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EVERYDAY METEOROLOGY



The Thames in flood near Staines

Everyday Meteorology

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HUTCHINSON OF LONDON

HUTCHINSON & CO. (Publishers) LTD
178-202 Great Portland Street, London, W.1

London Melbourne Sydney
Auckland Bombay Toronto
Johannesburg New York



First Published 1958
Second Impression 1961

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*This book has been set in Times New Roman type.
It has been printed in Great Britain by Tonbridge
Printers Ltd., in Tonbridge, Kent, on Antique Wove
paper and bound by them*

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ACKNOWLEDGMENTS

The authors would like to express their gratitude to all who have helped in the production of this book by criticism, suggestion, or in any other way, and are particularly indebted to Miss F. M. Taylor for drawing the line diagrams in a form suitable for reproduction. Acknowledgments for illustrations are also due to the following, the Roman figures referring to plates and the Arabic figures to diagrams in the text.

Aerofilms (*Frontispiece*, vii, xx), B.B.C. (vi), Kemsley Picture Service (xxi), Murray Hardy (x), F. H. Ludlam (xi), C. E. May and Son (v, xiv), D. R. Maxted (xvii), Miss P. Peacock (xvii), E. Pestell (xvi), R. Savigear (xix), R. H. Simpson and the American Meteorological Society (xv), the Director-General of the Meteorological Office (ii), Eidg. Institut fur Schnee und Lawinenforschung (xxii), the Headmaster of St. Crispin's School, Wokingham, for permission to photograph the school weather station (i), Donald L. Champion (67), and the Director-General of the Meteorological Office and the Controller of Her Majesty's Stationery Office for the use of Crown Copyright material in the Daily Weather Report, Daily Aerological Record and Monthly Weather Report (11, 14, 18, 19, 20, 21, 23, 27, 28, 29, 30, 31, 32, 33, 35, 38, 39, 41, 43, 44, 45, 46, 47, 49, 58, 62, 68, 70, 71, 73).

INTRODUCTION

For the defeat and destruction of the Spanish Armada, British seamanship and British weather have shared the credit about equally, and on many another occasion when the fate of nations hung in the balance, the weather at the moment of crisis has been the instrument of destiny. At the other extreme, every one of us can recall in our own little lives moments made memorable by the weather; an hour of happiness heightened by golden sunshine, the trivial tragedy of the spoilt picnic or the sinking feeling on reaching the gates at Lords to be met with the announcement 'No play this afternoon.' The fatalist takes it as it comes, muttering to himself:

Whether the weather be cold
Or whether the weather be hot,
Whether the weather be fine
Or whether the weather be not,
We must weather the weather
Whatever the weather,
Whether we like it or not.

The pessimist whose guiding principle is 'Prepare for the worst' keeps a spare muffler, overcoat, mackintosh and umbrella at the office and goes out to meet the weather suitably dressed for the weather as it is. Even so he may be caught unprepared between home and office, an indignity that he might have escaped had he added wisdom to caution and learnt a little meteorology.

But the hazards of the daily journey to indoor work, discomforting though they may occasionally be, are nothing to the impact of the weather on those who, for work or for pleasure, spend their day in the open air. Let us list a few examples: the farmer or market-gardener who wants to know the needs of his crops for warmth and water, how far these needs are being met naturally, how much irrigation water to supply to make good the deficit of rain in dry weather, whether to cut the hay meadow today, whether to cover up the strawberries and drain the car radiator before going to bed: the shepherd who needs to foresee snow in the hills and to be prepared for the weather at lambing time: the fisherman, the sailor and the yachtsman who want as much foreknowledge as possible of the strength and direction of the wind and the state of the sea: the

mountaineer for whom storm, ice or hill fog are perils that may be fatal: the glider or model aircraft enthusiast who takes advantage of knowledge of winds, calms and 'thermals': the captain of ship or aircraft who regularly challenges the elements on the sea or in the air, not to speak of the understandable interest—if not anxiety—felt by those who are merely his passengers.

How many more of us, less directly exposed to the weather, are nevertheless compelled to pay heed to it? There are the civil engineers, concerned with rainfall in relation to flood risk, storm damage and water storage: heating engineers who want an estimate of next week's temperatures in order to anticipate fuel demands: builders and plumbers, interested in the weather as it affects the setting of mortar and concrete or outdoor painting or the risk of burst pipes: housewives worried about hanging out the washing and the milk souring: department store managers and caterers for whom the estimate of demand for ice-cream, face-cream, hot drinks, snow boots or sunglasses, may be a profitable piece of foreknowledge: promoters and organisers of outdoor sports who wish to know as soon as possible whether the pitch will be fit for play, what the 'gate' may be, or whether to include the spin bowlers in the eleven.

For all these an understanding of the weather is at least an advantage and often an essential, and though they may have gathered a considerable store of experience and empirical knowledge, their skill will be enhanced by an understanding of the principles underlying their private weather lore and by an explanation of what they see with their 'weather eye.'

Where an older generation would have relied on the time-hallowed tenets of weather lore, the present generation increasingly takes the advice of the weather forecast on the radio or TV and feels a dreary satisfaction in blaming a government service on the inevitable occasions when the forecast goes wrong. Nobody blamed the anonymous author of the weather saying if the red sky at night was followed by the very reverse of the shepherd's delight. But today many people evidently feel that they are entitled to completely accurate prediction from a state meteorological service backed, as it is, by all the resources of modern science and moreover, a service for which they, the public, ultimately foot the bill. Three misconceptions, however, underlie this widespread attitude: firstly, that all meteorological facts are known: secondly, that the best possible use is made of those facts: thirdly, that the recipient of the weather forecast may sit back and accept it as inspired prophesy and is excused all responsibility for interpreting it.

The atmosphere is world-wide and very deep, and its qualities and properties have to be assessed from samples of tiny bits of it, far too widely spaced in many parts of the world. Moreover, nearly all these

samples are of the surface layers. Despite great progress in flying and the use of radio and radar sounding techniques, the vertical structure of the atmosphere is still quite inadequately charted. Meteorological knowledge is built up by means of intelligent generalisation from these samples. True, there is much more to come. A great deal is currently being learned by new and exciting techniques like probing the upper atmosphere by means of V-2 type rockets with instruments replacing the explosive war-head. The time is already here when important knowledge may be obtained from equipment carried by earth satellites, the most spectacular of all the projects achieved during the International Geophysical Year. Despite such prospects, there is good reason for doubting whether our knowledge of atmospheric behaviour will ever be more than merely approximate and it is very likely that there will always remain a large residue of uncertainty in weather forecasting because of the inherent unknowableness of the subject.

All this would be true even with the best imaginable meteorological service. British forecasters, working against odds in a region of notoriously capricious weather, have a deservedly high reputation, but one is entitled to ask (and we frequently do) if our meteorological organisation is as good as it can be under the circumstances. The answer to this may well be that we, the public, get the forecasting service we deserve, or are willing to put up with or, in the long run, are prepared to pay for. By maintaining the Meteorological Office on a total annual budget of little more than the cost of a single modern bomber, we are perhaps attempting to get our weather service on the cheap, but this is a matter to be argued in columns other than these.

More to the point is whether or not we get the best out of the forecasts as they are. This is where a certain responsibility devolves upon us, for the better we understand the difficulties of forecasting the more sympathetically we can make allowances; and the more we appreciate the working of weather processes the better chance we have of foreseeing the possibility of failure and of correcting the forecast for our own needs.

It is important, for example, to know what can and what cannot be forecast. On August 15th, 1952, ten inches of rain fell on Exmoor and the floods resulting from this cloudburst devastated the seaside resort of Lynmouth (see Chapter 11). Yet many parts of Southern England had no rain that day. The Daily Weather Report issued in the morning stated, 'In districts of England and Wales south of a line from Pembroke to the Humber it will be mainly cloudy with outbreaks of rain and with thunderstorms in places . . .' On the warm afternoon of May 19th, 1952, a tornado uprooted trees, brought down telephone wires and damaged 100 houses in the Derbyshire village of Tibshelf (see Chapter 7), but either side of a narrow strip of destruction all was calm. The Daily Weather

Report forecast, 'Many places will have fine very warm weather though there may be some scattered thunderstorms . . .' These were not careless or ill-considered forecasts. Such events are due to highly unusual and perhaps fortuitous combinations of circumstances which can hardly be forecast though they can be analysed in retrospect. In each case the situation had its potentialities, but the realisation of these potentialities was a 'thousands to one chance.' But the 'thousands to one chance' is certain to come off somewhere at some time. So it is that sooner or later some place is bound to achieve ephemeral fame with, say, 'the highest temperatures ever recorded at — in the second week of February.' It is like the bowler called Grover

Who bowled seven wides in one over.

It had never been done by a schoolmaster's son

Before luncheon, in August at Dover.

What each of us really wants is a forecast strictly applicable to our own back garden or journey to work, but this is to demand the impossible. Quite apart from the busy housewife in East Cheam who, lending half an ear to the B.B.C. forecast, hears the words 'fine and warm' but misses the qualification 'in Northern Ireland' as she switches on the vacuum cleaner, not enough people realise that the most detailed forecast heard on the air or read in the papers is regional in its scope, that is to say, it is generalised to embrace a region as a whole, although there may well be considerable variation of local weather within it. The problem is very much simpler if only a small area is under consideration, and the shorter the notice required the easier it is. A route forecast for an air-line flight based on the latest 'hot' news of weather along the route rarely leads the pilot astray. In fact, a specialist enquirer usually gets more satisfaction from the forecasting service than does the general public, though this is partly because few people are aware that they too can get up-to-date forecasts if they wish (see Chapter 2).

It is our aim to put the reader in a position to appreciate the potentialities of the general situation and to judge for himself, from an understanding of the effect of local weather-generating factors, what is likely to happen to him and how likely it is to happen. We do not expect that, as a result of reading this, or any other book, he will be able to rival the professional in weather acumen. In the first place, he lacks the detailed data available to the forecaster and, in the second place, he lacks the techniques of handling them. At the same time we think it well worth while suggesting ways of gathering a modicum of weather 'gen,' and we describe some elementary tools for making something of it.

Open any text-book of advanced meteorology and you will conclude that meteorology is a branch of mathematics; look into any research

journal and your fears will be confirmed. But do not, if you are alarmed at the prospect, leap to the conclusion it is not for you. It would be foolish to belittle the contribution of mathematics to meteorology: it is essential for the detailed understanding of the involved physical processes of weather and it is now becoming increasingly valuable as a tool in forecasting itself. But those trained to use it know best that the logic of mathematics is too precise to be applied to the weather and that the quality of the result obtained in the end from an electronic computer, as from a sausage machine, depends on the quality of the materials fed into it. Thus there is good reason for leaving the application of the mathematical analysis of weather phenomena to the experts, pointing out that this is a tool they must use with caution, and there is some justification for presenting a 'Met. without Maths,' which is what we intend to do.

Our presentation will be at the same time descriptive and explanatory, and our approach is through the appreciation of the Air Mass. There is nothing particularly original in this: the present generation of meteorologists has progressively adopted it as the unit of weather analysis since the statement of the principle by the Norwegian School at the end of World War I. While not disregarding subsequent modification, it is no exaggeration to say that the work of V. and J. Bjerknes and their associates provided the key that unlocked the major mystery of middle-latitude weather (though doubtless there are other doors beyond and further keys required). The distribution of rain, wind direction and shifts, temperatures, cloud forms and pressure changes found, in most (though not all) cases, a rational explanation and all the obscurely related or seemingly unrelated elements of the weather fell at once into recognisable and harmonious patterns. What nature has put together let us not take asunder: there is no sense any longer in piecemeal description and analysis of each individual weather element.

The senior meteorologist at the Central Forecasting Office looks down on the weather, as it were, from a great height, seeing it as a pattern of lines representing the broad air flow at various levels, his mind pre-occupied with various means of projecting the basic patterns twenty-four hours into the future. By the time the fruits of his deliberations—aided now by the whirring of the electronic computer—reach the average meteorological office where everyday problems are faced, they must be translated into terms of air masses and fronts and their associated phenomena. For us, the public, who, so to speak, see the weather from below, these are the elements of our weather experience.

These discoveries arose out of the suppression of weather information by the belligerents of the First World War. Neutral Norway, thus deprived of knowledge of the broad pattern of northern hemisphere weather, was driven to an intensive study of her own weather. A close

network of stations was established and continuous reports enabled the meteorologists to track each air stream, and to detect the boundaries or fronts along which air streams of different character, coming from different sources, met and over-rode or under-thrust each other. Here they found rain was made—broad stripes of steady rain at what came to be known as warm fronts, narrow stripes of squally rain at what came to be known as cold fronts—and thus the basic structure of our familiar depressions was revealed.

But on February 8th, 1868, commenting on the weather reports then reaching Greenwich from the United States over the newly laid Atlantic Cable, the magazine *Once a Week* had observed 'The report will, as time goes on, and as other links are added to the chain of observing stations, prove of great value for the study of *the movements of the Polar and Equatorial air-currents and ultimately for predicting the elemental wars to which the confliction of these gives rise.*' How near they had come to grasping the key! And thirty years earlier still Luke Howard, describing the climate of London in 1833, had written 'A Southerly current, charged with vapour from a warmer region' (now called Tropical Maritime) 'may be passing Northward at the same time that a Northerly current' (Polar Maritime) 'may be returning towards the South, in the immediate neighbourhood of the former; and these two may raise each other, the colder running in laterally under the warmer current and causing it to flow over laterally in its turn, while each pursues in the main its original course. In this case the country for a considerable space from about the line of their junction' (the front) 'far into the Southerly current may be the seat of extensive and continuous rain.'

The Bjerknes 'discovery' then was an independent rediscovery of old knowledge forgotten for nearly a hundred years and overlooked in the enthusiasm for the new world-wide synoptic meteorology made possible by the telegraph and the wireless. In the years between, dynamic meteorologists, dazzled by the new electric toys, got bemused with isobar patterns and lost sight of the simple clue until, the bright light of radio communication being blacked out for the First World War, Bjerknes lit the homely lantern of close observation and, searching diligently among his own weather, found again the long-forgotten key.

Now that the radio silence and the black-out of the Second World War are over, world meteorology is illuminated by a brighter light than ever before. The radio, the television and the daily charts in the newspapers keep us informed concerning the weather. We have no difficulty in fixing our position in the world pattern of weather events. Armed with some understanding of meteorology we know more of the potentialities of our environment and so are the better equipped to order our day.

GATHERING THE INFORMATION

Meteorological Observations—Significant and Otherwise

Measurements and observations of the state of the atmosphere form the raw material of weather studies. As a descriptive science, meteorology is of ancient origin: as a physical science, it dates from the seventeenth century and the invention of such instruments as the thermometer and the barometer. Since then, these instruments have been much improved, others have been devised, and methods have been established of observing qualitatively those weather elements that cannot be measured.

So that meteorological measurements and observations shall be reasonably accurate and strictly comparable, whoever does the observing, the Meteorological Office has laid down precise standards and methods, designed to make them as far as possible representative of the general condition of the air, an aim not easy to achieve. For example, if we hang up a thermometer out of doors on a sunny day, let us not delude ourselves that it will register the true temperature of the air. Different substances exposed to the sun's radiant heat warm up to differing extents which depend on their physical properties. The air (as we shall see in a later chapter) is warmed very little by direct solar radiation: on the other hand, a thermometer in the sun warms up considerably and records its own temperature which is much higher than that of the air. Brick walls, masonry and paving, too, warm up strongly in the sun, and we ourselves may feel comfortably warm in the sunshine of a bright winter day, while the actual air temperature is surprisingly low. For our thermometer to register faithfully the temperature of the air, we must ensure that it is shielded from direct solar (and other) radiation. At the same time, it must be well ventilated, that is, cut off as little as possible from the free-moving air, if it is to record a representative temperature.

The official thermometer housing in use in this country, the Stevenson Screen, consists of a wooden box painted white to reflect away solar radiation, with a double roof which further insulates the interior, and with louvred sides and floor to ensure reasonable ventilation (see Fig. 1 and Plate I). The hinged door faces north so that there is the least chance of solar rays entering when readings are being taken. The screen is mounted on a stand of such a height that the thermometers are about four feet above the ground: this is an attempt to escape the disturbing influence of

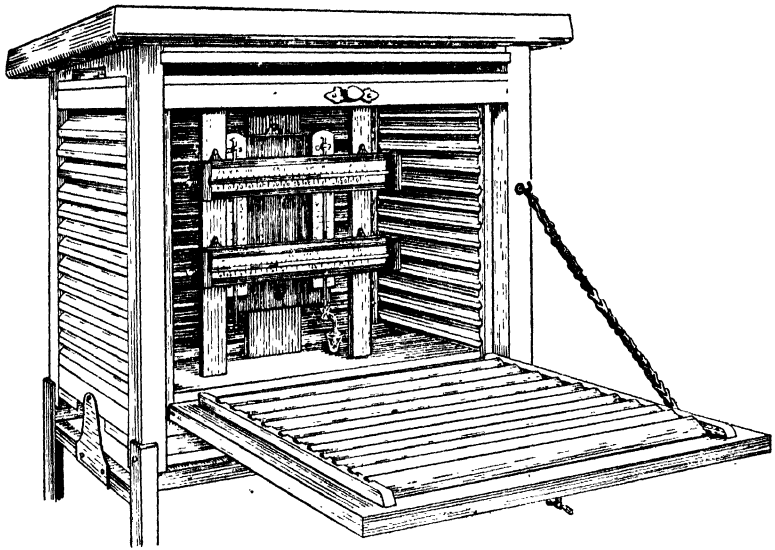


Fig. 1. Stevenson screen, open to show the arrangement of the thermometers

the ground surface which often strongly affects temperature in the lowest air layers. Quite incidental is the fact that screen thermometers record temperatures at about the height at which we breathe, and therefore provide some information about the atmospheric conditions significant to human beings. The weather and climate experienced by the dog, or the growing plant, however, may be a different story.

Because of the complications that might be introduced by different types of ground surface, the screen is always sited over close-cropped grass, which is a kind of standard meteorological surface for land observations. The ground should be fairly level, both at the site and for some distance around, as the air behaves curiously on and near slopes. The screen must not be placed near surfaces that heat up strongly in the sun and in turn unduly warm the adjacent air. A site equally to be avoided is a small enclosure, like a walled-in garden or an interior courtyard, where the air remains stagnant. All these precautions are designed to ensure that the temperature readings are representative and do not reflect purely local peculiarities.

Comparable difficulties attend the accurate recording of rainfall. A rain-gauge is essentially no more than a funnel leading into a bottle so that falling rain can be caught and measured. But in the standard type of gauge used in this country (see Fig. 2), several precautions are taken. The metal (generally copper) funnel, which has a deep rim to prevent rain being

splashed or blown away and a tiny opening to cut down loss by evaporation, fits into and is supported by a cylindrical copper can which is partly sunk into the ground. Within this can is an inner vessel in which the collecting bottle stands. In the rare case of the bottle overflowing, the excess water collects in the inner vessel which is easily lifted out, and there is no need to uproot the whole gauge. Two standard sizes of gauge are in use (the diameter of the mouth of the funnel being either 5 inches or 8 inches) and with each goes a glass measuring cylinder of appropriate scale into which the water is poured and read off (in parts of an inch or in millimetres).

The outer vessel is embedded firmly in the ground so that the mouth of the funnel is perfectly horizontal and the rim is exactly one foot above the ground. This height is governed by the following considerations. Nearest the ground the strength of the wind is at its least: rain therefore tends to fall most nearly vertically, and the maximum catch is recorded. Higher up (and this is why roof-top sites are not recommended) the stronger wind blows the rain across the mouth of the gauge and the catch is reduced: the eddies and swirls occurring near buildings emphasise this. At the other extreme, if the rim of the gauge were only slightly raised from the ground, there might be an unwelcome addition to the catch by splashing in from outside.

Like the Stevenson screen, the rain-gauge must not be situated near buildings or trees which may shield it from some of the rain it should rightfully receive: on the other hand, curiously high rainfalls have been recorded by gauges placed just below the edge of a sloping roof. There are disadvantages also in too open an exposure, for if the gauge is the only obstacle projecting above the ground on a wide level expanse, it will itself set up air eddies which will diminish the catch.

