantic find the Pyrenees and the Alps an almost impassable barrier, but the Cévennes are much less formidable; cold waves are thus able to surge forward across the Gulf of Lyons into the western Mediterranean while still arrested over Gascony and the Alpine Foreland.

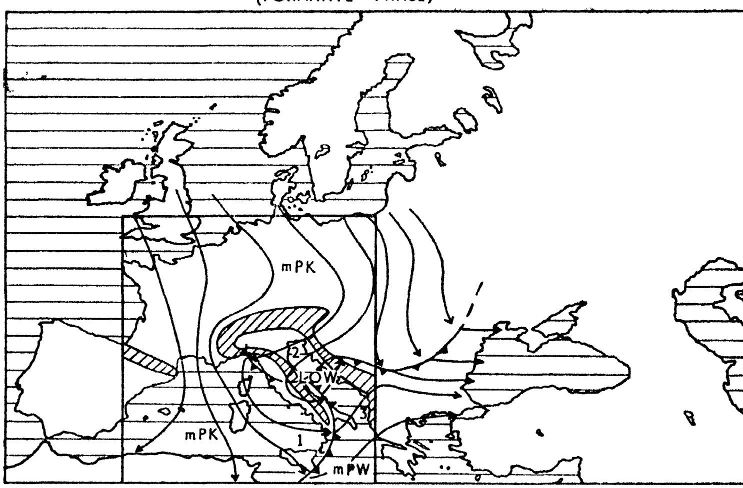
Further east, the Dinarics and the Balkan ranges and plateaus continue the high barrier separating cold airmasses from the Mediterranean. The narrow northernmost sections of the Dinarics are, however, very often surmounted by streams of polar air which pour through the range and across the Adriatic and northern Italy. The Morava-Vardar depression also allows a relatively free passage.

It should be emphasized that the effectiveness of these barriers in hindering the sweep of cold airmasses into the Mediterranean depends on the shallowness of the currents concerned; such air is normally overlain at no great height by warmer westerlies, and the stratification is hence very stable. The cold air at the surface is unable to climb over the range because of this stability, and the pressure distribution adjusts itself so as to render the climb unnecessary. The cold air is able, however, to continue unchecked through the various gaps, and does so with a considerable increase in velocity; the flow through the Provence gap, for example, amounts to a veritable jet which is injected into the Mediterranean basin. Each major gap in the barrier is hence subject to frequent and often violent passages of cold air in the cooler seasons. Among the better known we can identify the following:

(i) *The Mistral*, celebrated as the "scourge" of Provence. The name is applied in Languedoc, Provence and the Rhône Valley to the very fierce northerly wind which occurs at frequent intervals all through the winter. We can apply the term "Mistral Stream" to the broad current of generally north-westerly direction (see fig. 16) which passes through the whole of the Provence Gap.



THE FORMATION OF AN ALPINE LEE-DEPRESSION
(FORMATIVE PHASE)



THE FORMATION OF AN ALPINE LEE-DEPRESSION
(MATURE PHASE)

Figs. 16 and 17

Formative (Fig. 16 above) and mature (Fig. 17 below) stages of an Alpine lee-depression. The Mistral stream is labelled 1, the Bora stream 2 and the warm Mediterranean airstream 3.

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The locally intensified Mistral of the Rhône Valley is a very strong canalized part of this current which is often accompanied by low temperatures, clear skies and frost. It is a serious hazard to shipping, approaching Marseilles, which stands directly in its path. A similar wind, though less fierce and cold, blows from the west-north-west through the Gate of Carcassonne, the gap between the Cévennes and Pyrenees. The Mistral blows whenever a cold front arrives over the Cévennes from the Atlantic and is able to pass through to the Mediterranean.

(ii) *The Bora* is a similar wind blowing through the gap between the Alps and the high ground of the central Dinaric ranges. Severe bora winds are confined to the saddle region behind Trieste and Bora, but the intensity of the flow more than compensates for the restricted character of the gap. Bora winds in the southern Adriatic are more often of a local katabatic character.

(iii) *The Vardarac* is a smaller scale wind which blows through the Morava-Vardar gap after the southward passage of a cold front across the Balkans. It affects the Salonika region and the surrounding plain.

It has already been pointed out that a large proportion of all Mediterranean depressions forms over or near northern Italy in the lee of the Alps. These "Genoan lee-depressions", as they are often called, have no precise parallel anywhere. Their mode of formation and associated weather are described in the following paragraphs. Figs. 16–17 show the main stages in the evolution.