The barometer is the only meteorological instrument which performs its normal task indoors. Outside, it proves extremely sensitive to gusts and lulls of wind and records a rapidly fluctuating pressure. Even indoors, draughty locations should be avoided. A mercury barometer has also an incorrigible tendency to behave as a thermometer (not surprising in view of its construction), and should not be exposed to direct sunshine or the heat of fires or radiators. At official meteorological stations, barometer

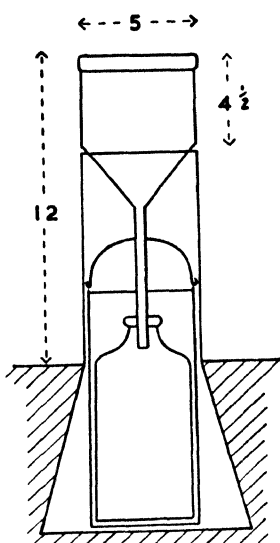


Fig. 2. Rain-gauge: cross-section diagram (dimensions in inches)

readings are adjusted to a standard temperature, and other corrections are applied—for latitude, with which the force of gravity (hence the weight of the mercury) varies, and for certain errors peculiar to the individual instrument. Usually the biggest adjustment is for altitude. Atmospheric pressure is simply the weight of overlying air and therefore varies with altitude (decreasing in the first 5,000 feet by about 1 inch in every 1,000 feet of ascent). The often considerable variations due purely to the relief of the land must be eliminated before the true pressure pattern can be studied. This is one thing, and we shall meet others later that may make nonsense of the legends—‘Rain,’ ‘Change,’ ‘Fair,’ etc.—which still unaccountably appear on commercially produced ‘hall’ barometers.

A School Weather Station

The foregoing remarks concern some of the precautions which make the difference between significant and meaningless meteorological observations. They may interest, for example, the teacher proposing to establish at his school a simple meteorological station which would serve admirable educational purposes and at the same time yield valuable local records. We hope he will aim high: many schools have surprisingly well equipped weather stations (Plate I). What follows now is intended to assist the teacher (or anyone else) who aspires to run a station: it is however no more than an introduction, and he must not be long without the Meteorological Office publication *The Observer's Handbook* which is the essential and indispensable reference book.

The first requirement is a suitable site on an open level lawn which allows a correct exposure for the outdoor instruments. Certain rules govern this. If, near the site, there is an obstruction (wall, building, tree, etc.) which is x feet high, then both thermometer screen and rain-gauge must be at least $2x$ feet away from it. (This means too that screen and gauge must be some 10 feet apart.) If such a site is not available, then this is an unfortunate start from which complete recovery is not possible. The teacher must then use his judgment to secure the best alternative. A rooftop site is better than nothing: the records will still be of great interest in their day-to-day fluctuations, like those of the famous thermometer on the Air Ministry roof at Kingsway, but, again like the latter, will not be comparable with others from properly exposed stations.

Next arise the problems of equipment—and finance. Thermometer screens and stands are expensive to buy, but they can be home-made, and this offers a nice challenge to the handicrafts class. A pamphlet, M.O. Form 63, *Instructions for Making Thermometer Screens of the Stevenson Type*, can be obtained from H.M. Stationery Office. Similarly, a very

serviceable rain-gauge of standard pattern could be fashioned out of readily accessible materials in a school with a good workshop. Thermometers must be purchased, and these are of various qualities and various prices. Best and most expensive are the 'sheathed' types, with the graduated stem enclosed within an outer protective glass sheath. Others, less expensive, have the stem mounted on a wooden or plastic back-board which carries the figuring of the scale. With these, the unprotected figuring tends to wear off in time, and the stem may become displaced in relation to the scale. On the other hand, the mounted types might well have a longer expectation of life, especially if the bulbs are protected by metal guards.

The screen houses four thermometers (see Fig. 1). Two are of ordinary type, the 'dry-bulb' and 'wet-bulb' thermometers: the latter has its bulb encased in a muslin bag kept moist by a cotton wick which trails into a

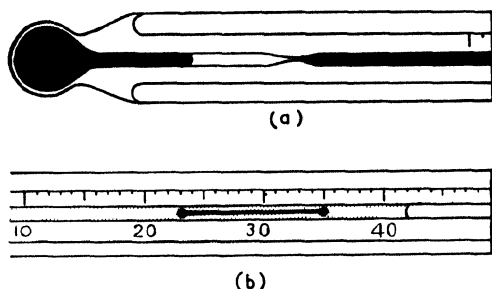


Fig. 3. (a) Maximum thermometer: detail showing the constriction in the bore. (b) Minimum thermometer: detail showing the index

container of distilled water. These two instruments give respectively the 'dry-bulb' (or ordinary air) temperature and 'wet-bulb' temperature (the significance of which we shall explain later), at the time of observation. From the two temperature values can be derived also an expression of the humidity of the air. The muslin and wick quickly become dirty, especially in towns, and they may require changing at least once a week.

The other two thermometers are self-registering: they provide maximum and minimum temperatures over a given period. The maximum thermometer differs from the ordinary type only in having a constriction in the bore about an inch from the bulb (Fig. 3, a). When the morning rise of temperature occurs, the expansion of the mercury forces it through the constricted part of the bore: when temperatures fall in the evening, the mercury thread beyond the constriction cannot flow back under its own weight and, remaining detached from the mercury at the bulb end, registers for us the highest temperature of the day. This process leaves a gap in the

thread between the constriction and the bulb (where the mercury has receded). After the 'max.' has been read, it must be 'set,' that is, the gap must be closed and the thread re-united. We achieve this by swinging the thermometer once or twice at arm's length: the action is equivalent to the shaking of the clinical thermometer, which is a specialised version of the maximum type. The minimum thermometer contains not mercury but spirit, and within the spirit column is a light index, rather like a pin with heads at both ends (Fig. 3, *b*). When the temperature falls at night, the spirit column retreats towards the bulb end and drags the index with it by a surface tension effect: when the temperature rises again after dawn, the spirit flows past the index which is left behind, registering the lowest temperature reached during the night. What we read therefore at the morning observation is not the end of the spirit column (which records the present temperature) but the end of the index, and, after reading, we set the 'min.' by tilting it so that the index slides down into contact with the end of the liquid.

A minimum thermometer of sheathed pattern can also serve as a grass minimum (or terrestrial radiation) thermometer. This is exposed at night

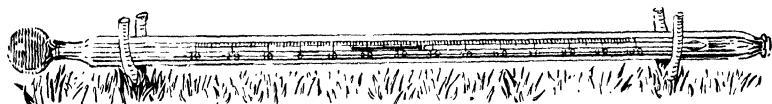


Fig. 4. Grass minimum thermometer in position over short grass

on an open patch of close-cut lawn, supported by two pegs or twigs, so that the bulb is in contact with the grass tips (Fig. 4). The instrument registers the lowest temperature reached over the grass during the night. A grass minimum of below 30.4° F. constitutes a 'ground frost.'

A school with a science laboratory will no doubt possess a mercury barometer of some sort, and every science teacher seizes the opportunity of repeating Torricelli's original experiment and showing how an air column extending the depth of the atmosphere is balanced by a column of mercury some 30 inches high. This also indicates the origin of the measurement of pressure in terms of length. Today a unit of force—the millibar—is generally adopted (a pressure of 1,000 mb is equivalent to that of just over 29.53 inches of mercury). Probably only the older pupils will easily manage the fine adjustments involved in the reading of a standard pattern mercury barometer. For younger meteorologists, the less accurate aneroid barometer is probably more suitable. This type consists basically of a disc-shaped metal box which is partly exhausted of air; the underside of the box is fixed, but the extremely thin and flexible lid is more or less depressed in response to changes in atmospheric pressure. The tiny movements so produced are magnified many times by a system

of levers, and arranged to move a pointer over a calibrated circular scale.

With the measurement of temperature, rainfall and pressure thus arranged, our teacher is entitled to be content with the instrumental recording, and to turn his attention to eye estimations and observations. These can be classified under the headings of surface wind, cloud and weather. There are instruments for measuring wind speed (anemometers), but these are expensive and exacting in their requirements of exposure. The best method where fine accuracy is not essential is the estimation of wind force in terms of the Beaufort Scale, which is a series of descriptions of the way common outdoor objects—smoke, wind vanes, the leaves and branches of trees—respond to different wind conditions. The Beaufort Scale (which is reproduced in the Appendix) gives better training in observation than the reading of a pointer on an anemometer dial. Wind direction may be estimated in various ways: by means of a wind vane which can be home-made (but unless this is well away from buildings and trees it will show directions considerably influenced by the peculiar air flow near obstacles), or a light flag on a pole (the same proviso applies here), or smoke from a tall factory chimney (if fairly near), which will give a direction even with the lightest (Force 1) wind.

The identification of cloud types provides a great deal of interest: a convenient little cloud atlas is included in the *Observer's Handbook* (or can be bought separately), which may help with cloud observations (see also Plate II). It is also usual to make a rough estimate of the amount of cloud, in 'oktas' (eighths) of sky covered. Finally, 'weather'—the look of the sky, precipitation if any, the state of the ground, etc.—is described by the convenient short-hand of Beaufort letters, of which a selected series is given in the Appendix.

The remaining problems are of organisation, and here every teacher will have his own solutions. The time of observation should be the same each day and 0900 hr. G.M.T. if possible. In addition to the permanent record of observations, many teachers will devise ways of displaying the current information, perhaps graphically, so that day-to-day fluctuations can easily be seen and appreciated.

Official Weather Stations

The organisation of weather observing in this country is in the hands of the Meteorological Office, a department of the Air Ministry. Stations which are official, in the sense that they report regularly to the Meteorological Office, vary from school weather stations like that just described to very fully equipped and staffed stations which provide up-to-the-minute information for forecasting purposes. The latter are known as full synoptic stations, where a wide range of surface observations and measurements

are taken and reported each hour of the day. There are also auxiliary reporting stations from which less comprehensive observations are reported less frequently. These two types of stations are generally located on airfields, but may sometimes be lighthouses or coastguard stations or even stations manned by individuals where these can fill gaps in the observational network in areas unsuitable for airfields.

Reports from the 90-odd synoptic stations are intended primarily for forecasting purposes, though they are preserved also for climatological records. Over 300 stations provide data purely for climatological needs: these are not Meteorological Office stations but are maintained by observatories, industrial concerns, local authorities, universities, schools, and private individuals. Since their information is not required immediately, such stations usually send in monthly returns. Some of these voluntarily maintained stations are very well equipped, and a few carry out specialised observations for research or other needs. Climatological stations are classified as 'normal,' if observations are made at least twice daily, or 'auxiliary,' if only once daily (0900 hr. G.M.T.). Some climatological stations are situated at seaside and other resorts and are distinguished as 'health resort stations': apart from the usual routine, these stations send in every evening to the Meteorological Office reports (in which special attention is paid to maximum temperatures and sunshine) which are later issued to the Press. Certain other climatological stations have additional duties especially concerned with measuring soil temperatures, and these 'crop weather stations' contribute to various agricultural research projects.

Official weather stations possess at least the basic equipment described in the previous section and often more besides. Most synoptic stations have a well-exposed anemometer and wind-vane to give accurate measurements of wind speed and direction. One of the standard types of anemometer consists of three metal cups held on arms mounted on a vertical steel spindle (Fig. 5): the cups rotate in the wind, at a rate proportional to the wind speed, and the motion activates a small electrical generator. The wind speed is shown on the indicator dial of what is really a disguised voltmeter, calibrated in m.p.h. or knots. The wind-vane too can be made to indicate automatically—the pointer fluctuating over a dial calibrated in degrees from true north.

Most of the other additional equipment is of the continuous-recording type. In all these instruments, the recording element, whatever it may be, controls the movement of a pen which marks a continuous ink trace on a chart wrapped round a drum rotated by clockwork. For example, in the thermograph, which provides a continuous temperature record, the thermal element is a coil formed of two strips welded face to face of different metals which expand at differing rates when heated. This bi-

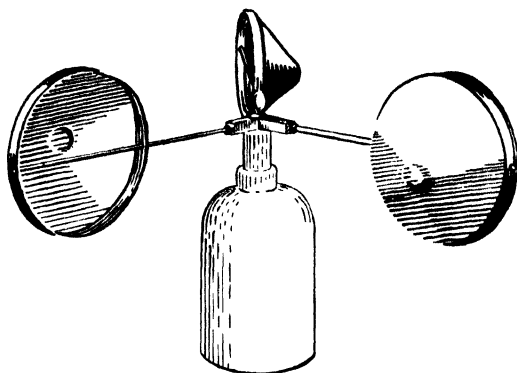


Fig. 5. Cup anemometer (*see also Plate I*)

metallic coil responds to temperature changes by uncoiling or coiling further, and this mechanical action is harnessed to activate the pen. The hygrograph, which gives a continuous humidity trace, makes use of a long-recognised property of human hair: the length of a hair, if cleaned of natural grease, varies with changing humidity. In the instrument, by an ingenious mechanism, the change in length of a bundle of hairs operates the recording pen. Both these instruments are accommodated, with the four thermometers, in large-size Stevenson screens. Broadly similar is the barograph, in which several aneroid elements provide the mechanical movement which is responsible for the trace. The pressure trace is especially valuable as the behaviour of the pressure during the preceding three hours forms an important item in the weather messages sent by synoptic stations. Some examples of thermograph, hygrograph and barograph traces are shown in Fig. 45.

A satisfactory continuous-recording anemometer is the pressure-tube anemograph invented by W. H. Dines. This is an intricate affair, but essentially it records the pressure (and so the speed) of wind blowing into a tube attached to a vane which keeps it facing into wind. The mechanism activates a recording pen which makes a continuous trace on a chart calibrated in m.p.h. Incorporated is a direction recorder, controlled by the vane, and a second pen moves over the lower part of the same chart, which is calibrated in degrees from north. To watch one of these expensive instruments in action is to be convinced of the restless nature of the wind: the movements of the pens reveal a ceaseless fluctuation of speed and direction (Fig. 45).

Many stations—particularly the health resorts—keep sunshine records. The standard instrument in this country is the Campbell-Stokes recorder (Fig. 6) which provides a daily record of the duration of bright sunshine. The principle is very simple: a glass sphere, acting as a lens, focusses

the sun's rays on to a sensitised card held in a metal half-bowl with which the sphere is concentric. With the apparent movement of the sun across the sky a brown trace is scorched on the blue card, and, as the latter is calibrated in hours, the duration of sunshine, whether continuous or interrupted by cloudy periods, may be read off. The recorder must be set up in the most open situation available, often on a rooftop.

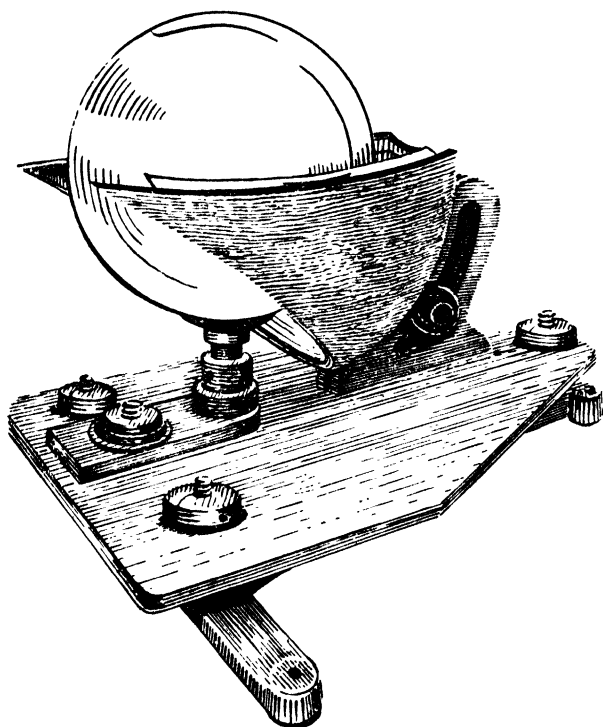


Fig. 6. Campbell-Stokes sunshine recorder

Finally among the continuous-recording instruments is the autographic rain-gauge, usually of the 'tilting-siphon' type. This consists of a large rain-gauge in which the water collected falls into a chamber, raising a float which activates a recording pen in the usual way. When no rain is falling, a horizontal trace appears at the bottom of the chart: when there is rain, the pen climbs the chart, and the steepness of its ascent depends on the rate of rainfall. When the chamber is full (and the pen is at the top of the chart), it tilts and empties through a siphon tube: the pen immediately returns to zero on the chart, and the processes of collection and recording can begin again. From the chart, it is possible to determine the amount and duration of rain, and the rate at which it fell (see Fig. 45).

Synoptic stations particularly are concerned to make detailed observations of weather elements like cloud and visibility which defy measurement in the ordinary sense, but which are of great significance to aviation. The height of the base of clouds is usually estimated, but with a very low 'ceiling' which may be marginal for aircraft, some method of measurement becomes necessary. During daylight, this is done by noting the precise time taken by a coloured hydrogen-filled balloon with a known constant rate of lift to disappear into cloud. At night, a 'cloud searchlight' is used: this directs a beam vertically upwards, which forms a characteristic patch of light on the base of the cloud. The angle of elevation of the light patch is measured from the station and the calculation of the height is a matter of simple trigonometry. Visibility is estimated by noting which of a series of 'visibility points' (landmarks ranging from nearby buildings to distant spires or hill-tops) at known distances from the station remains recognisably visible. At night, observers depend on the visibility of fixed lights at known distances.

The Amateur Observer

The amateur weather-watcher may well admire the equipment and techniques described in the foregoing sections. They represent a peak to which he may aspire, and even move some way towards. More than one amateur runs a fine station, and some even report to the Meteorological Office. Many more, however, must be content to keep a weather eye open, and perhaps to hang up a (preferably shaded) thermometer on the best possible site in a garden not designed with meteorological considerations in view. The time-hallowed north wall of house or shed is better than nothing.

Many amateurs add to the store of climatological records by keeping a rain-gauge and sending monthly returns to the British Rainfall organisation (a branch of the Meteorological Office). Provided the gauge is of standard type and the exposure suitable, new rainfall observers are always welcome, especially from remoter areas. Most of the 5,000-odd gauges in the country are maintained by amateurs.

Weather Stations at Sea

Weather forecasters need information from wide areas which include sea as well as land: the British forecaster in particular is interested in today's weather in the North Atlantic Ocean which may well be Britain's weather tomorrow or the day after. Much information now comes in from ships of all kinds. Some are especially equipped ocean weather ships, like the British vessels (*Weather Observer*, *Weather Explorer*, *Weather*

Watcher, and *Weather Recorder*) which look after two positions in the eastern Atlantic: further west, American ships perform similar tasks. Many ocean-going Merchant ships send observations on a voluntary basis, as do numerous ships in home waters, including certain lightships. Ship observations are broadly similar to those made at land stations, but certain problems arise on a floating 'station,' and modifications are necessary in some of the equipment and techniques. On the other hand, special information such as water temperatures and state of sea can be provided.

Information from the Upper Air

The foregoing sections have been concerned with the necessary but limited view of the atmosphere derived from surface observations. Weather is a three-dimensional phenomenon and the answers to many problems of meteorology lie in conditions in the upper atmosphere. There have been great advances in the investigation of the higher levels since the early hazardous explorations by manned balloons were undertaken in the 1860s. Today, aircraft ascents and 'radio-soundings' are the two main sources of upper air data.

Spiralling ascents by specially equipped aircraft were of great importance just before, and early in, the Second World War. The observer recorded pressure, temperature and humidity at various heights, and could also report on conditions of cloud, icing, weather, etc. Aircraft also undertook reconnaissances over wide areas, and reports every 50 miles were radioed to base. Today, aircraft flights are not the chief method of upper air observation, although vertical ascents and reconnaissances (often combined, with an ascent or descent at the terminal points of the 'legs' of the course) are still carried out.

Most important today is the radio-sounding (*radio-sonde*) method. The equipment consists of a balloon (some six feet in diameter) which carries up recording apparatus—an aneroid capsule for pressure, a bi-metallic coil for temperature, and a strip of goldbeater's skin sensitive to humidity—and also a radio transmitter which sends the information in the form of a musical note to the ground receiving station. Each recording unit is connected in turn, by means of a switch operated by a small windmill, to the transmitting unit. The result of changes in pressure, temperature or humidity is to vary the pitch of the note transmitted. At the ground station, the signal is received and applied to a cathode ray oscillograph. The observer has by him an oscillator similar to that in the radio-sounding transmitter, and this he tunes until the note it produces is of exactly the same pitch as the one transmitted: when both notes are in tune, the C.R.O. shows a simple stationary pattern. The frequency of the note is read off the oscillator dial and this figure converted by means of prepared

graphs into familiar pressure, temperature or humidity values: each element 'comes round' about once a minute. Meanwhile, at a certain limiting height (which is of the order of 60,000 feet) the much distended balloon bursts and the equipment then returns gently to earth by means of an attached parachute. Many sets are recovered.

Balloons also provide a means of measuring upper winds. One method employs a 'pilot balloon' with a known rate of ascent (this is controlled by the amount of gas): the balloon is released and 'followed' through a specially designed theodolite, from which the bearing and elevation are read off every minute. The path of the balloon depends on the winds at various levels, and both directions and velocities can be computed from the figures. Cloud and fog, however, limit the value of the pilot balloon method. The same purpose can be served by the radio-sounding balloon, which has the advantage that it need not be visible: the transmission of signals betrays its position, which can be determined by three radio-direction-finding stations. The height of the balloon can be calculated from the transmitted data, though radio methods of measuring elevation as well as bearings have been developed in the United States. The most recent techniques of upper wind measurement have made use of radar. No transmitter is required, simply a less costly reflector (usually of metallised paper) which is carried by the ascending balloon. A radar signal is emitted from a powerful ground transmitter, and the reflected echo is picked up by a receiver with directional aerials. Only one ground station is needed in this 'radar-theodolite' method.