The initial phase, in which incipient cyclogenesis becomes apparent, is best described as follows. A cold front has reached the foothills of the Pyrenees and Alps, and behind it a strong current of cool, fresh mP air is sweeping south-eastwards. At this stage the general fall of pressure which precedes any rapid-moving cold front is observed to be intensifying over northern Italy, and a small trough of low-pressure or closed centre of low-pressure, appears in the lee of the Alps, often over the Po Valley. Rain and thick cloud tend to develop over much of northern Italy, but especially over the Po Valley, the Alpine foothills and the head of the Adriatic.

The main cyclogenetic or formative phase is illustrated in fig. 16. The front has become stationary north of the Alps and Pyrenees, but has burst through the Provence gap; it has also begun to round the eastern end of the Alps. The advancing cold wave has split into two currents, the Mistral stream (labelled 1 on the maps) and an incipient Bora stream (2). It can readily be envisaged that a ridge of high-pressure must have developed over the Alpine Foreland to permit this split. Meanwhile the developing lee-depression is now more pronounced, and is centred close to Genoa. It constitutes a non-frontal vortex developed in the relatively warm and moist current (3) which covers most of the western Mediterranean. By this time there is general rain all over northern Italy, of a thundery character in autumn, with snow over the Alps. Exceptionally heavy rain or snow is likely in the Carnic and Julian Alps and in the Dinarics. We may note that, if the advancing cold air is very deep, the cold front may surmount the Alps at high levels, and so overrun the developing cyclone over northern Italy, greatly increasing the instability. This overrunning cold front is sketched in fig. 16.

Fig. 17 depicts the ultimate development. The lee-depression, having become a well-established system, has begun to move south-east, and will probably continue to do so until it reaches Asia Minor or the Levant. The cold air (1) which came from Provence is still travelling south-east south of the centre; the cold front ahead of it will probably be traced to the Nile and beyond. The Bora (2) has set in over northern Italy, and is in contact with the Mistral current west of the centre. Quite often a cold front develops between them, as the Bora is usually the cooler current. Over the Alpine foothills and the Po Valley a north-föhn may have set in, giving clear skies and warm weather over a limited area. Rain is now largely confined to the southern Dinarics.

Though there are differences in detail, most Atlantic fronts that approach the Alps at all vigorously lead to cyclogenesis in the manner described; quite obviously the peculiar nature of the mountain barrier is responsible for the particular patterns the development takes. Probably more than 60 per cent of all Mediterranean cyclones are of this curious type, which is responsible for much of the rainfall of the central and eastern Mediterranean lands as well as the Levant, Iraq and even northern India and Iran.

We may note in passing two special features of the weather of the western Basin. One is the stormy weather experienced along the coasts of Morocco, Algeria and Tunisia as the cold front leading the Mistral stream (1) arrives Heavy rainfall occurs on all the north-facing slopes of the Atlas, and the instability showers occur long after the front itself has gone by. The wet weather which so much affected the north African campaign of 1942–3 was of this type. The other is the peculiar winter weather of the Po Valley, surely one of Europe's most unpleasant spots in winter. The surrounding girdle of mountains traps the surface air, and radiative cooling lowers its temperature. The middle and upper parts of the valley are hence occupied for a large part of winter by a shallow, inert pool of air in which there is little motion. Fog is very widespread; Milan reports an average of twenty-two foggy mornings in January, and on no less than fourteen days there is still fog in mid-afternoon. Turin is at least as bad. The circulation of the lee-depressions described above very often fails to dissipate the foggy layer, and steady rain may fall for hours to increase the gloom. It is a little hard to reconcile these figures with our impressions of the Italian climate, but the fact remains that Milan's winter is more unpleasant than Manchester's.

Though the Alpine lee-depression is distinctly the dominant cyclonic type, we must not neglect the others. Fairly widespread rain may occur in Portugal, Spain, Morocco, or southern France when shallow frontal waves travel from the Bay of Biscay or Atlantic into the western Mediterranean. Such depressions may travel eastwards along the Basin for considerable distances. Slightly different in character are the disturbances which form over the Mediterranean when pressure is high over continental Europe. At such times a flow of cP air sweeps westwards from the Balkans across Italy towards Spain. There are usually shallow depressions strung out along the Mediterranean Basin, though in many cases they are stationary or erratic in movement. Heavy and frequent showers occur within these depressions, presumably because of convergence and the heating of cP air by the sea as it circulates around the centres. Occasionally true eastward-moving wave cyclones develop in these circumstances.