Radio and radar techniques are now being applied increasingly to problems of meteorological observing. For example, lightning activity is located by radio: the flashes that produce 'atmospherics' in our ordinary wireless sets act as radio transmissions and can be registered on the cathode-ray-oscillograph. The British SFERIC service (begun during the war) 'fixes' thunderstorms up to about 1,500 miles away by means of four direction-finding stations. This provides valuable information for the forecaster.

Radar is now being employed to locate cloud and rain, since it is now known that water droplets exceeding a certain size produce echoes which appear as patches on the radar screen. In this way the rain-bearing clouds of thunderstorms and frontal systems can be seen and followed. Two types of equipment are used: the Plan Position Indicator gives the bearing and distance of the 'radar cloud' from the station: the Height-Range Finder gives the height and distance. This method, already successfully used in research into thunderstorm structure, has great possibilities for forecasting, especially if the range of operation (now about 70 miles) can be extended.

ASSEMBLING THE INFORMATION

Between the initial gathering of meteorological information and the final use of it lies a series of intermediate steps that involve a high degree of organisation. Little more need be said here concerning reports intended purely for climatological or research purposes: these are directed into quieter backwaters of our meteorological organisation (the Climatological or appropriate research branch of the Meteorological Office). What follows in this chapter concerns the swift sequence of events from observation to forecast, and, while the details given refer particularly to the British meteorological service, the arrangements are broadly similar in most national organisations.

The Organisation of a Modern Meteorological Service

At all synoptic and auxiliary reporting stations the observations described in the previous chapter are carried out at standardised times, using standardised instruments and following standardised procedures. The reports are entered in a standard form of observation book which serves as a permanent record. For the urgent needs of forecasting, this information must be made available as quickly as possible wherever it is required. A rapid system of communications is thus an essential requirement of an efficient meteorological service. As the first step in the exchange of information, the reports are converted into a form of coded message which can be easily and quickly communicated.

There was no method of rapid exchange of weather messages before the advent of wireless telegraphy (the first telegraphed weather reports were published in the *Daily News* in 1848). Today, wireless transmission in Morse code remains the system of communication used by ships and aircraft and between more distant countries. Within the British Isles the main internal means of communication is by land line (teleprinter), although wireless is available in an emergency. Some 190 British stations are currently linked by the teleprinter network, and lines now extend to the nearer Continental countries. The latest development in meteorological communications is photo-telegraphy or facsimile transmission (known as FAX), an ingenious process by which completed weather

charts are 'broadcast' by one station and 'received' at others. This method, already extensively used in the United States and Canada, and (experimentally) in Britain, may eventually revolutionise the exchange of meteorological data.

At present, however, information is received by the forecaster in the presentation of this information in some way that allows an analysis of the current weather situation. There are standardised procedures for the form of coded messages on the teleprinter. The next stage is the replotting of surface data on prepared outline maps: such a map when completed provides a general view (or synopsis) of the weather over a wide area, and is, in fact, known as a 'synoptic chart.' The upper air data similarly find their way on to aerological charts and diagrams of various kinds. The climax of all this activity is the preparation of the forecast itself. At this stage, the organisation, having provided the available evidence, recedes into the background, and the individual forecaster comes into the centre of the picture. All now depends on his reading of the facts, on his skilful application of certain trusted techniques, and on his general knowledge of weather habits.

No national meteorological service can be self-sufficient. The weather recognises no political frontiers and it is not surprising that in normal times few services are as effectively internationalised as the exchange of meteorological information. Every country makes its own weather reports available for the common use. This involves a high degree of co-operation between the individual state services, which are in fact associated in a World Meteorological Organisation. The functions of this body are to ensure uniformity in observing, reporting and plotting procedures, and to co-ordinate arrangements for the exchange of information.

The 'Met. Office'

We may clothe this bare outline with some detail by viewing the structure of the organisation from bottom and top respectively, *i.e.* at an ordinary satellite meteorological office, and then at the Central Forecasting Office. The typical 'met. office' combines the functions of synoptic reporting station and forecasting unit, and is usually located on an airfield, R.A.F. or civil. If we visit such an office our first visual impression is likely to be one of charts, which cover the working benches and most of the wall space. At the same time we become aware of the constant chatter of the teleprinters (housed usually in an adjoining room). The office is manned normally by a forecaster and one or more assistants.

Concentrating our attention first on the assistant, we note that he (or she) becomes galvanised into activity at approximately ten minutes before the hour: this is the time for the hourly observations or 'ob.' The visit to



PLATE 1. The meteorological station at St. Crispin's School, Wokingham. Much of the equipment, including the scale model of the 'Skylon' carrying the anemometer and wind-vane, was made by the pupils



PLATE II. A miniature cloud atlas. Cumulus and stratocumulus, layered medium cloud and distant cirrostratus and cirrus can be seen in this R.A.F. photograph

the instruments, outdoor and indoor, and the expert appraisal of the sky are carried out at great speed. Still before the hour has struck, the 'ob' is duly recorded in the register and also condensed into a coded message of the following form (with meanings indicated):—

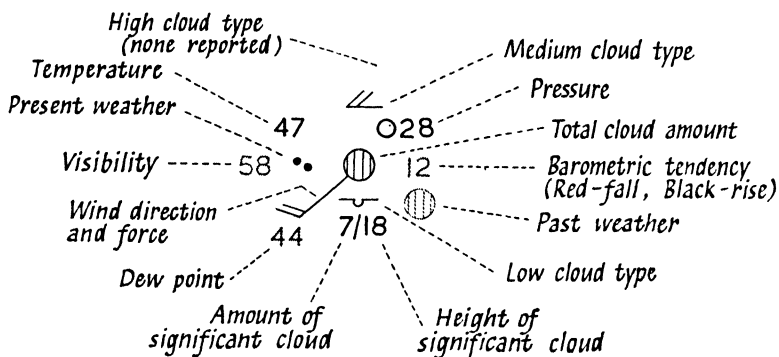
7 7 2 Station number	8 2 2 1 8 8-eighths cloud Wind direction: 220° Wind speed: 18 knots	5 8 6 1 2 Visibility: 8 Km Present weather: slight rain Past weather: cloudy	0 2 8 4 7 Pressure: 1002.8 mb Temperature: 47° F.
7 5 4 2 - 7-eighths low cloud Form of low cloud: Sc Height of base: 1,000–2,000 ft. Medium cloud: As No high cloud	4 4 7 1 2 Dew point: 44° F. Barometer falling Fall of 1.2 mb in past 3 hours	8 7 6 1 8 Indicator figure 7-eighths significant cloud Form of significant cloud: Sc Height of this cloud: 1,800 ft.	

At a signal, the 'ob' is transmitted on the teleprinter. The assistant types the message on the keyboard of the machine, and we may watch the five-figure groups appearing on the unwinding roll of teleprinter paper. They are also appearing simultaneously on the teleprinter at the Group Station which controls this particular satellite and a number of others, and acts as a regional centre for the collection and distribution of reports. If we continue watching after the 'ob' has been sent, we observe the teleprinter chattering on, as the messages from the other satellites in the group are duly sent and acknowledged.

Every other Group office in the country is at this moment similarly receiving the 'obs' from its respective satellite stations. Once collected, the messages are passed from the regional centres to the national centre, that is, the Central Forecasting Office at Dunstable. This task is completed by 10 minutes past the hour. At that moment, every station on the teleprinter network is switched via its Group centre to Dunstable, and the

entire series of now classified messages is relayed directly to all concerned. It takes 30 minutes for the British Isles and near Continental data to be transmitted: these are followed by other foreign reports, ships' observations, upper air data, and sundry less regular communications. At 5 minutes before the next hour, the transmission ceases: shortly, the cycle will begin anew.

The long strips of teleprinter paper contain the information which keeps the forecaster in touch with current weather conditions from hour to hour. At certain times (generally at midnight, 6 a.m., noon, and 6 p.m., and sometimes at intermediate hours), synoptic charts are plotted. These may vary from small-scale charts of Western Europe and the eastern North Atlantic to large-scale charts of the British Isles alone: much depends on the forecasting responsibilities of the office and on the particular weather situation. If we glance over the assistant's shoulder as he plots, we see that he is rapidly translating coded teleprinter messages into clusters of symbols and figures on the chart. The plotting technique is based on a standardised 'station model' and on internationally agreed codes and symbols. The stations are located on the chart by circles each identified by the station number, and the information is plotted in and around these circles. The coded message given on page 33 would be plotted thus (the four dotted features being given in red ink):—



The completed task of the assistant represents the raw material on which the forecaster must work. Rapidly he undertakes his analysis: his pencil weaves among the plotted 'obs,' drawing in the isobars (lines of equal pressure) which indicate the pressure distribution over the area: here, perhaps, a coloured line traces the position of a front: there a tentative arrow suggests a probable direction of movement. All the time, the forecaster refers back to preceding charts, for his analysis must be consistent with the past history of the developing situation. Several examples of fully analysed charts are given in Chapters 5 and 6.

The assistant meanwhile deals with other information. The data of balloon soundings are represented on a complex diagram (the 'tephigram') which depicts at a glance the variations of temperature and humidity with height, and is a forecasting tool of great value. Upper-air charts are plotted which show wind directions and speeds at heights up to 40,000 feet: essential information in these days of high-flying aircraft. Of especial interest perhaps to the visitor is the 'prebaratic'—a forecast chart of surface conditions 24 hours ahead, issued at 6-hourly intervals by Dunstable.

All such evidence considered, the forecaster prepares his story. Forecasts take various forms, depending on the requirements of the airfield. A local area forecast is generally issued as a matter of routine at least once daily: this takes care of local short-distance flying. Here the forecaster needs to know well the weather peculiarities of his own locality. Longer flights mean that special route forecasts are required: here again, obviously enough, the task of a forecaster at an R.A.F. fighter unit differs considerably from that of a forecaster concerned perhaps with trans-Atlantic flights at a civil airport. Often at Service stations there are periodic briefings for the crews. The duties of the forecaster do not end here. Enquiries by telephone (there are several on the bench) and by personal callers (the 'met. office' is by tradition a hospitable place) must all be answered. Never far from the forecaster's elbow is the Enquiries Book in which all enquiries and replies are duly noted. Sometimes there are special forecasts or warnings to be made known to individuals or organisations on or off the station. The 'met. office' can be a very busy place, especially when bad weather is threatening.

Dunstable

The Central Forecasting Office (C.F.O. or simply 'Dunstable') occupies a modest group of single-storey buildings on the outskirts of that town. Here are performed a number of vital functions, of which we gave some hint in the previous section. In the first place, C.F.O. is the nerve centre of the communications network, charged with the day-and-night task of collecting and redistributing with the utmost haste all the data required for forecasting purposes. The teleprinter room at Dunstable, with its 80-odd machines, is impressive both to the eye and to the ear. The messages received from Group centres are recorded on tape and rebroadcast at high speed by an automatic transmitter. Rivalling the teleprinter room in size is the wireless reception room where data from foreign sources, ships and meteorological reconnaissance aircraft are received. There is also a wireless transmitting room, for, apart from its rôle in the British service, Dunstable is a sub-continental centre in the European meteorological

logical scheme and is responsible for relaying weather reports from the British Isles and certain neighbouring areas.

In the Forecast Room at Dunstable, meteorologists of long experience concentrate their attention on the basic analyses and prognostic charts which serve to guide forecasters at satellite stations in their more local and immediate problems. Four times daily, surface charts are plotted which encompass most of the Northern Hemisphere: at intermediate hours, the charts drawn cover most of Europe and the North Atlantic. A host of assistants is kept in full employment. Upper air conditions are analysed too over a wide area, and again forecast charts are prepared. These distributions are transmitted in the form of coded messages to all stations. Also regularly broadcast for the benefit of local forecasters are the 'General Synoptic Reviews' which give an appreciation of the current weather situation and the expected future developments.

In addition to those forecasts that move through the usual channels of the organisation, C.F.O. is responsible for providing weather information of various kinds for Government Departments, public services, shipping, and for the general public. Thus B.B.C. and Press forecasts originate at Dunstable. Warnings of special weather conditions—such as gale, fog or frost—are given out for general guidance. Certain organisations—for example, British Railways, the British Electricity Authority, the Gas undertakings, River Boards—are warned specifically of impending weather that may affect their operation. Many commercial undertakings too—like film production units concerned about sunshine—arrange to receive special forecasts for their particular purposes.

The Forecasting Research Division is also located at Dunstable, in fact in the new Napier Shaw Laboratory (named after a distinguished pioneer meteorologist and one-time Director of the Meteorological Office). Dunstable is the control station of the SFERIC service, and is concerned in experimental activities, like the transmitting of FAX. An adjacent printing plant produces the *Daily Weather Report* (of which more shortly) and many of the blank charts and forms required in the synoptic service.

Weather Information and the Public

The energies of our meteorological organisation are mainly devoted to meeting those specialised needs of which we have sought to give some notion. Readers of this book, however, may be particularly concerned with the meteorological information ordinarily and regularly available to those members of the public whose sole claim to attention is their interest in the weather.

Forecasts of a non-technical nature are issued to all newspapers and Press Agencies. The amount of information supplied by C.F.O. may how-

ever differ vastly from what the newspaper actually publishes. We should not blame Dunstable for the twenty-word statement devoted to weather in our morning or evening paper. In fact, no less than nine weather bulletins are supplied daily to the Press and three of these include forecasts of some detail for different parts of the country. Without doubt, the forecasts published by some newspapers are so condensed and generalised as to be practically useless. On the other hand, certain papers—which deserve our gratitude—print, in addition to detailed forecasts, Dunstable's 'prebaratic' for noon on the day of issue. We should remember that a morning paper prints a forecast issued at 9 p.m. the previous evening: it will sometimes happen that the forecast is out of date by the time it is read.

The B.B.C. weather forecasts are familiar enough. On the Home Services we need to take notice of the land area bulletins (which include regional forecasts) read immediately before the News broadcasts rather than of the often over-summarised forecasts read during the News itself. The shipping forecasts (broadcast since April 1956 only on the Light Programme on 1500 m.) are of a rather more technical nature and contain some synoptic details (see Chapter 11). The Television weather forecasts bring to the viewer the synoptic chart of the afternoon and a 'prebaratic' for the following day. This visual presentation of the weather situation and prospects, accompanied by a commentary delivered so aptly by the now-famous TV weather-men from the Meteorological Office at Kingsway, seems entirely successful: nothing else is doing as much at present to create a weather-conscious public.

However, the most rewarding B.B.C. weather broadcast, instituted in 1946, was the now suspended 'Airmet' service: this was a radio broadcast functioning during daylight hours, and including, hourly, statements of the weather situation and expected developments, actual weather reports from selected stations and informal talks by a forecaster. Although intended primarily for small airfields lacking meteorological offices, Airmet was invaluable to all who were interested in weather. Quite apart from its practical value, it was highly suitable for teaching purposes, and the hourly reports enabled the amateur to draw his own synoptic charts. In 1950, unfortunately, the service was suspended owing to shortage of suitable broadcast wavelengths.

Not all people are aware that they may obtain up-to-date information and forecasts by telephoning certain meteorological offices specially designated to deal with public enquiries. There is no cost to the enquirer beyond that of the call. The number of the nearest office may be found at the beginning of the local area telephone directory. Recently in the London area the automatic telephone weather service has come into operation, by which dialling WEA 2211 evokes a brief forecast (for the

London area) just as painlessly as dialling TIM calls forth the correct time. It is hoped to extend this service further afield.

The Daily Weather Report

For the amateur who likes to follow the march of weather events, for the teacher in school or university who wishes to use synoptic data for teaching, and for numerous research needs, the *Daily Weather Report* and the *Daily Aerological Record* are invaluable sources of information. Both publications can be supplied by post to any address on application to the Meteorological Office.

The *Daily Weather Report*, in its present form, consists of four pages. Two are covered with detailed weather reports (for 0000 and 0600 hr. G.M.T. of the day of issue, and 1200 and 1800 hr. G.M.T. of the day before) from 56 British stations and a number of North Atlantic ships. The other two pages carry four synoptic charts (for the same hours as the weather reports), of which one is a small-scale chart covering most of the Northern Hemisphere, and the others, of larger scale, covering Western Europe. We reproduce as Fig. 49 a map based on the Northern Hemisphere chart of a *Daily Weather Report*. Included also are brief statements of the general inference and further outlook and some explanation of the symbols and figures used on the charts. The *Daily Weather Report* is posted in the afternoon and arrives at most destinations next morning.

Although the *Daily Weather Report* reaches us a day late, it nevertheless offers an explanation of past weather and the background of current developments against which we may check both our own personal observations and whatever forecasts we see or hear. Since the loss of Airmet we can no longer continue the sequence with simple charts of our own construction. We can however use the station data of the *Report* to plot past charts in considerable detail. For this some knowledge is needed of the more common codes and symbols, and a selection of these is supplied in the Appendix. Reference to these and a little practice will make it easy to convert from the tabulated reports to station models like that shown on page 34.

The companion *Daily Aerological Record* may seem at this stage of more remote and academic interest, especially as it is issued about a week in arrears to allow for careful checking of data. It is a six-page report including observations from the nine British radio-sounding stations, the two ocean weather ships, occasional aircraft ascents, and upper-air charts for various levels. The radio soundings and aircraft data enable us to construct a temperature-height diagram—like that shown in Fig. 18—

which is a simple version of the forecaster's 'tephigram,' and which we shall learn to use in a later chapter.

The upper-air charts must seem even further removed from our view of the weather from the ground. Modern meteorology, however, pays increasing attention to the upper air, and we must return to these charts at a later stage. At present, sufficient to say that they consist of 'contour' charts, which depict the air flow at various levels, and 'thickness' charts, which show the distribution of warm and cold air within given layers of the atmosphere.

The *Daily Weather Report* and *Aerological Record* place some of the professional forecaster's tools in our hands. To enable us at least to understand how he uses them and perhaps to play with them ourselves, is the aim of the following chapters of this book.

THE CAUSES OF WEATHER

The atmosphere above us is in a state of ceaseless and sometimes violent activity. Vast bodies of air weighing many millions of tons are moved great distances over the face of the earth: horizontal wind speeds of 200 m.p.h. and vertical currents of 20 m.p.h. are often recorded: an average thunderstorm releases power equivalent to the combined output of several large electrical generating stations—or to that of ten atomic bombs: even the energy involved in the explosion of a hydrogen bomb is quite dwarfed by that dissipated in a tropical hurricane. Atmospheric activity requires an input of energy on a stupendous scale. The prime question in meteorology is that of the source and nature of the energy that powers the atmospheric machine.

The sun-worshippers of old and the sun-bathers of today are alike in recognising, by their homage, the importance of the sun as the benevolent donor of the light and warmth which permit life on the surface of the earth. There are no alternative sources of light and heat: the escape of heat from the earth's interior is so slow as to be insignificant: even the light and heat we obtain from burning coal or oil represent originally sun-derived supplies stored in the rocks over millions of years. To the meteorologist, light and heat are a form of energy—*radiant* energy—and it is solar energy which is made available to the earth's surface and atmosphere.

Energy cannot be destroyed but its various forms are interconvertible under certain conditions. Such transformations readily occur in the atmosphere, so that incoming solar energy may be converted, by indirect means, into the energy of motion of moving air masses or the electrical energy of lightning discharges. To state, then, that the sun is the source of atmospheric energy is to imply more than the explanation of our supplies of light and heat. We shall show that practically all the atmospheric processes and conditions which make up the weather depend ultimately on variations in the amount of solar energy received over different parts of the earth's surface. This is the first cause of weather.

Heat Transfer in the Atmosphere

We must at this stage attach some technical labels to methods of energy transference which are familiar enough in everyday life, but which have

special significance in meteorology. A poker inadvertently left stuck in the fire soon becomes hot at the handle end. This involves the process of *conduction*, that is, the transference of heat energy by actual contact from hotter to colder particles. The heat of a body is really the energy of vibration of its constituent molecules: conduction along the poker thus implies a movement of this energy from the more agitated molecules of the hot end to the more sluggish molecules of the handle. Conduction is most important in solid bodies, in which the molecules are densely packed and tightly held together. Liquids and especially gases, however, are poor conductors of heat. Conduction is therefore negligible as a means of heat transfer in the atmosphere generally, but it may become important in the lowest layers where air is in contact with the solid ground.

In liquids and gases, mass flow is possible, and heat acquired in one place may be transported to another by the actual movement of the fluid. This we call *convection*, and it is, of course, the process by which heat circulates in a domestic hot-water system. In the mixture of gases which constitutes the atmosphere, convection is a most important means of heat transfer. Meteorologists usually reserve the term for up-and-down movements of air, and distinguish horizontal movements, which may transport heat on a much larger scale, as *advection*. Associated with both convection and advection is the transfer of heat by means of the movement of water in the atmosphere: this we shall consider later in the chapter.

None of these methods of heat transfer, important as they are in the atmosphere, can be responsible for the transmission of energy from the sun to the earth. Conduction and convection require matter in which to operate. Practically all the 93 million miles which separate sun and earth is empty space, there being little atmosphere for the purpose even 100 miles above the earth's surface. One means only exists by which heat can be transferred through matterless space, and that is by *radiation*. This is again a familiar process in everyday experience. The ordinary open fire warms us, if at all, by radiation: clearly, convection is unimportant, for most of the warmed air vanishes regrettably up the chimney. A so-called 'radiator,' however, heats us comparatively little by radiation and much more by convection, by warming the adjacent air and setting up a circulation throughout the room.

The Nature of Radiation

We may infer from this that a hotter body radiates more intensely than a colder one. The 'radiator,' although in effect misnamed, nevertheless radiates to some extent, as, in fact, all bodies do, whatever their temperature. Physicists regard radiation as energy emissions in the form of waves (analogous to sound waves in air or ripples on water), which can be

conveniently described in terms of the speed of travel, the number of waves per second (*frequency*), and the distance between successive wave crests (*wave-length*). Certain phenomena we may not normally associate—light, heat, wireless waves, X-rays—all consist of waves of this kind, having the same velocity (186,000 miles per second), but differing in frequency and wave-length, and, most significantly for us, in the way they affect objects placed in their path. We have become familiar with the idea of wave-length from the dial of our wireless receiving set: wireless waves, which are very long, have no effect on us personally, but can be picked up by tuning a suitably constructed electrical circuit. If we imagine the dial of our wireless extended enormously in the direction of shorter wave-lengths, we would come eventually to a range within which radiation affected us physically by warming us: this is radiant energy which we perceive as heat. At still shorter wave-lengths, radiation not only warms us, but also affects our eyes as light. At the extreme short-wave end of our protracted wireless dial come emanations such as X-rays and cosmic rays.

Meteorologists are concerned with the relatively long waves that are effective as heat and the shorter waves that are effective as heat plus light. The wave-length at which a body radiates depends on its temperature: the hotter the body, the shorter the wave-length of its radiation. The glowing open fire emits short-wave radiation which is visible as well as warming, whereas the 'radiator,' with a much cooler surface, sends out long-wave, invisible heat radiation only. Similarly, the sun, which has a surface temperature of nearly 11,000° F., sends out short-wave radiation which provides warmth, and part of which is visible as sunlight: in contrast, the earth's surface and atmosphere, with an average temperature of about 60° F., send out only long, invisible heat waves. In addition, solar radiation is much more intense than earth (terrestrial) radiation.

Ordinary sunlight may be split up into various colours which constitute what is known as the visible spectrum. These colours—which some of us remember by the mnemonic ROYGBIV—differ physically only in their wave-lengths, red being at the long-wave end of the spectrum and violet at the short. Solar radiation is most intense within the visible range, but actually includes invisible waves shorter than the violet (the 'ultra-violet') and longer than the red (the 'infra-red'). Terrestrial radiation is of wave-lengths entirely within the infra-red. Thus we see that the spectrum is not concerned only with colours but covers all wave-lengths whether visible or not: in fact the entire range from very long radio waves to ultra-short cosmic rays is spoken of as the *electro-magnetic spectrum*. Some of these facts concerning radiation are summarised in Fig. 7.

What happens to an object placed in the path of a radiation beam, whatever its wave-length, depends on the physical properties of the object.

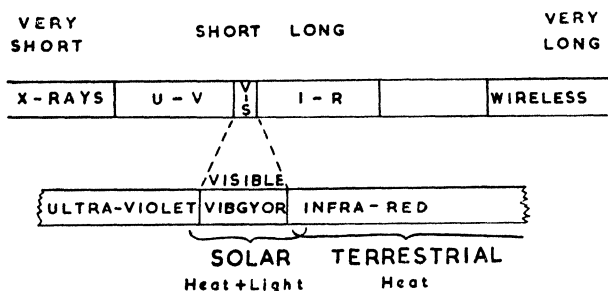


Fig. 7. The extended wireless dial (electro-magnetic spectrum)

A body may *transmit* radiation, just as glass allows the passage of light, in which case neither the radiation nor the transmitting body are affected in any way. Other bodies *reflect* radiation, as a mirror reflects light: in this case, the reflecting body again remains unaffected, and the radiation is altered only in the changed direction of its path. Still other bodies *absorb* radiation, which means that the radiant energy becomes converted into the internal (heat) energy of the absorbing body. As a result of this energy transformation, the temperature of the absorbing body is increased. It is by this process that the surface of the earth is warmed by the radiant energy from the sun. But a body warmed by absorption of radiation is not a passive recipient, for it must itself radiate, and does so at a wave-length appropriate to the temperature it acquires.

In summary, the sun emits short-wave radiation, including the visible wave-lengths. Practically all of this radiant energy is wasted from our point of view, but the planet Earth intercepts and absorbs a tiny part (less than a millionth) of the total, is warmed in the process, and itself radiates in the long-wave (infra-red) part of the spectrum. The earth's surface as a whole is both an efficient absorber and emitter of radiation. Within its surrounding atmosphere however, the response to radiation is more complex, and some of the constituent gases respond differently to radiation of different wave-lengths. This is important in meteorology, both as regards the radiation exchange of the earth as a whole and day-to-day differences in weather over a particular spot.

The Response of the Atmosphere to Solar Radiation

Physicists measure solar radiation indirectly by means of the *pyrheliometer*, an instrument that records a rapid increase of temperature proportionate to the radiation it absorbs. By measurements taken from high mountain observatories under the clearest conditions, and after certain adjustments, an average value has been obtained for the rate of

energy received at a point just outside the earth's atmosphere. This is known as the *solar constant* (it will not help us to know the exact value). Of more concern, however, to the meteorologist is the solar radiation received at the earth's surface after having passed through the atmosphere: this we call *insolation*. At any point on the earth's surface, the amount of insolation clearly depends on the solar constant (which actually fluctuates slightly), the angle of the sun's rays (this varies with time of day and with season: the more oblique the ray the less the insolation), and on the effect of the atmosphere.

The atmosphere consists of a mixture of gases, of which some (like oxygen and nitrogen) are present everywhere in fixed proportion, while others vary in concentration with time and place: these include water vapour and carbon dioxide, which are mainly concentrated in the lower layers, and ozone, which is most important at great heights. The atmosphere also contains various solid impurities—dust, smoke, salts—which are found mainly in the lower layers. Each of these constituents reduces incoming solar radiation in various ways.

A small part of the sun's energy is absorbed by certain gases in the atmosphere, mainly oxygen and ozone at high levels and a little by carbon dioxide and water vapour nearer the earth's surface. A substantial portion is reflected by water droplets, particularly from the upper surfaces of clouds, and largely lost into space. Thick clouds reflect nearly four-fifths of the radiation falling upon them, thin broken clouds much less. The earth's surface (especially where it is snow-covered) also contributes a little to the loss by reflection, but the major factor is the amount of cloud. Taking the earth as a whole, about half its surface on average is cloud-covered, so that half the possible reflecting cloud top is present: this means that, over all, about two-fifths of the incoming solar radiation is reflected away. This average value (40 per cent) is sometimes referred to as the *reflectivity* or *albedo* of the earth.

Particles like those that make up dust or smoke haze, which are smaller than water droplets (in fact, having a diameter less than the wave-length of the radiation), scatter rather than reflect. *Scattering* resembles ordinary reflection in sending off the radiation, unaltered, in all directions, but with the difference that it is more effective for the shorter wave-lengths. Thus the bluer light is more easily scattered than the longer red rays. Some of the scattered light is lost into space, but some is directed downwards as a diffuse radiation from the sky in general. A cloudless sky appears blue because we see it by diffuse light which has undergone much scattering and is rich in the blue rays. On the other hand, when we see the sun through a smoke haze, or through a great thickness of atmosphere when it is low in the sky, it looks reddish: its light has been robbed of much of its blue and is correspondingly richer in the red rays.

Depleted in these various ways, less than half the incoming solar radiation on average survives its passage through the atmosphere, and is finally received and absorbed by the earth's surface. Equally noteworthy is the comparatively small proportion of the available solar energy that is absorbed by the atmosphere—about one-seventh only, and that mainly at great heights. We express this in the statement that the atmosphere as a whole is largely *transparent* to solar radiation. It follows from this that the atmosphere is warmed very little by direct solar energy.

The Response of the Atmosphere to Terrestrial Radiation

The atmosphere responds quite differently to the long-wave radiation emitted by the earth's surface. While it is not far from transparent to solar radiation, the atmosphere absorbs strongly in the wave-lengths of terrestrial radiation. This means that while it is true that the sun is the ultimate source of its energy, the atmosphere is heated directly from the earth's surface, not from above but from below. This is of first importance not only for the atmosphere as a whole but in connection with individual air masses, which acquire their temperature characteristics from underlying surfaces.

The water vapour of the atmosphere is by far the most important absorber of long-wave radiation (although carbon dioxide and ozone also absorb in these wave-lengths). This absorption continues even in perfectly clear skies, for there is always water vapour present at least in the lower layers of the atmosphere. If there are clouds, the absorption is even more effective: cloud surfaces, which reflect so much of the solar radiation falling upon them, reflect hardly at all in these longer wave-lengths. Actually, water does not absorb all the wave-lengths of terrestrial radiation: it is transparent to some, so that there is always a certain loss of radiation directly into space. This, however, amounts, on average, to only about a tenth of the total.

The remainder is absorbed by the atmosphere. The atmosphere, in its turn, being warmed by this absorption, acquires a temperature which is of the same order as that of the earth's surface (though lower), and must itself radiate in the long wave-lengths. Part of this radiation is directed upwards and lost into space, but part is emitted downwards to the earth's surface. Here it is again absorbed, transformed into heat, and re-radiated. Radiation and counter-radiation thus proceed continuously between earth and atmosphere.

The Heat Balance of the Earth and Atmosphere

The earth's surface and atmosphere are, on the long view, growing neither hotter nor colder. This means that all the energy derived from

the sun and converted into heat must eventually be lost by terrestrial radiation and other means: an over-all balance between incoming and outgoing heat energy must be achieved. In presenting a rough account of these complex energy transactions, we shall avoid further complexity by considering average conditions over the earth and atmosphere as a whole. It must also be a very approximate and tentative statement, as the values assigned to the various items of the energy exchange are merely estimates based on measurements and computations concerning which not all authorities agree and which are constantly under review.

The processes involved in maintaining the heat balance are depicted graphically in Fig. 8. The left-hand side of the diagram is concerned with

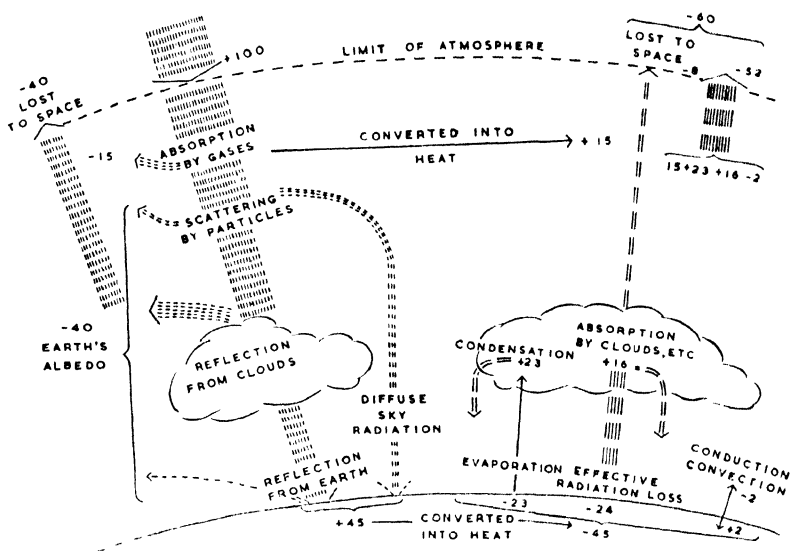


Fig. 8. The heat balance of the earth and atmosphere

incoming short-wave radiation, and shows that of 100 units of solar energy arriving at the outer limit of the atmosphere, depletion by the various means already discussed allows only about 45 to reach the earth's surface. Of this about one-third arrives as diffuse sky radiation.

The right-hand side of the diagram shows how the energy absorbed by earth and atmosphere is eliminated, thus preserving the heat balance. This involves more than one method of heat transfer. Of the radiation emitted by the earth's surface, the major part is absorbed and returned by the atmosphere: however, the effective radiation loss—the excess of outgoing radiation over that received—eliminates over half the 45 units absorbed from the sun. Of this terrestrial radiation, a third (8 units) is

lost directly into space and the rest (16 units) re-absorbed by the atmosphere. Almost as much heat (23 units) is lost to the earth's surface by evaporation, carried upwards in water vapour by convection currents, and later yielded to the atmosphere by condensation: the mechanisms of these processes we shall examine shortly. Conduction and convection in itself play only a small role in the effective heat exchange: the movement of heat may be upwards or downwards according to circumstances, and most authorities agree that the net transfer is from atmosphere to earth, but the amount is small (perhaps 2 units). This means that 52 units in all (15 absorbed directly from solar radiation, plus 16 absorbed from the earth, plus 23 from condensation, less 2 returned to the earth) remain to be radiated by the atmosphere back into space. This completes the balance.

It remains to draw attention once again to the selective response of the atmosphere to the radiation it receives. Of the incoming solar radiation, some 45 per cent is allowed through unhindered to be absorbed at the ground: but of the outgoing terrestrial radiation, only 8 per cent (in comparable units) is allowed to escape directly to space, the rest being absorbed by the atmosphere and in part returned to the earth. It becomes clear that the earth receives an extra ration of heat owing to the protective function—sometimes known as the Greenhouse Effect—of the atmosphere. It has been calculated that if there were no atmosphere, the average temperature of the earth's surface would be not 60° F. but something nearer 15° F., and the daily and seasonal extremes of temperature would be much greater than they are at present.

The Vertical Distribution of Temperature

The observed variations of temperature in the atmosphere, both with altitude—as shown by the evidence of radio soundings—and horizontally over the earth's surface—as established by the records of climatological stations—are the net result of the complex interplay of the various factors in the heat balance, modified by certain disturbing influences of a geographical nature. We have illustrated some of the features of the vertical distribution of temperature by plotting on a temperature-height diagram (Fig. 9) a number of radio sounding ascents chosen at random from the *Daily Aerological Record*.

These ascents, made at different times of day and year and in different kinds of air, are nevertheless strikingly similar in their general appearance on the diagram. These common features enable us to recognise at least three layers of the atmosphere, distinguished by the slope of the curve, which of course represents the way temperature varies with height. The greater part of each curve shows a fall of temperature with height:

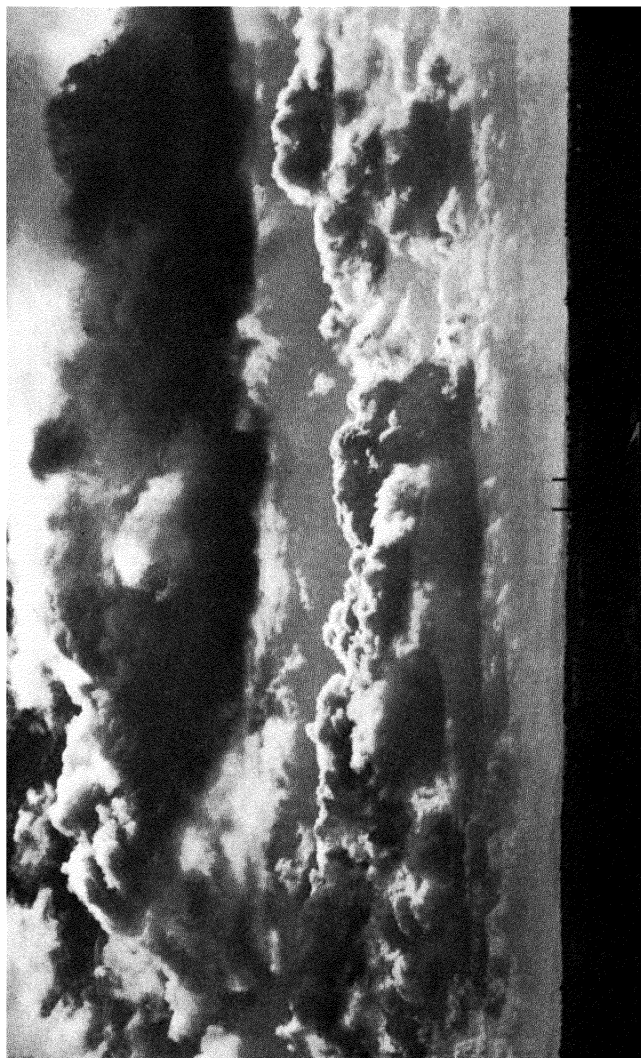


PLATE III. Polar maritime weather: an afternoon sky laden with cumulus clouds



PLATE IV. Polar maritime weather: the sun forces its way through heavy clouds after a shower

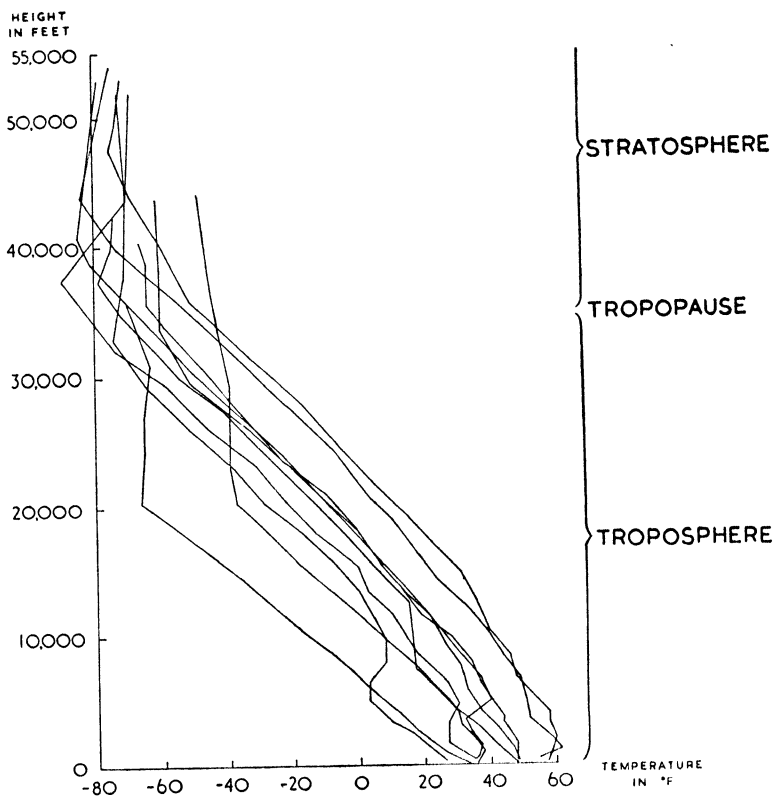


Fig. 9. Some typical radio-soundings

meteorologists apply the term *lapse rate* to the rate at which this decrease occurs and express it as a fall of so many degrees per 1,000 feet. A negative lapse rate, rather perversely, implies an increase of temperature with height: this is more often known as a *temperature inversion*. A layer within which temperature remains the same is called an *isothermal layer*.

In these terms, we can distinguish a lower atmospheric layer in which temperature decreases generally with elevation, and which is known as the *troposphere*: and an upper layer which is not far from isothermal or with a tendency towards a negative lapse rate, which is called the *stratosphere*. The level at which the lapse rate changes has been christened the *tropopause*: sometimes the transition is sharp, at others it occurs over a zone of some thickness. The average height of the tropopause over this country is around 30,000 to 35,000 feet, but, as the curves of Fig. 9 show, it fluctuates between wider limits. Many authorities separate out the lowest few thousand feet of the troposphere as a distinct layer. We can

see from the diagram that this seems to be justified, for in the lower troposphere, the various curves differ most: in particular, temperature inversions are most common in this layer adjacent to the earth's surface. In contrast, the middle and upper troposphere sections look remarkably similar from curve to curve. The average lapse rate in the troposphere as a whole, calculated from a large number of temperature soundings, is about $3\frac{1}{2}^{\circ}$ F. per 1,000 feet.

The thermal structure of the atmosphere, as revealed by Fig. 9, is roughly what we should expect from the nature of the heat balance. In general, the earth's surface is the warmest region: the atmosphere, being warmed from below, but losing heat both upwards and downwards, must be cooler on the whole than the earth's surface: and it seems entirely reasonable that those parts of the atmosphere nearer the ground should be warmer than those further away.