The last general class of cyclone requiring our attention is the Saharan type. These are storms forming as frontal cyclones over the northern Sahara with warm sectors of cT air; their origin has been discussed by the present writer (1943). Saharan depressions form chiefly in the trough south of the Atlas on fronts separating mP from cT air. As they move north-eastwards into the Mediterranean they bring a wave of cT air out across the African coast. Though they give little rain, the weather of the warm sector is a characteristic weather type of the southern Mediterranean. The hot, dusty, desert winds from the south-west which blow as soon as the warm front is past are known as the Ghibli, the Scirocco in southern Italy and the Khamsin in Egypt and the Levant. High temperatures, very low humidities and thick dust haze are characteristic. If the winds are strong, actual dust-storms or sand-storms occur east of Gabes, especially just before the arrival of the cold-front. Over the relatively cool Mediterranean the surface layers rapidly become saturated, and an unbroken fall of low stratus covers regions reached by the airmass (which aloft is known to have reached Scotland on at least one occasion).

It is worth noting that the very dry character of this air precludes heavy precipitation. The moisture added at low levels does not affect the body of the airmass, which is too stable, for convection to carry the moisture upwards. The rainy "sciroccos" of the Adriatic are south-easterly currents of maritime air circulating around low-pressure systems to the south and west.

Conclusion

And so, for five or six months, the winter goes. Rainy or showery weather associated with the cyclones alternates with brighter weather in the intervening periods. The characteristic sky is cumuliform, and there is no equivalent for the dull, gloomy weather of north-west Europe. Weather is everywhere mild except near the gaps taken by the cold Mistral, Bora and Vardarac. The Spanish Meseta, too, because of its height, has a colder winter.

By May the warm sun has restored temperatures to near-summer levels. The brief vegetative period is at its height, as the plants take advantage of the accumulated winter moisture and the warmth. Towards the end of the month the last major polar outbreak reaches the western basin, and the final cyclonic rains are experienced. Then the atmosphere settles down to the fourteen or fifteen weeks of unbroken tranquillity that bring the hot and arid summer.

CHAPTER XV

ON HOW TO READ FURTHER

THE title of this final chapter might perhaps be sub-titled "Or What to Read Instead". There are so many different motives behind the study of the weather that every book on it must disappoint someone. For the disappointed, and for those whose appetites have been whetted, the literature of weather study is briefly and selectively reviewed below. These references are quite independent of the specific references given in places in the foregoing text.

The literature will be reviewed under the following general heads:

1. Meteorology
    (a) Dynamic and Physical
    (b) Synoptic and applied
    (c) Reference volumes
2. Climatology
    (a) Dynamic
    (b) Physical and Statistical

1. *Meteorology*

(a) *Dynamic and Physical.* By this term we mean the study of atmospheric physics; i.e. of atmospheric processes as physical phenomena. This general field is mathematical in method, and will be inaccessible to most undergraduates in geography or the biological sciences. The following texts are arranged roughly in ascending order of difficulty mathematically:

(i) *S. Petterssen: Introduction to Meteorology.* McGraw-Hill, 1942. An excellent and simply conceived study by an outstanding leader in the field.

(ii) *H. R. Byers: General Meteorology.* McGraw-Hill, 1944. Rather more advanced in character than the preceding text, this book is very clearly written, and is of real value to the non-mathematical student.

(iii) *E. W. Hewson* and *R. W. Longley: Meteorology, Theoretical and Applied.* John Wiley, 1944. Roughly on a par with (ii) as far as difficulty is concerned, this Canadian reference work is refreshingly different in coverage. Strongly recommended.

(iv) *D. Brunt: Dynamical and Physical Meteorology.* C.U.P., 1937. This beautifully produced and presented book is perhaps the most often quoted reference book in the field. It is severely mathematical in method, and is outside the abilities of all but properly qualified students.

In the same category as item (iv) are B. Haurwitz' *Dynamic Meteorology* (McGraw-Hill, 1941) and J. Holmboe, G. Forsythe and W. Gustin, *Dynamic Meteorology* (John Wiley, 1945), though they cover different ground, and represent European work as much as American or British. A great pioneer study in the German language, *Physikalische Hydrodynamik*, by V. Bjerknes and collaborators, has been reprinted by Edwards Bros. of Ann Arbor, Michigan, but is quite beyond the range of all but the professional meteorologist.

(b) *Synoptic and Applied.* This includes the branch of weather study concerned with weather analysis and forecasting. It also deals with many of the newer ideas in meteorology which have developed in the course of chart analysis, and which are still treated under this heading rather than under Dynamical and Physical meteorology. References (ii) and (iii) above apply with equal force in this second branch of meteorology, and we can add the following:

(v) *H. C. Willett: Descriptive Meteorology.* Academic Press, 1944. This study by one of the pioneers of airmass meteorology in the U.S.A. is the best review of airmass analysis available. It presents some of the new ideas about the general circulation which have been gaining adherents in the U.S.A.

(vi) *V. P. Starr: Basic Principles of Weather Forecasting.* This book is unusual in both scope and method. It deals chiefly with the still immature but significant methods of analysis which have been sponsored by C. G. Rossby and his collaborators in the University of Chicago.