The exact slope of the typical temperature-height curve has not yet been explained to the entire satisfaction of the theoretical meteorologist. Computations have been made to show that something like the observed vertical distribution of temperature is necessary if heat is to escape from the atmosphere into space by radiation: the net flow of radiation must be from warmer to colder regions. But more than radiation is involved. Heat exchange by radiation alone would apparently result in a much greater concentration of heat in the lower atmosphere layers than is consistent with observations. It seems clear that the excess heat is carried upwards by convection currents which cause an effective stirring of a certain thickness of atmosphere and produces, by means which we must consider later, a uniform temperature lapse rate within it. Convection thus plays an essential role in the vertical distribution of temperature, and, as we shall see, is an important element in the day-to-day weather we experience.

The tropopause is visualised as the maximum height to which these convection currents penetrate, and therefore the effective upper limit of the influence of the earth on the temperature of the atmosphere. Supporting this is the fact that the tropopause is highest (about 55,000 feet) and the troposphere therefore thickest, over tropical regions, where surface heating is most intense and convection currents able to extend to greatest heights: whereas over the poles, where the least energy is available for convectional activity, the tropopause is lowest (about 25,000 feet) and the troposphere correspondingly thinnest.

The most up-to-date evidence shows that the stratosphere is not, as was once thought, a region of still and uniform conditions. The increase in temperature in the lower stratosphere shown in some of the curves of Fig. 9 is due to strong absorption of solar radiation by ozone, which exists in fairly high concentrations at these levels. The ozone concentration itself

varies with latitude, being greater above the polar regions than above the equator: because of this the stratosphere is warmer above high latitudes than above low. At elevations beyond the present reach of radiosonde balloons, indirect evidence suggests that very much higher temperatures exist even than those experienced at the earth's surface. But these speculations are far outside the scope of everyday meteorology and therefore of this book. We know that vertical movements take place in the stratosphere, though they are probably gentler than those observed in the troposphere, and it is certainly a region of sometimes strong horizontal winds. However, the stratosphere is extremely dry, and since weather as we know it is much associated with moisture, there is support for the popular notion that the stratosphere is 'above the weather.' Yet there may be some connection between conditions in the stratosphere and the weather we experience at the base of the troposphere. This is one of the many open questions of meteorology.

The Horizontal Distribution of Temperature

When we turn our attention to the distribution of temperature horizontally over the surface of the earth, we find that there are fewer unknowns and the pattern is more clearly understood. The major factor is, of course, the variation of insolation with latitude: the low latitudes, where the mid-day sun is always high and insolation strong, are in general the warmest, and temperatures decrease towards the polar regions where the rays from the low sun fall obliquely and have little intensity. Fig. 10 shows the computed variations with latitude of incoming solar and outgoing terrestrial radiation. As we should expect, incoming radiation exceeds outgoing in low latitudes, while the reverse is true of middle and high latitudes. The two curves cross at a latitude of 35° , which in terms of area, divides a hemisphere into halves. It would appear that while an over-all heat balance is maintained for the earth as a whole, there is a definite unbalance in the heat exchange at different latitudes.

We are entitled to ask why, under these circumstances, the low latitudes are not becoming steadily warmer and the high latitudes steadily colder. The answer can be only that some mechanism exists to transfer the excess heat from low to high latitudes. We shall see in Chapter 7 that this mechanism is the horizontal movement (*advection*) of air from one latitude to another, which combined with vertical movements and complicated by the rotation of the earth, constitutes the general circulation of the atmosphere. In this way, the wind systems, and the consequent transportation of weather from place to place by advection, result from the unequal heating of the earth's surface.

The temperature conditions at a given spot, however, depend on more

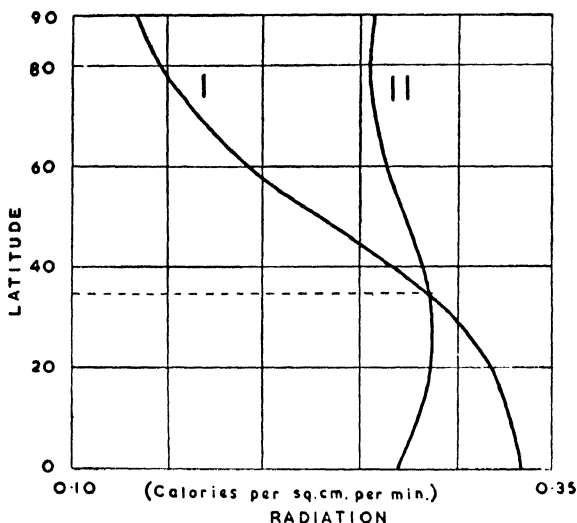


Fig. 10. The unequal heating of the earth's surface. Curve I, incoming solar radiation. Curve II, outgoing terrestrial radiation

than the radiation exchange. One of the most important factors is the nature of the surface on which the insolation initially falls: and of the various types of surface, those that contrast most in their response to insolation are land and water surfaces respectively. The physicists tell us that the average *specific heat* of the solid ground is about one-third of that of water. This means in ordinary language that it needs three times as much heat to raise the temperature of a quantity of water by so many degrees as is required to heat the same quantity of solid earth by the same number of degrees. More significantly, the solar radiation which falls on the land is immediately absorbed in a very thin surface layer, from which conduction downwards proceeds very slowly and to a depth of a few inches only: in short, the heating effect is concentrated in a thin layer, and the rise of temperature is therefore considerable. On the other hand, radiation penetrates relatively deeply into water, and in further contrast to the solid ground, the vertical mixing and stirring which are possible in a fluid spread the heating effect over a considerable depth. It follows from this that the rise of temperature at any point is extremely small.

The result of these contrasted properties is that water responds only slightly, but land considerably, to the same amount of solar heating. Similarly at night when outgoing radiation dominates, land temperatures tend to drop sharply, water temperatures very little. Since the temperature of the lowest air layer adapts itself rapidly to that of the underlying surface, this means that the temperature variation between day and night is

negligible above the sea (generally less than 1° F. except over shallow coastal waters) while over land it is often 30° or more.

In addition to this contrast in the diurnal temperature regimes, the seasonal variation of temperature over the sea is much less than that over the land. The slow response of the oceans to the heating and cooling processes also results in a substantial seasonal lag in temperature, so that the month of maximum temperature round Britain is August or September and the minimum is reached in February or March, in each case a month or two later than the corresponding period over land. The heat distributed throughout a substantial depth of water is not lost, but merely stored, and may be transported in the circulation of ocean waters to be later liberated in colder regions. This difference between land and water profoundly modifies the general atmosphere circulation, and introduces important differences in the weather and climate over the oceans on the one hand and over land on the other.

Temperature and the State of the Atmosphere

We can conveniently consider at this stage the effect of varying atmospheric conditions on day-to-day temperature fluctuations at a given spot. With clear day-time skies, the depletion of incoming solar energy is at its least, and probably two-thirds of the possible radiation is absorbed at the earth's surface. Over land areas in the absence of strong winds (which, as we shall see, are unfavourable to temperature extremes), the highest maximum temperatures possible for the region, the time of year, and the type of air are reached. Similarly, a cloudless night will allow the greatest loss of heat by terrestrial radiation, and low minimum temperatures must therefore be expected (again assuming calm conditions). This is a weather situation most worrying to farmers and horticulturalists, especially in the marginal temperature conditions of spring and autumn when a low minimum may mean a damaging frost.

Under clear conditions, the temperature rises from a minimum about dawn to reach a maximum around 2 p.m., that is, two hours beyond the time of maximum insolation (noon). This lag between temperature and sun is due to the fact that up to 2 p.m. the gain of heat by insolation exceeds radiation losses, but thereafter the balance tips the other way, outgoing radiation overtakes the diminishing insolation and the temperature falls.

With cloudy skies, incoming solar radiation is cut down to a minimum: most of it is reflected from the cloud tops, and only about one-fifth is absorbed at the earth's surface. Under these conditions, temperatures rise only slightly and low maxima are recorded. Cloudiness is however an advantage during the night when outgoing terrestrial radiation is reduced

by absorption by, and back-radiation from, the cloud cover. The effectiveness of this reduction depends largely on the height of the cloud: a complete cover of low cloud reduces the nocturnal cooling to about an eighth of what it would be with a cloudless sky, but with very high cloud there is almost as much cooling as under quite clear conditions. With this exception, cloudiness at night tends to maintain relatively high minimum temperatures. It follows from these considerations that clear skies both day and night favour temperature extremes, that is, a large diurnal variation, while continuous cloudiness makes for a small diurnal temperature range. Other combinations are of course possible; we can readily see that clear days followed by cloudy nights give us the best possible temperature conditions, while with cloudy days and clear nights we experience the terrestrial heat exchange at its least advantageous.

Cloudiness not only reduces the diurnal range but may also blur the times of the temperature extremes. For example, a spread of heavy cloud during the late morning may advance the time of maximum temperature to midday or earlier. The replacement of one air mass by another, with the accompanying change in temperature (as well as other air mass properties) may completely upset the normal diurnal march of temperature. The daily variation depends also on the wind strength. Only still or gently moving air is subject to great temperature extremes, cloud permitting: strong winds (see Chapter 4) diffuse the effects of heating and cooling upwards through a fair thickness of atmosphere, which makes for a reduced range at the surface. Thermograph traces illustrating some of these conditions are reproduced in Fig. 11.

Water in the Atmosphere

The existence of water in the atmosphere is a major element in any review of the basic causes of weather, and, as already indicated, it plays an important rôle in the atmospheric heat balance. Water is present in the atmosphere in all three states of matter, gaseous, liquid and solid, in other words, as vapour, droplets and ice particles. Whatever its condition, moisture reaches the atmosphere from underlying surfaces by the process of *evaporation*: this takes place readily not only from bodies of water but also from soil and vegetation-clad surfaces. It is worth remembering that individual air masses derive their moisture content from underlying surfaces. Evaporation signifies the change from the liquid to the vapour state: sometimes, as when snow patches disappear without first melting, the change occurs directly from solid to vapour: this is known as *sublimation*, less important than evaporation as a means of supplying water to the atmosphere.

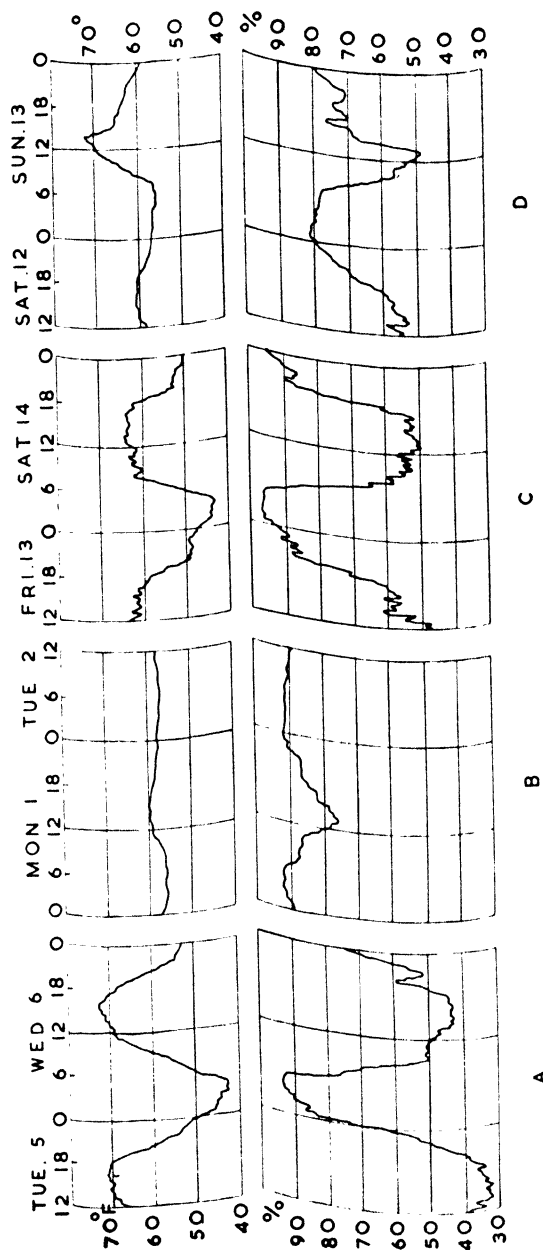


Fig. 11. Some typical thermograph and hygrograph traces. (a) Clear skies night and day. (b) Cloudy night and day. (c) Clear night followed by cloudy day. (d) Cloudy night followed by clear day.

Evaporation

Everyday experience teaches us the various factors which affect the rate of evaporation. The higher the temperature of the evaporating surface, the more evaporation will occur: we apply heat in order to boil a kettle of water, and evaporation occurs as an inevitable though unintended by-product of this process. The amount of evaporation also depends on the moisture content or humidity of the air above the surface, for air has only a limited capacity for holding water in the vapour state: in still, moist, muggy air, wet streets or damp washing take a long time to dry. A third important factor is the wind strength. With strong winds and the ceaseless eddying and swirling of air which accompany them, there is constant replacement of the air above the evaporating surface, so that successive volumes of fresh air arrive to take up their quota of water vapour before moving on. In this way, a good deal of moisture can be evaporated: every housewife hopes for windy conditions on washing day.

Now part of the atmosphere, the water vapour is spread upwards by convection or horizontally by advection to influence weather events perhaps far removed from the initial evaporating surface. But the process of evaporation itself affects the heat exchange between earth and atmosphere. As a mechanism, evaporation requires heat: energy is needed to free the water molecules from the tighter bonds of the liquid state. We see this illustrated daily again when we boil water: its temperature rises to boiling-point (212° F.) and there remains even though heat is still being applied. This further heat is used to convert water liquid to water vapour, and physicists call it the *latent heat* of vaporisation of water (but they define it more precisely than this). In the atmosphere, the necessary latent heat is taken from the heat content of the evaporating surface. Both the surface and, secondarily, the air in immediate contact with it are thus cooled by evaporation.

The heat thus acquired remains part of the internal energy store of the water vapour and is transported with it. But when, as later happens, the water returns to the liquid state by the process of *condensation*, the latent heat is liberated and serves to warm the air. Thus the atmosphere eventually gains the heat originally lost from the earth's surface. But condensation in the atmosphere concerns us for reasons apart from its significance in the heat exchange: it is the process by which invisible water vapour is converted into water droplets which are visible as cloud.

Condensation

Condensation occurs when certain conditions are fulfilled. In the first place, the air must be *saturated*, that is, incapable of holding any more

water in the vapour state: as we noted earlier, its capacity for doing so is strictly limited. There are two ways of inducing saturation in air. Firstly, we can go on evaporating water into a sample of air until it is saturated, when some of the water vapour condenses out: a condition of balance is reached when as many water molecules leave the vapour and enter the liquid as move in the opposite direction. This method of saturating air is of only limited importance in the atmosphere. The other way rests on the fact that the capacity of air for containing water vapour depends on its temperature, and is greater the higher the temperature. Conversely, cooling a sample of air reduces this capacity and brings it eventually to saturation point. This is the most important means by which saturation and subsequent condensation are brought about in the atmosphere: we must later examine in some detail the various processes which result in the cooling of air.

Laboratory experiments have shown that it is extremely difficult to induce condensation in air which has been cleaned of all impurities: under such (highly artificial) conditions, water vapour can exist in a *super-saturated* state. The second essential requirement for the condensation process seems to be the presence in the air of minute particles or droplets—known as *condensation nuclei*—around which water droplets can form. The most likely substances to be effective as condensation nuclei are those that are *hygroscopic*, i.e. have a chemical affinity for water. It happens that the atmosphere is well supplied with nuclei, although their nature and concentration vary from place to place. Ordinary sea salt seems the most common: salt particles reach the atmosphere in sea spray, being left suspended when the water evaporates. Certain products of the combustion of fuels act as nuclei, especially sulphuric acid droplets, which form from a combination of sulphur dioxide and water: these nuclei are most concentrated over industrialised urban areas.

By the process of condensation, some of the water vapour in the atmosphere is converted into the droplets which make up clouds. These are a fascinating part of the weather we observe, and the different forms they assume often reveal the mechanism of their origins. This will claim our attention later. There is no need to regard all clouds with suspicion: some are pleasant accompaniments of fair or fine weather: but others are associated with yet another phase of the movement of water in the atmosphere—its return to the earth's surface as *precipitation*. In meteorology, this term is used to denote all water deposits derived from the atmosphere (including, for example, dew and hoarfrost). At present, however, we shall concern ourselves only with precipitation which falls from clouds as rain and snow, two forms which, as we shall see, are often related.

Precipitation

The difference between the droplets that make up clouds and the drops that fall as rain is one of size and weight only. Cloud droplets are too light to fall, and remain suspended in the air: raindrops are some hundreds of thousand times heavier. How the one grows into the other is a meteorological problem to which the answer has only recently come to be understood and then only qualitatively. It was once thought that condensation continued upon a nucleus until the drop formed was heavy enough to fall out of the cloud. But we know now that continuous condensation does not produce large drops but merely a very high concentration of small droplets which have no tendency to fall. The problem is basically that of the initial introduction into the cloud of some droplets which are larger than their fellows: once this has occurred—and no great number seems necessary—the physical meteorologists explain the subsequent stages of the precipitation process with confidence. The larger droplets, being heavier, tend to fall relative to surrounding smaller ones. In its downward path, a large droplet is bound to collide with a number of small ones and each time this occurs the small ones coalesce with the large one. By this process of capturing moisture, a falling droplet becomes progressively larger and heavier until it leaves the cloud as a raindrop. The question is what gives certain cloud droplets the advantage in size and weight which sets the process going.

One answer has been given by the Norwegian meteorologist Bergeron; according to him the raindrop begins as an ice crystal. The Bergeron theory is based on the co-existence within the same part of the cloud of both water and ice, and we must first explain how this is possible at the same temperature. Those parts of a cloud which lie above *freezing level*, *i.e.* are colder than 32° F., are not necessarily composed of ice. The formation of ice crystals, like that of water droplets, requires nuclei, and not only are these *crystallisation nuclei* apparently rare in the atmosphere but it is not at all certain what they are. Nor is it clear whether ice crystals form directly from water vapour or by the freezing of water liquid. Furthermore, crystallisation nuclei do not seem to be effective until temperatures are as low as 5° or 10° F. At temperatures higher than this but below 32°, the cloud consists of water droplets in what is known as the *supercooled* condition. A large rain-cloud can thus be divided up into several temperature zones. The lowest and warmest part, above 32° F., consists of water droplets in the normal state: higher up, with temperatures between 32° F. and 5-10° F., is the zone of supercooled droplets: above this is the all-important zone where these droplets mingle with increasing numbers of ice crystals: while only in regions colder than about -40° F. are cloud tops composed entirely of ice (Fig. 12).

Feet.

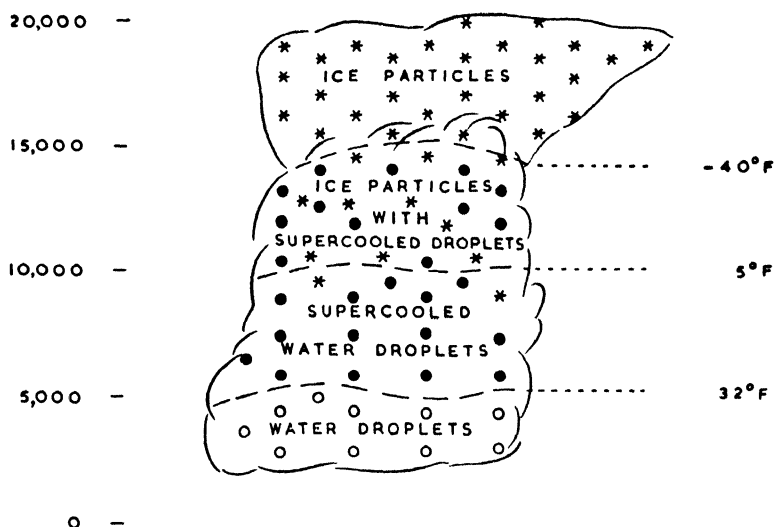


Fig. 12. The structure of a rain cloud. (The scale of feet is included only to give a rough idea of the extent of the temperature zones. This varies, of course, with the air-mass.)

The 'Bergeron mechanism' operates in the transitional part of the cloud where a few ice crystals exist together with supercooled droplets, both surrounded by air saturated with water vapour. We have explained saturation as a condition of balance in the interchange of water molecules between a water body and the adjacent vapour: a similar state of equilibrium can be achieved between an ice particle and surrounding water vapour. But ice saturation differs from water saturation. Air that is just saturated in relation to water is more than saturated in relation to ice. So that when ice and water exist together in vapour-laden air, the water molecules flock from the vapour to the ice crystals: this leaves the air unsaturated with respect to the liquid droplets which evaporate in order to retrieve the balance. The process continues with the result that the ice crystals grow rapidly at the expense of the water droplets. Eventually, the crystals are heavy enough to fall as snowflakes, but once below the freezing level, they melt to form large raindrops, which, in their descent through the normal water zone of the cloud, grow even larger by coalescence with small droplets.

There is considerable evidence that the Bergeron process is responsible for most of the rain of temperate regions at least. Observations show that rain is most likely from big clouds with considerable vertical extent, in fact those that reach levels at which ice crystals can exist. In winter, when

these levels are sooner reached, precipitation may fall from clouds of limited upward development: on the other hand, summer clouds may reach considerable heights without achieving the level necessary for rain formation. Similarly, in the cold air of Polar regions, precipitation—which remains in the snowflake form—can fall from quite thin, low clouds: while great towering clouds in warmer regions may yield not a drop of rain. And often enough we hear on the B.B.C. forecast that showery weather is expected in northern districts, but that showers will fall as snow over high ground.

Not all rain, however, can be explained by the Bergeron process. There is clear evidence from warm regions that substantial rain often falls from cloud which does not reach freezing let alone the ice crystal level. For the explanation of this 'warm rain' from 'warm clouds' is needed some alternative mechanism for introducing large droplets which can grow at the expense of their fellows. It seems—and we may not yet know the complete answer—that the large droplets result from condensation on to extra-large hygroscopic nuclei (probably of salt, reaching the air from breaking wave crests): these would grow more rapidly than those on smaller nuclei. Once the large droplets existed, they would fall and grow by capturing smaller droplets in their path. This is the *coagulation process*, also known as the 'Bowen-Ludlam process' after two meteorologists (the former Australian, the latter British) who have contributed much to our knowledge of the mechanism.

So far the evidence of this type of precipitation has come from tropical marine climates, but we cannot be sure that it does not occur in temperate latitudes. Thick clouds are necessary for coagulation, so that the falling drops can collect many droplets. Probably, in temperate regions, with the required depth of cloud, the tops are in any case in the ice-plus-supercooled-water zone, and the Bergeron mechanism is available. Even so, coagulation can operate in the later stages. While there is more to be learned, both processes are understood sufficiently to provide the theoretical basis for practical experiments in rain-making (see Chapter 11).

AIR-MASS PROPERTIES

In everyday conversation we describe the weather in such homely terms as warm, cold, bright, muggy, wet, thundery, and so on: in doing so we align ourselves with the professional meteorologists, for we are in most cases—whether we think so or not—discussing the properties of the prevailing air mass. Many of the weather elements we observe are associated with particular air masses, and a change of weather is generally the result of a change of air mass, with the added complication that the boundaries between air masses are in any case regions of disturbed weather. It is putting the matter a little crudely to state that the weather of the moment is (shall we say) warm and muggy because it comes from a warm and muggy place, for air masses may change character as they move: but this commonsense view of the weather is not too far removed from modern air-mass concepts. We are not in this chapter concerned with the origin and modification of air masses, but with the properties by which we describe and identify them.

An air mass is defined as a large portion of the atmosphere with approximately uniform properties in the horizontal direction: ‘approximately uniform’ because of local variations (which we examine in Chapter 9), and ‘in the horizontal direction’ because, as the last chapter made clear, there are changes vertically in all kinds of air, although the particular way in which the properties vary with height is again characteristic of the air mass.

In Chapter 3 we noted, for future reference, that the temperature and humidity characteristics of the atmosphere are acquired from below. Temperature and humidity are the basic or *primary properties* of air masses. It is because the processes of transferring heat and moisture between surface and air vary over the face of the earth that the atmosphere is divided up into the distinctive portions we call air masses (see Chapter 5). Meanwhile we must show that all the other characteristics of air masses—including cloudiness, visibility, likelihood of precipitation and others less readily defined—depend on the primary properties of temperature and humidity and on their vertical distribution.

The Surface Temperature

The temperature of the surface air—as measured by the dry-bulb thermometer in the Stevenson screen—is the most obvious air-mass property and the one that most readily impresses itself upon us. But there is not always a direct relationship between the true air temperature and our subjective impression of warmth or cold. We have already noted the difference between warm air and simply feeling warmed by exposure to direct sunshine, while our sensation of warmth or cold is influenced also by humidity and wind strength.

We enjoy thermal comfort when the excess heat produced in our bodies as a by-product of our internal chemical processes (*metabolism*) is successfully eliminated by body cooling. This can occur by means of conduction and convection losses from the skin to the adjacent air, by radiative cooling to surrounding colder objects, and by evaporative cooling due to perspiration. Within a range about an ideal temperature (which for most temperate land dwellers is around 65°-70° F.), body cooling is controlled automatically by the nervous system through the regulation of the flow of blood near the skin. At temperatures below the limit of operation of this natural thermostatic control, we feel cold and must check excessive heat loss by wearing suitable clothes or by heating our dwellings. At temperatures above the comfort zone we gain rather than lose heat by radiation and convection, and can lose it only by evaporation, that is by sweating.

Humidity now becomes important as little evaporation is possible in moist air (see Chapter 3) and, unable to cool sufficiently in spite of continued sweating, we feel hot, sticky and oppressed. In dry air, however, evaporative cooling is more effective and we can tolerate higher temperatures without discomfort. Cold damp air is unpleasantly raw and chilly since moisture condenses on clothing and reduces its insulating properties. Cold dry air, on the other hand, is keen and stimulating, at least in small doses. Strong winds are unpleasant with low temperatures as they cause excessive cooling by conduction and convection, but breezes are welcome in hot weather when they encourage evaporation from the perspiring body.

The diurnal variation of surface temperature has been described in Chapter 3. Temperature is an element very subject to local peculiarities and these are considered in Chapter 9. We now enquire what happens to the temperature of air when, by some means, it is lifted off the surface in a vertical current.

Temperature Changes in Vertical Air Currents

When air rises in the atmosphere it becomes cooler, and when air at some height sinks it becomes warmer. This is not in any direct way

connected with the general decrease of temperature with height in the atmosphere. It must be remembered that heat conduction in air is negligible, and we can therefore regard a vertically moving body of air as effectively insulated from the surrounding air. No actual heat exchange is involved, and the temperature changes occur solely because the air moves to a region of different pressure. We call such a process *adiabatic*. In contrast, the transfer of heat between the very lowest air layers and a warm or cold surface is a *non-adiabatic* process.

When what the meteorologists like to call a 'parcel' of air descends in the atmosphere it moves into regions of progressively higher pressure: in other words, it becomes compressed. In this process, energy is expended on the descending air and is transformed into the energy of motion of its molecules: as a result, its temperature rises. Cyclists are familiar with an everyday illustration of warming air by compression, when the lower end of the bicycle pump warms up in the process of pumping. When, on the other hand, a 'parcel' of air rises, it moves into a region of lower pressure and therefore expands: in this case, the expanding air is doing work in pushing aside the surrounding air, and the energy required for expansion can come only from its internal energy store. This manifests itself as a fall in temperature. Familiar illustrations of cooling gases by allowing them to expand are less easy to find, but the method has found an application in certain types of household refrigerator.

In the atmosphere then we associate the ascent of air with adiabatic cooling and descent with adiabatic warming. The rate at which this cooling or warming occurs depends on immutable laws governing the behaviour of all gases, and is quite constant, assuming the air remains unsaturated. We call it the *Dry Adiabatic Lapse Rate* ('dry' here meaning simply unsaturated), and it is a rate of temperature change with height equal to 5.4° F. per 1,000 feet. We shall refer to it for convenience as the D.A.L.R.

Stability and Instability in Dry Air

Whether or not air is likely to move upwards in vertical currents depends on the vertical distribution of temperature within the air mass, *i.e.* on the lapse rate. A glance at the radio-soundings plotted in Fig. 9 will remind us how far particular lapse rates may depart from the average tropospheric value of $3\frac{1}{2}^{\circ}$ F. per 1,000 feet. Whatever it may be, we term the lapse rate as derived from observations the *Environment Lapse Rate* (E.L.R.): this serves to remind us that it refers to the general distribution of temperature vertically in the air mass, and to distinguish it from the D.A.L.R. which describes solely the behaviour of unsaturated air 'parcels' moving vertically within the environment air (and also from still another lapse rate which we have yet to introduce).

The readiness or otherwise of the air to submit to vertical movement is a highly important air-mass property which we describe in the terms *stability* and *instability*. We define stability as a condition in which vertical motions are not encouraged: conversely, instability is the state in which such movements are favoured and frequently occur. The distinction can be appreciated by comparing, at given levels, the temperature of the surrounding or environment air with that of an isolated 'parcel' moving upward within it. We illustrate the point first with the aid of Table 1.

TABLE 1

Height Ft.	Column 1.	Column 2.	Column 3.
	E.L.R. 5° F./1,000 ft.	D.A.L.R. 5.4° F./1,000 ft.	E.L.R. 6° F./1,000 ft.
5,000	25	23.0	20
4,000	30	28.4	26
3,000	35	33.8	32
2,000	40	39.2	38
1,000	45	44.6	44
0	50	50	50

Column 1 represents the observed change of temperature with height in the lowest few thousand feet of an imaginary air mass in which we suppose a uniform E.L.R. of 5° F. per 1,000 feet. Let us further imagine that we can, by some means, apply an upward push to a part of the surface air in an attempt to create a rising air parcel within the air mass. Such an ascending parcel has no alternative but to cool at the D.A.L.R. and its change of temperature with height is shown in Column 2. By the time this air has risen to a height of 1,000 feet it has become 0.4° F. cooler than the still air around it. With increasing upward movement, the rising air becomes progressively cooler than the environment: at 5,000 feet the difference has become 2°. This temperature difference is significant because the *density* (the weight of a unit volume) of a gas depends on its temperature: if this is increased, the density decreases and the gas becomes lighter: conversely, if the gas is cooled, it becomes denser and heavier. Our rising air is all the time becoming cooler and therefore heavier than the surrounding air. Clearly then, it has no inherent tendency to continue rising and, if left to itself, will sink back to the surface: it will rise, unwillingly as it were, only if we persist in pushing it upward ('forced ascent' which is described later). This is, of course, what is meant by the stable state—one in which vertical movement is resisted.

In contrast, consider another air mass, this time with a uniform E.L.R. of 6° F. per 1,000 feet, as represented in Column 3. Again we apply a push to part of the surface air, again the ascending parcel cools at the D.A.L.R. (Column 2): we compare Columns 2 and 3, and find that in this case the rising air is 0.6° F. *warmer* than the surrounding air at 1,000



PLATE V. Polar continental weather: skating on Whiteknights Lake, Reading, early 1947



PLATE VI. Polar continental weather: rime accretion on the B.B.C. television mast at Holme Moss, Yorkshire, March, 1954

feet, 1.2° warmer at 2,000 feet, and 3° warmer at 5,000 feet. As it ascends, the rising air becomes progressively warmer and lighter than its environment, in other words, progressively more buoyant: there is nothing to stop it rising with increasing vigour. This clearly is an unstable condition.

It is now obvious that air is stable if its E.L.R. is less than, and unstable if its E.L.R. is greater than, the Dry Adiabatic Lapse Rate. There is, of course, a further possibility: the E.L.R. may exactly equal the D.A.L.R.: in which case vertical motion may or may not occur, and we describe the condition as one of *neutral equilibrium*.

In practice, the forecaster ascertains the stability conditions of an air mass by comparing the E.L.R. and D.A.L.R. graphically on a form of

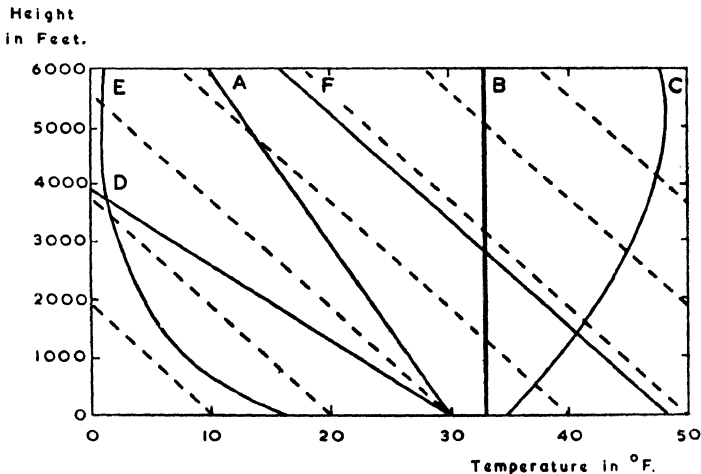


Fig. 13. Some lapse rates in the lowest layers of the atmosphere. The broken lines are dry adiabats

temperature-height diagram. Fig. 13 is a diagram of this kind: on it any straight line (though only a convenient few are inserted) of slope 5.4° F. in 1,000 feet represents a D.A.L.R.—a so-called *dry adiabat*. A number of sample E.L.R.s have been drawn in to illustrate various conditions of stability and instability. We immediately recognise a stable E.L.R. (line A) as a line of steeper slope than the dry adiabat: the more vertical the E.L.R. the more stable is the air—line B represents an isothermal layer—while, clearly, an inversion (C), in which the E.L.R. actually leans to the right, implies a condition of extreme stability. Unstable air is represented by a line of lesser slope, or one leaning more to the left, than the dry adiabat (D): line E represents air which is unstable below but stable above. Finally, the case of neutral equilibrium is depicted by a line parallel to, or, more precisely, coincident with, a dry adiabatic (F). All these con-

ditions are liable to be found at different times in the lowest layers of an air mass.

In these lowest layers there is often a diurnal variation of lapse rate—and therefore of stability conditions—which is closely associated with the diurnal variation of surface temperature referred to in Chapter 3. Assuming clear calm conditions over land, the surface temperature varies from early morning minimum to early afternoon maximum: with increasing height above the surface, however, the effect of this cooling and heating diminishes until a level is reached at which no diurnal variation is discern-

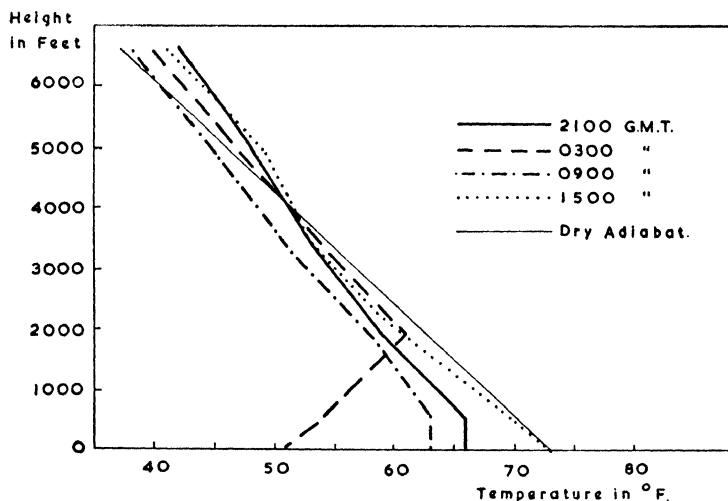


Fig. 14. Diurnal variation of lapse rate in the lowest layers (inland). Aldergrove, Northern Ireland, 5th-6th June, 1951

ible. Fig. 14 shows an actual example. The evening lapse rate is everywhere stable and the lowest layer is isothermal. During the night the ground and the adjacent air become increasingly chilled and the lapse rate even more stable so that the early morning curve shows a well marked temperature inversion. Later, as the rising sun warms the ground, the change proceeds in reverse, and the mid-morning situation resembles that of the previous evening. With increasing insolation the ground warms up strongly, especially in calm summer weather, and the mid-afternoon lapse rate is decidedly unstable in the lower layers. This diurnal change of lapse rate does not occur over the open sea.

Vertical Motions in Dry Air

In the previous section we resorted to a hypothetical 'pushing up' of bodies of air in different environments in order to illustrate the conditions

in which vertical air movement is likely or unlikely. It is now time to replace this artifice by a consideration of the actual processes which produce these up-currents in nature. In Chapter 3 we described all vertical movements in general terms as convection, but in fact there are three types which should be distinguished by separate names.

Orographic Ascent

The most obvious type—and the one that is nearest our hypothetical example—is the forcible lifting of horizontally moving air by high ground in its path: this is *forced ascent* or *orographic ascent*. Clearly enough, under these conditions, the air must ascend whether it is stable or unstable: but in the first case, the amount of uplift will be limited, while, if the air is unstable to begin with, the orographic lifting will simply set off its inherent tendency to upward movement. Orographic ascent is of great significance for the weather of hilly or mountainous terrain. We shall consider it together with other orographic effects in Chapter 9.

Turbulence

On a much smaller scale, whenever air moves horizontally over a more or less rough surface, it is apt to develop vertical motions. The lowest air layers are continually striking against irregularities and being joggled into a series of small-scale circulations or eddies which are carried along with the wind. This frictionally induced up and down movement is known as *turbulence*. Observe the behaviour of smoke pouring from a factory chimney on a windy day. With increasing distance from the chimney the plume widens and becomes less dark: in fact, the smoke is being diffused both upwards and downwards by vertical motions. It is this eddying, sometimes moving with the general flow and sometimes against it, that is mainly responsible for the gusts and lulls we experience as wind, and which are revealed so clearly by the trace of a pressure-tube anemometer (Figs. 23 and 45).

Turbulence is more pronounced the stronger the wind and the rougher the surface, being generally more effective over land than over sea. Turbulence is favoured by instability: it occurs in stable air, but tends to be damped down in the extreme stability of a marked surface inversion. Essentially a mixing process, churning the air as a paddle-wheel churns up water, turbulence redistributes heat and moisture throughout a depth of air which is usually the lowest 1,500 or 2,000 feet: clearly, it plays an important part in the formation and modification of air masses. Within the turbulent layer, because of the vertical motions, the lapse rate tends to become dry adiabatic (provided the air remains unsaturated).

Convection

The third type of vertical motion, *convection* proper, is thermally induced, the result of the heating of the surface air by contact with the warm ground: the air becomes light and buoyant, and rises as a convection current—the *thermal* of the glider pilot. Between the up-currents, air sinks slowly in compensating down-currents so that a circulation is set up rather like that produced by turbulence but generally affecting a much greater depth of air. Convection is dependant on the initial instability of the air: it occurs only when the lapse rate near the ground is *super-adiabatic*, i.e. exceeds the dry adiabatic, as shown, for example, by the mid-afternoon curve in Fig. 14. It follows that, over land areas, convectional activity undergoes a diurnal variation, obeying that of stability, reaching its maximum during the afternoon and ceasing at night when the heat supply is cut off.

A good deal is known about the conditions permitting convection, but it is not easy to picture the physical process actually happening above us in the atmosphere. Convection is most apparent to the observer when it gives rise to cloud (i.e. in saturated air), but so-called *dry convection* is rarely evident, except that it occasionally becomes visible on a small scale as the heat shimmer observable above strongly heated road surfaces or motor-car bonnets. It is simplest to regard convection currents as continuous columns of warm air rising above locally over-heated patches of ground: but the British meteorologists Ludlam and Scorer picture them as 'bubbles' of warm air rising periodically from the heat sources (rather as air bubbles rise in water) leaving behind a wake of slightly warmed air which is a favoured area for the rise of further bubbles.

Pure convection occurs alone only under conditions of calm or very light winds: with even moderate winds, convection is inevitably mixed with turbulence in the lower air layers. In calm air a thermal rises vertically, but in even a moderate breeze, its ascent may be at an angle of only 15° from the horizontal. The height to which a thermal will ascend depends on the stability of the air: the ascent will cease when the rising air has cooled to the temperature of the surrounding air, i.e. when it is no longer buoyant. With a low inversion this will happen quickly and convectional activity is thus limited to the air layer below the inversion level.

Certain types of ground surface warm up more rapidly in sunshine than others, and these are the favoured areas for the development of convection currents. More detailed consideration of these heat sources belongs properly to Chapter 9, but we may list some of them here as dark surfaces like bare rock or tarmac roads, built-up areas generally, fields of ripe cereals (but not of green vegetation), bare soil sometimes, and high land, especially sunny south-facing slopes.

Much is currently being learned about the nature of convection by careful observations of certain birds (like vultures, buzzards and swallows) and insects (locusts and dragon-flies) that utilise thermal up-currents as an aid to travel. Their ability at thermal soaring is now being emulated by glider pilots, who are themselves making valuable contributions to our knowledge of the subject. We learn from them that the speed of up-currents ranges from a few feet per second in dry thermals near the ground to well over 30 feet per second in some large convection clouds. Given suitable conditions, experienced pilots can travel several hundreds of miles using exclusively lift from thermals, while nearly 30,000 feet has been reached in up-currents in cloud. There are however types of soaring other than thermal and pilots also use lift due to high ground (Chapter 9).