(vii) *S. Petterssen: Weather Analysis and Forecasting.* McGraw-Hill, 1940. By far the most comprehensive and definitive study of applied meteorology. This book is absolutely indispensable to anyone concerned with the atmosphere, including the climatologist.

(c) *Reference Books.* There are several works of an encyclopaedic character which are of great value to the student. Among these we can list:

(viii) *Sir W. Napier Shaw: Manual of Meteorology.* C.U.P., 1926–34. The four volumes of Shaw's Manual contain, besides much theoretical and descriptive text, great quantities of statistical tables and maps, many of which have never been superseded.

(ix) *F. A. Berry, E. Bollay* and *N. R. Beers: Handbook of Meteorology.* McGraw-Hill, 1945. The three editors of this compendium are instructors at the United States Naval Academy. The book contains useful working tables, reference material about codes, charting procedures, etc., and a more or less complete review of theoretical and applied meteorology by various authors, including C. G. Rossby, H. U. Sverdrup and other authorities.

(x) *U.S. Dept. of Agriculture: Climate and Man, Yearbook for* 1941. This unique book is in no real sense a yearbook. It contains great quantities of statistics about U.S. and world climates, but also contains some thirty-eight articles on various aspects of climatology and meteorology, some of which are of fundamental value. It should be on the shelf of every geographer, biologist and meteorologist, to say nothing of farmers and administrators.

2. *Climatology*

(a) *Dynamic*. By this term we mean the explanatory description of climates in terms of the circulation or disturbances of the atmosphere. There is little literature in this little-explored field, which was first defined by T. Bergeron in the article "Richtlinien einer Dynamische Klimatologie" in the Meteorologisches Zeitschrift for 1931. The one reference fully within this field is:

(xi) *B. Haurwitz* and *J. A. Austin: Climatology*. McGraw-Hill, 1944. This study contains a brief review of some major physical features of the atmosphere, but the main part of the work is a regional study of world climates on a dynamic basis. It assumes a considerable technical equipment on the part of the reader.

(b) *Physical*. Physical climatology is the branch of the subject dealing with climate on the basis of the physical elements, viz. temperature, humidity, rainfall, cloudiness and the like. The treatment is descriptive and statistical, and is usually organized on a regional basis. Workers in this field are usually either meteorologically or geographically trained, rarely both. The metereological climatologists have been especially concerned with statistical method. The two most recent studies in this field are:

(xii) *H. Landsberg: Physical Climatology*. Pennsylvania State Col., 1940.

(xiii) *V. Conrad: Methods in Climatology*. Harvard U.P., 1944. Both volumes deal with methods of rationalizing and utilizing observational materials. Conrad is a noted Austrian climatologist and his book contains valuable material for anyone making a study of European climates. A new edition (with L. W. Pollak) has just been published (1951).

Among works by geographical climatologists we can select the following:

(xiv) *W. Köppen* and *H. Geiger: Handbuch der Klimatologie*. Geb. Bornträger, Berlin. Published at intervals over the past twenty years, this remarkable series represents the work of a large number of regional specialists, gathered into a series of volumes edited by Wladimir Köppen and his junior editor, and presented on a uniform plan. Most of the world is covered in detail, and statistics support the authoritative text. Köppen was the great pioneer of systematic climatology, and this work is a fitting tribute to him. Unfortunately it is now hard to get hold of, but it is to be hoped that it will be reprinted.

(xv) *W. G. Kendrew: Climatology*. Oxford U.P. This book, long established as a favourite among geographers, is hard to place. The treatment is physical, but not regional. Each element of climate is treated separately, and the work suffers from a lack of integration; the dynamic aspects of climate are inadequately treated.

(xvi) *W. G. Kendrew: Climate of the Continents*. Oxford U.P. The companion volume of (xv) applies the principles set out in that volume to the climates of the world on a regional basis.

There are many other works similar in scope, such as A. A. Miller's *Climatology*, R. de C. Ward's *Climate*, G. T. Trewartha's *Introduction to Weather and Climate*, and T. Blair's *Climatology*. The latter is very sound as to physical principles.

Statistical materials are collected in many publications. The most complete world climatic normals are contained in *World Weather Records*, published by the Smithsonian Institution at Washington. For the British Isles, the Meteorological Office publication *Book of Normals* is valuable. Most of the general climatological texts referred to above also include tables of climatic normals.

Finally, one must add the *Compendium of Meteorology*, edited by T. Malone, published in 1951 by the American Meteorological Society. It fits into all the categories listed above. Its 1,334 pages include 108 articles on every aspect of the subject, and its list of authors is drawn from many countries. Practically every leading contemporary research worker has contributed.

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(N.B. In the case of certain much-quoted items, the principal reference is in bold).