The meteorological consequences of convection in unsaturated air are most apparent in desert and semi-desert climates. In the semi-arid southwest of the United States, dry convection has been observed even to a height of 10,000 feet. In deserts violent convection is readily initiated above super-heated bare rock or sand dunes, and is sometimes rendered visible by carrying up dust or sand. *Whirlwinds*, *dust-devils*, *sand-whirls* and *sand-pillars* are among the names given to these manifestations of extreme instability. For reasons which are still obscure, they often have a whirling motion (in either direction) about a vertical or slightly inclined axis. They reach heights generally of a few hundred feet but have been known exceptionally to exceed 3,000 feet: usually they last for a few minutes only, and often for less than a minute. They are occasionally observed in temperate regions.

However, the main importance of convection for our everyday weather is in the characteristic cloud and precipitation forms it produces. So far we have ignored the existence of moisture in the atmosphere in order to focus attention on temperature as an air-mass property. This separation of temperature and humidity has been convenient for the purposes of discussion, but, except in the driest regions of the earth, is a long way from the realities of weather. It is now time to put humidity, the other primary air-mass property, into its due place in the picture.

Humidity and its Expression

The word humidity signifies merely some measure of the water vapour content of the air, but is not itself a precise meteorological term. There are, however, several definite expressions, most of them concerned with the relative proportions of water vapour and dry air in a given quantity of the air as a whole. The water content is sometimes expressed in terms of weight of vapour contained in a unit volume of air (*absolute humidity*), or as the weight of vapour in a unit weight of air (*specific humidity*), or as

the weight of vapour per unit weight of dry air, that is, air minus vapour (*humidity mixing ratio*). Alternatively, that part of the total pressure of the air which is contributed by its water vapour content—the *vapour pressure*—is also a precise measure of the humidity, this time expressed in millibars.

For many practical purposes we are less interested in the actual amount of water vapour contained in the air than in the nearness of the air to saturation point, for this is a measure of the imminence of condensation and perhaps subsequent precipitation. This information is given by the *relative humidity* which is the ratio of the amount of water vapour actually present in the air to the amount that the air could hold, at the same temperature: it is expressed as a percentage, and may be visualised in the following way

$$\text{Rel. Hum.} = \left(\frac{\text{actual amount}}{\text{possible amount}} \times 100 \right) \%$$

at that temperature.

Thus if the air is saturated (actual amount equals possible amount), the relative humidity is 100 per cent: if the air contains only half the water vapour required to saturate it at that temperature, the relative humidity is 50 per cent.

The proviso 'at that temperature' is necessary because, as explained in Chapter 3, the capacity of air to hold water as vapour varies with its temperature. If the temperature is kept constant and more water is evaporated into the air, its relative humidity will increase. If the water vapour content of the air is left unaltered but the temperature is raised, the 'possible amount' is increased and the relative humidity falls: conversely, if the temperature is decreased the relative humidity rises. This means that, assuming no change in the actual water content of the air, the relative humidity undergoes a diurnal variation which reflects the temperature variation but in the opposite sense: actual moisture changes, such as may accompany a change in air mass, of course complicate the picture. Some examples of diurnal variations of relative humidity as shown by hygrograph traces—the length of the hairs reacts to the relative humidity—are reproduced in Fig. 11.

The disadvantage of the relative humidity is that the same value means different actual water contents at different temperatures. Thus the relative humidity may be 70 per cent on both a winter day and a summer day but, in the latter case, the absolute humidity may be twice that in the former. Provided, however, that we remember its meaning and limitations, the relative humidity is a useful measure of atmospheric dampness in terms of nearness to saturation.

Suppose now that we cool a sample of air without altering its water

vapour content. Its temperature falls and its relative humidity rises, ultimately to 100 per cent when the air becomes saturated. This will happen at a certain temperature, which is known as the *dew-point temperature* (or simply *dew-point*), and which may be defined as the temperature to which air must be cooled for saturation to occur. Cooling below the dew-point will result in condensation in some form or other. Used in conjunction with the air (dry-bulb) temperature, the dew-point gives an impression of the moistness of the air, for the closer are these two values the nearer is the air to saturation.

Relative humidity and dew-point may be evaluated from readings of the dry-bulb and wet-bulb thermometers (described in Chapter 1), which in combination form what is known as a *hygrometer* or *psychrometer*. The dry-bulb thermometer records the ordinary air temperature, the wet-bulb thermometer records a temperature usually lowered by cooling due to the evaporation of moisture from the wet muslin (see Chapter 3). If the air is very dry considerable evaporation can occur, resulting in considerable cooling, so that the wet-bulb thermometer shows a temperature depressed far below that of the dry-bulb. If the air is moist, there can be relatively little cooling and consequently only slight depression of the wet-bulb temperature, while if the air is saturated, clearly the two thermometers will read alike. The dry-bulb temperature and the depression of the wet-bulb together give a measure of the nearness to saturation of the air, and from Hygrometric Tables, based on the formulæ relating all these variables, the relative humidity and dew-point are easily found.

Stability and Instability in Saturated Air

At this stage we return to the consideration of a small body of unsaturated air moving upwards, whether due to convection or turbulence or orographic lift, within the larger body of the air mass. As it does so it must cool (at the D.A.L.R.) and we must now bear in mind too that its relative humidity must increase and it is in fact approaching saturation. Ultimately, with continued ascent and cooling, the air becomes saturated—at the dew-point temperature—and some of its water vapour condenses out as liquid water. But condensation is accompanied by the release of the latent heat of vaporisation (as described in Chapter 3), and this has the immediate effect of reducing the rate of cooling. If the air—now saturated—continues to ascend, its rate of cooling is cut down to about half the dry adiabatic—to about 2.7° F. per 1,000 feet. This introduces a new lapse rate—the rate of cooling of rising saturated air (or the rate of warming of descending saturated air)—which is termed the *Saturated Adiabatic Lapse Rate* (S.A.L.R.).

The S.A.L.R. differs from the D.A.L.R. in not being constant. As

saturated air ascends higher into colder levels, its water holding capacity decreases, there is less condensation and less liberation of latent heat. Consequently the S.A.L.R. increases with decreasing temperature and at very low temperatures is much the same as the D.A.L.R. This can be appreciated by reference to the temperature-height diagram shown in Fig. 27, in which the *saturated adiabats* are curved lines, steeper than the dry adiabats in the lower layers but becoming parallel to them at great heights.

We now have a new criterion for stability and instability when dealing with saturated air: the environment lapse rate in saturated air must clearly be related to the S.A.L.R. By analogy with Table 1, we can readily see that saturated air is stable if the E.L.R. is less than the S.A.L.R., and

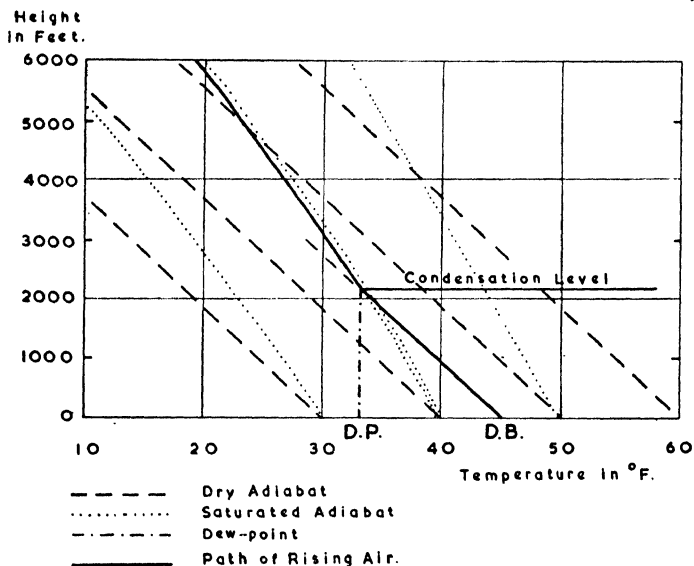


Fig. 15. Rising air depicted on a temperature-height diagram

unstable if the E.L.R. exceeds the S.A.L.R. On the temperature-height diagram the stable E.L.R. line is steeper than the S.A.L.R., while the unstable E.L.R. leans more to the left.

The change of temperature undergone by rising air may be illustrated by means of Fig. 15 which is part of a temperature-height diagram incorporating both dry and saturated adiabats. Let us depict the behaviour of a small body of air at the surface having a dry-bulb temperature of (say) 45° F. and a dew-point of 33°. If this air is lifted it cools dry-adiabatically and its change of temperature with height is represented by the dry adiabat drawn from 45° (continuous line on the diagram). Meanwhile the dew-point of this air decreases so slightly with ascent (as long as the air remains unsaturated) that we may fairly ignore this and represent the

near-constancy of the dew-point by a vertical line drawn from 33° . Eventually the rising air will cool to its dew-point—in terms of the diagram, the dry adiabat and dew-point lines meet: at this point the air becomes saturated and the height at which this occurs (about 2,200 feet) is the *saturation level*. With further ascent the now saturated air cools at the S.A.L.R. and its behaviour is depicted by a saturated adiabat drawn from the intersection. Incidentally, if this saturated adiabat is continued downwards it will intersect the surface line at a point which gives the wet-bulb temperature of the surface air (39° in this case): in other words, while the dry-bulb temperature varies dry-adiabatically with vertical motion, the wet-bulb temperature varies moist-adiabatically (*i.e.* along a saturated adiabat), and, at saturation point, all three temperatures—dry-bulb, wet-bulb and dew-point—are of course the same.

The saturation level in Fig. 15 is also the *condensation level* since above it water vapour is condensing out of the rising air to form water droplets or cloud. We may now begin to appreciate the use of the temperature-height diagram in providing information about the behaviour of air in vertical motion. The continuous line on the diagram represents the path of rising air both before and after saturation, and the condensation level is effectively the *cloud base*, under these particular conditions. But the possibility of cloud and the extent of vertical motion in a given air mass can be decided only when the stability conditions at all heights are known.

Types of Instability

We are now in a position to summarise those features of the vertical distribution of temperature and humidity in an air mass that are expressed in the terms stability and instability. We have seen that different criteria for these conditions must be applied depending on whether we consider unsaturated or saturated air. In nature the same body of air rising in a convection current or a turbulent eddy may be unsaturated near the ground and become saturated higher up. All the circumstances we have so far discussed can be fitted into one of three conditions which readily define the possibilities of vertical motion. The conditions are illustrated by three environment curves in Fig. 16: in each case the path of rising air has been drawn in (it being assumed that saturation occurs at 2,000 feet) so that the stability criteria can be applied.

Air is described as *absolutely stable* when its environment lapse rate is less than the S.A.L.R. (line A). The path for rising air is always to the left of the environment line (*i.e.* rising air is always cooler than its surroundings) and this is a stable condition whether the air is saturated or not. It represents a state in which convectional activity is not possible. An inversion is obviously an extreme case of absolute stability. This con-

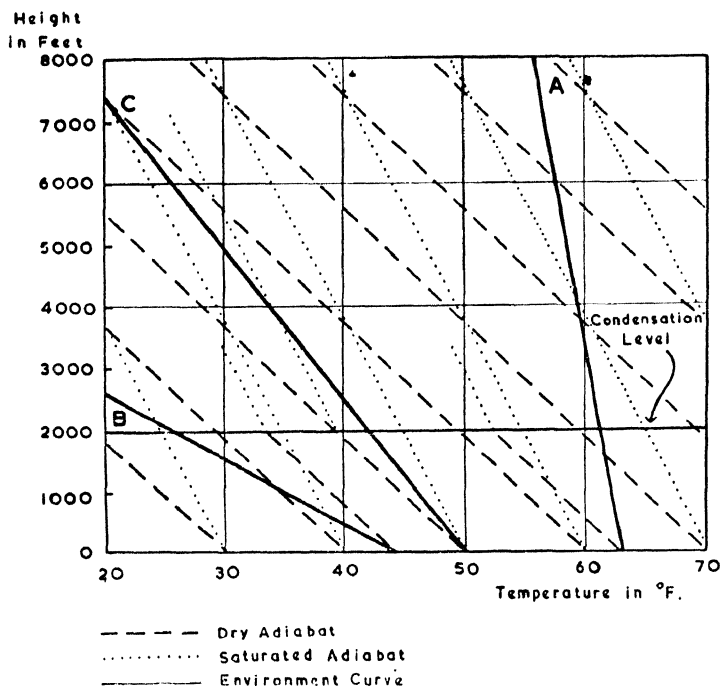


Fig. 16. Illustrating absolute stability, absolute instability and conditional instability

dition prevails in the lowest air layers due to the nocturnal cooling of the ground or when warm air moves over colder ground: it may also occur in the upper air due to subsidence (see Chapter 7).

Secondly, air is *absolutely unstable* when its E.L.R. exceeds the D.A.L.R. (line B): rising air under these conditions is always warmer than its environment. This state is practically limited to the air layer very near the ground on hot afternoons (page 68): the development of convection, which it strongly encourages, tends to reduce the lapse rate higher up to the dry adiabatic.

The third condition is so frequent and important that we shall illustrate it first by comparing the lapse rates of rising and surrounding air numerically. The criterion is a lapse rate between the D.A.L.R. and S.A.L.R., that is between 5.4° and 2.7° per 1,000 feet: in Table 2, Column 1 represents the observed decrease of temperature with height in air with an E.L.R. 4° F. per 1,000 feet, and Column 2 the temperature change in a body of air rising within this environment. It is assumed that the rising air becomes saturated at 2,000 feet. Below this level, the rising air, cooling at the D.A.L.R., becomes steadily colder than its environment: this signifies stable air. Above 2,000 feet, however, the ascending air cools less

TABLE 2

Height Ft.	Column 1. E.L.R. 4° F./1,000 ft.	Column 2. Rising Air	Lapse Rate
6,000	26	28.4	2.7° F./1,000 ft.
5,000	30	31.1	
4,000	34	33.8	
3,000	38	36.5	
	Saturation		
2,000	42	39.2	5.4° F./1,000 ft.
	Level		
1,000	46	44.6	
0	50	50	

rapidly: its temperature almost equals that of the surrounding air just above 4,000 feet, and it becomes progressively warmer with further ascent—thus satisfying the criterion for instability. In short, this air is stable while it remains unsaturated, but becomes unstable after saturation occurs.

This state is known as *conditional instability*. It is instability which lies, as it were, dormant and its realisation is conditional upon the air becoming saturated. It is illustrated diagrammatically by line C on Fig. 16. Conditional instability may be 'triggered off' in a number of ways. The requirement is sufficient lift to saturate the air: this may be provided by turbulence, by orographic forced ascent, or by surface heating and convection during the warmest part of the day. Clearly too, the moister the air (*i.e.* the sooner saturation occurs) the more likely will conditional instability become a reality.

There is a fourth type of instability that depends on the vertical distribution of moisture in the air and becomes apparent only by the large-scale lifting of the whole air mass. This type—known as *potential or convective instability*—is illustrated qualitatively in Fig. 17. The points X and Y represent the temperatures at bottom and at top respectively of a layer of air: the line XY then represents the E.L.R. (taken to be uniform) within the layer, which we recognise as absolutely stable. It is assumed that the humidity decreases with height in this layer. If the layer is lifted, all the air cools initially at the D.A.L.R., but the air at X, being moister, reaches its condensation level sooner than the air at Y. Subsequently, while the air at Y is still cooling dry-adiabatically, the air at X is cooling more slowly at the S.A.L.R. By the time the top of the layer has become saturated, the point X has moved to X' and the point Y to Y': the line X'Y' represents the lapse rate of the air layer after lifting, and, clearly, it is now much less stable.

If the humidity increases with height in the layer, the opposite occurs and the air becomes more stable on lifting. But the other vertical distribution of moisture is more common, and, in fact, most air masses are

Height
in Feet.

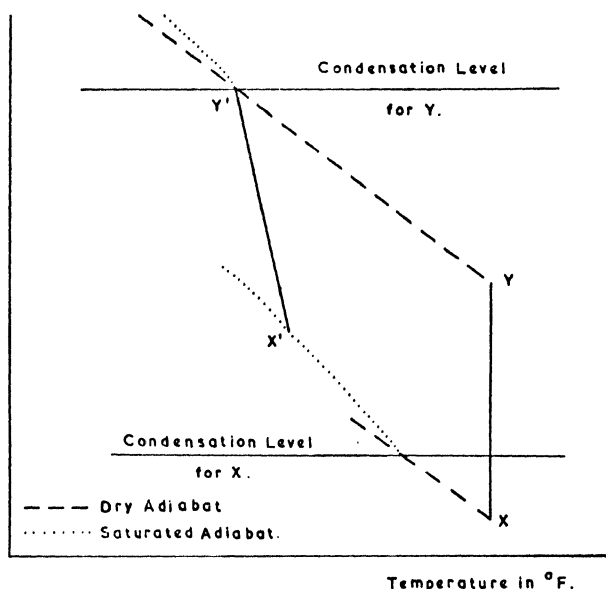


Fig. 17. Illustrating potential instability

potentially unstable if lifted sufficiently. This lift may be provided by mountain ranges or frontal surfaces.

Clouds as Characteristics of Air Masses

All clouds are the result of the cooling of air by uplift to and beyond its condensation level. This lifting may occur in various ways each of which produces typical cloud-forms. We are at present concerned only with the clouds which are characteristic of stable and unstable air masses: in the latter case the dominant cloud-forming mechanism is convection, in the former it is turbulence. The clouds associated with uplift at fronts and over high ground are dealt with in Chapters 6 and 9 respectively.

Unstable Air Masses

When convection currents rise above the condensation level each is capped by a cloud of the familiar *cumulus* variety (Plate III). These are often beautiful clouds, displaying pleasing contrasts of light and shade, hard and white when lit directly by the sun, dark and with the proverbial silver lining when the sun is behind them. To us on the ground they appear to have quite flat bases, but glider pilots report that the base is typically bowl-like, being highest at the centre (where presumably the up-

draught is strongest). The cumulus tops are cauliflower-shaped, bubbling and changing as we watch, suggestive of the upsurging currents responsible for the existence of the cloud.

The pattern of clouds in a cumulus sky renders visible the initial pattern of thermals rising from the ground. This depends on the distribution of heat sources and on the prevailing wind. Sometimes the grouping of clouds seems quite haphazard, at others, it is surprisingly regular. On occasions the cumulus are arranged neatly in rows—*cloud streets*, as they are called—and this sometimes signifies the drifting of successive thermals downwind from the same source. With the growth of large cumulus, order

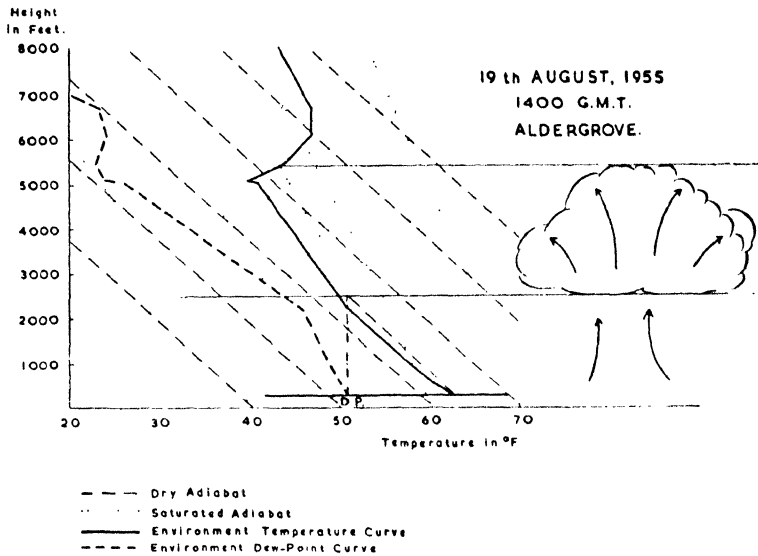


Fig. 18. Fair weather cumulus on the temperature-height diagram

in the pattern of the sky is less likely: vigorous convection tends to be self-checking, for the shadows spread over the ground by large clouds may suppress further heating and the development of fresh thermals.

In Figs. 18 and 19 the convection process is depicted on the temperature-height diagram. Both examples are of balloon ascents made at the warmest time of day when convection is most likely. In both cases, dew-points are plotted as well as temperatures: the distance apart of the two curves gives a ready indication of the moistness of the air.

The E.L.R. in Fig. 18 is absolutely unstable up to 600 feet and then conditionally unstable to about 5,000 feet, above which there is a pronounced inversion with much drier air. We now carry out the simple construction described on page 72. The dry adiabat drawn from the base

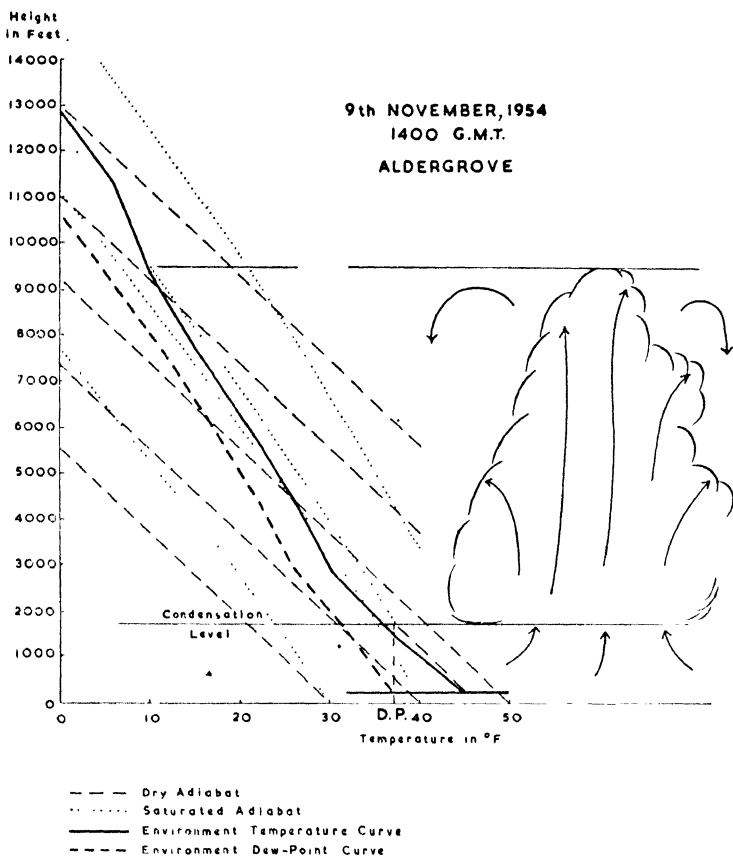


Fig. 19. Large cumulus on the temperature-height diagram

of the E.L.R. and the vertical line drawn from the surface dew-point meet to give a condensation level at about 2,500 feet. From the intersection, a saturated adiabat (drawn parallel to the nearest one on the diagram) represents the continued ascent of (now saturated) air: where this line meets the E.L.R. is the approximate cloud top, for here the rising air has cooled to the temperature of the environment and is no longer buoyant. In this example, the presence of the inversion clearly limits the cloud to shallow fair-weather cumulus, with base about 2,500 feet and tops about 5,000 (Plate VII).

In Fig. 19 there is no such inversion. Above a somewhat deeper super-adiabatic layer the E.L.R. is conditionally unstable to about 3,000 feet and only barely stable above. In addition the air is moist at all heights: clearly, the possibilities of cloud are greater in this example. Repeating the construction we find that a large cumulus, with base about 1,700—

1,800 feet and top towering to over 9,000 feet, can be produced. This cloud, unlike the fair-weather cumulus, could possibly yield a shower, as the top just reaches the ice crystal level (see Chapter 3).

Stable Air Masses

Turbulence is the main cloud-producing process in stable air. The most typical form is the *layer* or *sheet cloud* produced in the lower air layers, often beneath an inversion, by the turbulence that accompanies a moderate or stronger surface wind. Such conditions occur, for example, when warm air moves over a colder surface. The effect of turbulence is to transport heat downwards and to distribute moisture from the surface upwards. The churning motion thus modifies the initial lapse rate and tends to establish a D.A.L.R. throughout the layer. If the moisture content of the air is high, however, the colder air in the upper part of the turbulent layer may well become saturated, so that a low cloud sheet forms, its base perhaps only a few hundred feet above the surface, its top depending on the depth of turbulence.

The process is illustrated in Fig. 20 which shows successive ascents at an East Anglian station in a situation with strengthening easterly winds, in which low turbulent cloud was forecast for Eastern England. The mid-night curve, with a pronounced surface inversion and dry except near the surface, is quite transformed by noon. Turbulence has stirred the air to

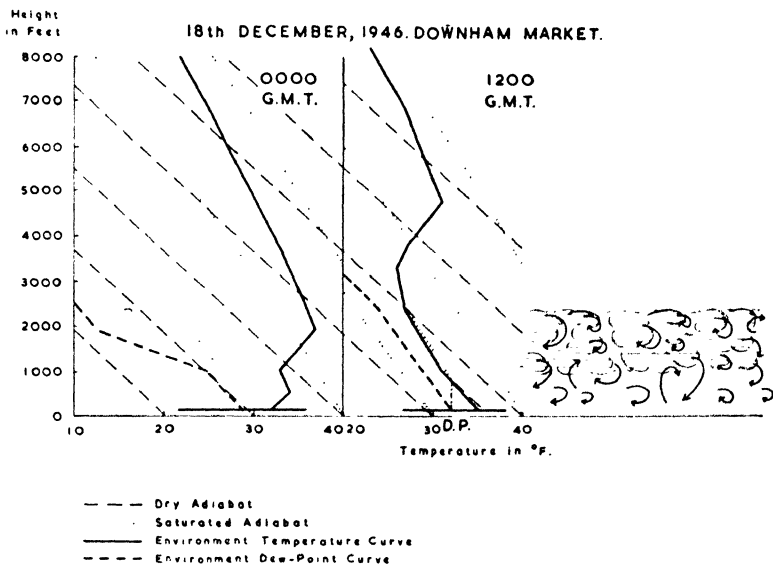


Fig. 20. Turbulent stratocumulus on the temperature-height diagram

about 2,400 feet: below this the lapse rate is practically saturated adiabatic above and dry adiabatic below, and the layer as a whole is considerably moister than at midnight. With a dew-point of 32° , a condensation level is possible at 800 feet, the base of a cloud layer some 1,600 feet thick.

The most typical sheet cloud is known as *stratus* but probably the commonest low turbulent cloud is *stratocumulus* (Plates II and VIII). This appears to us as a collection of individual cloud masses, quite large in size, sometimes rounded, sometimes elongated in bands or rolls. Often the individual masses are separated by a little blue sky, and we can appreciate that the cloud elements represent the rising currents of the turbulent motion and the intervening blue sky the down-currents of warming and drying air. With thicker stratocumulus, no blue sky may be visible but a definite pattern of light and dark suggests the same structure. Sometimes, however, the cloud sheet appears quite uniformly grey and featureless: this is *stratus* proper. It resembles fog lifted off the ground, and frequently is: at other times it represents a stage of further thickening of stratocumulus.

Because clouds of this type have only limited vertical extent they do not tend to give precipitation beyond a little drizzle or light snow (and this is often aided by orographic lift). More generally they are associated with dry but cloudy and dull weather.

Cloud Transformations

Nothing is static in meteorology: the stability conditions of an air mass or part of an air mass may change and clouds are frequently transformed from one type to another. There are several forms transitional between layer types and those with vertical development. For example, when convectional activity is limited by a strong inversion the cumulus, being unable to extend higher, spread out laterally to form a stratocumulus layer. The difference between fair weather cumulus and stratocumulus formed convectionally in this way is thus one of degree only. Similarly, at the medium cloud levels (from about 7,000 to 20,000 feet) the upper parts of large cumulus, particularly when they are dissipating, may spread out into *altocumulus* (the medium level counterpart of stratocumulus).

There is also a general tendency for a stable layer cloud to become unstable, unless it is being cooled from below, for example, by continued movement of the air mass over a cold surface. Otherwise, the base of the cloud is constantly absorbing long-wave radiation from the ground and warming, while the upper surface cools by radiation to the colder overlying atmosphere. In this way, the lapse rate within the cloud layer increases and tends towards instability. (This probably explains an otherwise

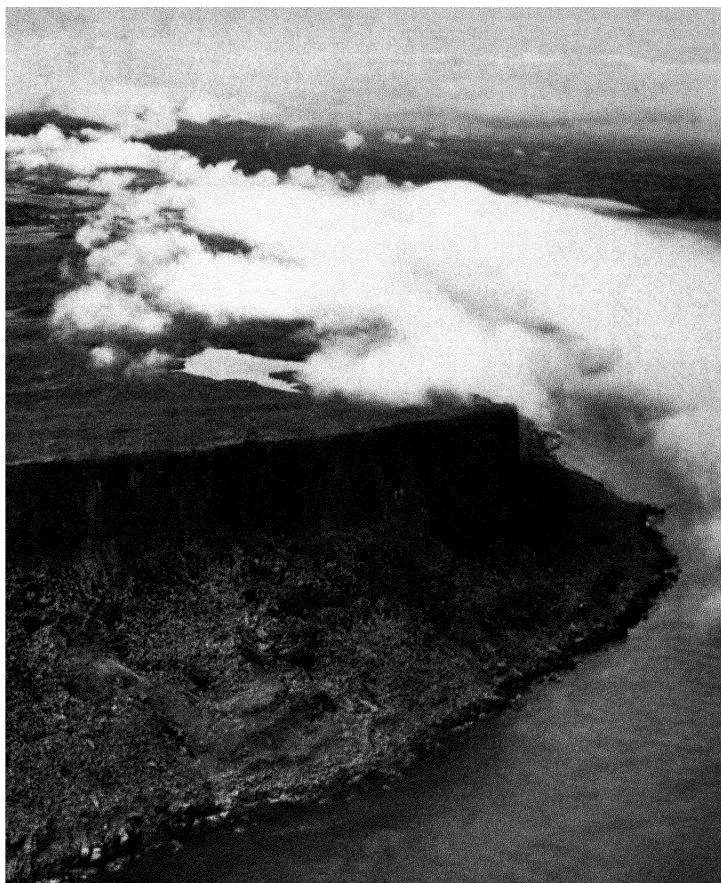


PLATE VII. Tropical maritime weather: sea fog drifting in and dispersing over the Antrim coast, Northern Ireland



PLATE VIII. Tropical maritime weather: a dull stratocumulus sky over Reading

puzzling phenomenon, that convection cloud development over the oceans reaches its maximum at night.) A stratus sheet thus breaks up into a stratocumulus layer: if the process continues the cloud eventually dissipates by warming and drying out from below. Low turbulence-formed cloud tends to persist day and night over the sea, but over the land, especially in summer, it is likely to break up during the day, due to surface warming, and to return during the night: its diurnal variation is quite the reverse of that of convection cloud.

Sheet clouds at higher levels, formed usually by frontal ascent, also tend to become unstable and be transformed into clouds with some vertical development. Thus *altostratus*, the sheet cloud of medium levels, becomes *altocumulus*, which is similar to stratocumulus except that the individual cloud elements, being higher, seem smaller: and *cirrostratus*, the sheet cloud of high levels, composed entirely of ice crystals, becomes *cirrocumulus*, a beautiful and less common type composed of tiny white masses or ripples (sometimes known as 'mackerel sky').

Extreme Stability—Fog

To the meteorologist the term *fog* signifies any atmospheric condition that obscures surface visibility to less than 1 kilometre (1,100 yards), and there are various intensities of fog also rigidly defined. With poor visibility, between 1 and 2 kilometres, *mist* is used if the condition is produced by water droplets in the air, and *haze* if it is due to dust or smoke. The man in the street, the motorist and the aeroplane pilot no doubt have their own differing ideas on what constitutes good or bad visibility. Deferring till later the effects of fog, we concentrate in this section on the condition itself. We can regard fog as essentially a cloud formed on the surface, ignoring for the moment bad visibility due to solid particles in the atmosphere (see Chapter 9). A simple classification of fogs separates them initially into those characteristic of stable air masses, which are our present concern, and those associated with fronts (Chapter 6).

Of the three types of air-mass fog, two result from the cooling of air to its dew-point by contact with a cold surface, and are distinguished by the manner in which the cooling occurs. When warm moist air moves horizontally over a cold surface, *advection fog* may form. When moist stagnant air is chilled by contact with a surface itself cooled by nocturnal radiation, the result may be *radiation fog*.

Advection Fog

Advection fog is formed by the same process as produces turbulent cloud. In the case of a stratus sheet, because of the heat transported downwards by turbulence, the very lowest air layer is too warm for

saturation and remains dry below the stratus base. But if there is less wind and consequently less turbulence, the air remains cold near the surface and the entire turbulent layer becomes saturated. The formation of this type of fog thus depends on the wind speed: too little means insufficient air movement and too much produces not fog but stratus. In a pioneer investigation of advection fog off Newfoundland, G. I. Taylor found that fogs were most frequent with a Force 3 wind (around 10 m.p.h.).

Advection fog, which is generally a few hundred feet thick, tends to be persistent over the sea, but inland, like stratus, it often disperses with day-time warming only to return at night. The clearing process may, however, go no further than the intermediate stage of 'lifted fog' or low stratus, when high ground remains cloud-capped—or in *hill fog* (depending on the view-point of the observer). An increase in wind speed will also 'lift' the fog, that is, convert it to stratus because of increased turbulence.

Radiation Fog

Advection fog may form over land or sea, but radiation fog is purely a land fog. It is associated with a surface inversion due to nocturnal radiation cooling, and it is only over land that such cooling can occur. Clear skies and relatively little air movement are required for the formation of this type of fog and we shall see that these conditions are found most often in anticyclonic situations (see, for example, Fig. 31). A typical temperature-height diagram is shown in Fig. 21; the saturated air in the lowest layers denotes fog nearly 1,000 feet deep, though many radiation fogs are shallower than this.

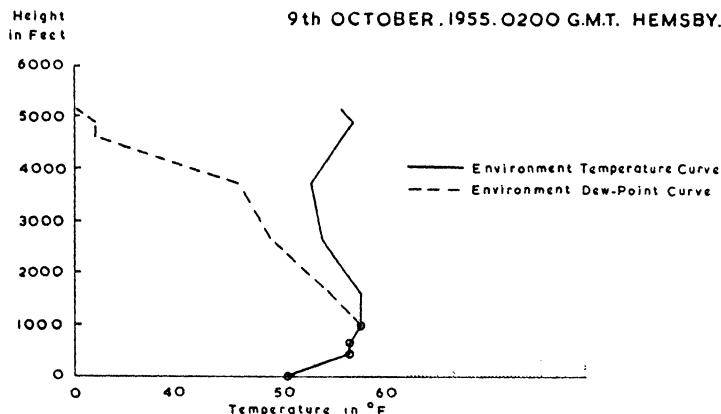


Fig. 21. Radiation fog on the temperature-height diagram

The wind speed is again the crucial factor. In the complete absence of wind and turbulence, the cooling of the air by conduction and radiation alone does no more than deposit dew on the ground and produce a layer of ground fog chest-high at most. For the development of a substantial radiation fog, some stirring is necessary to spread the cooling upwards over a sufficient depth of air. On the other hand, too much turbulence results in the formation of stratus rather than fog. Most radiation fogs form with wind speeds of 2 to 6 m.p.h. More so than the advection type, radiation fog tends towards localised occurrences, and the cleanliness of the atmosphere, the shape of the land and the nature of the surface all influence its formation (Chapter 9).

Radiation fogs occur most frequently during the long nights of autumn and winter. They occasionally persist all day and even several days on end in an unchanging situation. At other times of the year, the sun is sufficiently powerful to penetrate the fog and warm the ground, after which the fog layer is rapidly dried out from below. Summer radiation fogs are very fleeting, and are no part of the experience of those of us who rise at a normal hour. An increase in wind speed will also clear, or at least lift, a radiation fog.

Steam Fog

The third type of air-mass fog is the result of the transport of cold air over a warm water surface. This places it in the advective category. But under these conditions the saturation of the air is due not to cooling but to intense evaporation. The essential condition is a large temperature contrast between the water and the air. The cold air is quickly saturated and the resulting condensation is visible as steaming. The type is known as *steam fog* or *steaming fog* or as *Arctic Sea Smoke*, for it is best seen over relatively warm Arctic seas when very cold air blows off adjacent ice- and snow-covered land. It is a fog of no great depth, sometimes 50 feet, occasionally 100. On a smaller scale, steaming occurs in cold dry air above lakes and rivers, and from any wet surface warmed by the sun (for example, town streets after an afternoon shower in summer), not to mention the bathroom on a winter bath-night.

Extreme Instability—Showers and Thunderstorms

When air is unstable to a considerable height large towering convection clouds may form. If the cumulus tops enter into regions cold enough for ice crystals to appear among the supercooled water droplets the Bergeron mechanism operates, as described in Chapter 3, and precipitation occurs: this is of the *shower type*, characterised by its relatively short duration

and sometimes formidable intensity. With an even greater depth of instability, and always provided there is a sufficient supply of moisture, convection clouds ascend still higher and changes occur that are qualitative rather than a mere matter of degree. The cumulus develops into a *cumulonimbus*, a dark menacing thundercloud, and *thunderstorms* may occur, in which particularly heavy showery precipitation is associated with a characteristic behaviour of surface wind, temperature and humidity, and, above all, with more or less spectacular manifestations of atmospheric electricity.

The transition between cumulus and cumulonimbus occurs when the cloud pushes up into the coldest regions of the troposphere where water can exist only as ice: ice formation (*glaciation*) therefore occurs in the cloud top, which loses its sharp rounded outlines and becomes fibrous and diffuse. Just as condensation liberates the latent heat of vaporisation at lower levels, so glaciation here releases more latent heat (that of fusion of ice): this supplies energy for renewed ascent of air. With further development, the top spreads upwards and outwards into a dense mass of *cirrus* (ice cloud) which often assumes a characteristic anvil shape: the top is flat because here at last further ascent is prevented by the presence of a stable environment (which may well be the tropopause), and the cloud spreads laterally with the prevailing winds at that level (Plate XI). The base of a cumulonimbus may be as low as 1,000 feet, the tops may reach 30,000 feet in this country and 50,000 in the tropics. Sometimes the observer may see little wisps hanging from the base of a distant cumulonimbus: these are rain trails or *virga*, falling drops that evaporate before reaching the ground. At a later stage, the shower is visible as a dark curtain between cloud and ground. Sometimes, too, before the onset of precipitation, the black threatening underside of the cloud is bulbous, hanging in unmistakable pouches (*mammatus structure*), suggestive of turbulent down-currents in the cloud (Plate XII).

A detailed investigation of thunderstorms—in which surface stations, upper air equipment, aircraft flights and long-range radar were all enlisted—was undertaken soon after the war in the United States. The Director of the Thunderstorm Project, Dr. H. R. Byers, visualises the development of a thunderstorm in terms of a 'life-cycle' having three distinct stages: these are illustrated in Fig. 22.

In the first or *cumulus stage*, that of an ordinary cumulus cloud, the main feature is that updraughts prevail throughout the cloud, sometimes with speeds exceeding 50 feet per second. In the surrounding dry air slow compensating downflow completes the circulation or cell. Temperatures are higher within the cloud than at corresponding levels outside. Although the rain-forming process may be operating the ice crystals are not yet heavy enough to prevail against the updraughts.

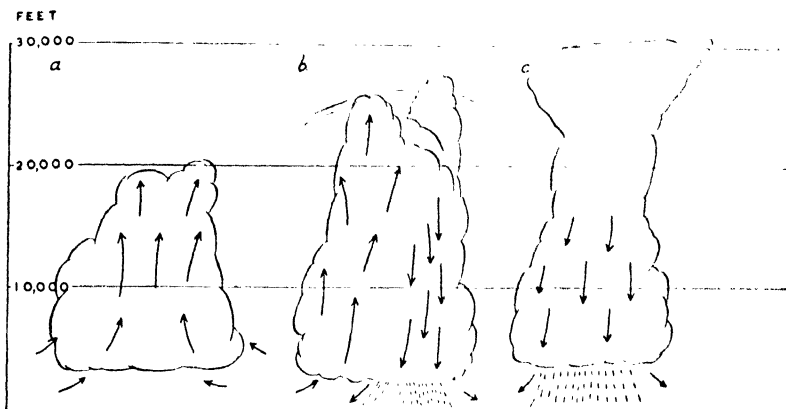


Fig. 22. Life-cycle of a thunder cloud according to Byers.
(a) Cumulus stage. (b) Mature stage. (c) Dissipating stage.

Eventually these can no longer support the increasing weight of ice crystals and water drops, and these suddenly fall as precipitation. The great weight of falling water drags down by friction cold air from high in the cloud: this is the *downdraught* which initiates the second or *mature stage*. Precipitation and downdraught reach the ground together, and the cold air spreads out as a chilly gusty wind. At this stage, there are both updraughts and downdraughts within the cloud.

The area of downdraught gradually extends throughout the cloud, and when downward movement dominates, the supply of energy is cut off and the cell has reached the final or *dissipating stage*. The precipitation dies away and the cloud dissipates from below, often leaving residual patches of altostratus (which may give light rain for a while) and cirrostratus (the remnant of the original anvil). The whole life-cycle may take an hour, but very often large thunderstorms consist of several cells combined, often at different stages, which may give a much longer duration of rainfall.

The sequence of surface weather beneath a moving thunderstorm is illustrated by continuous traces in Fig. 23. As the storm approaches a gentle wind may be felt blowing towards the cloud: this is part of a general inflow of surface air which is drawn into the updraught. Sometimes ahead of the storm this inblowing wind is effectively countered by the prevailing surface wind: the result is a temporary calm—the 'calm before the storm.'

Then with the arrival of the downdraught there occur simultaneously the onset of precipitation, a rise in pressure (due to the sudden advent of cold heavy air), an abrupt drop in temperature and equally sharp rise in relative humidity, gusts of wind and a violent change of direction as the downdraught air blows with the storm. The heaviest precipitation comes

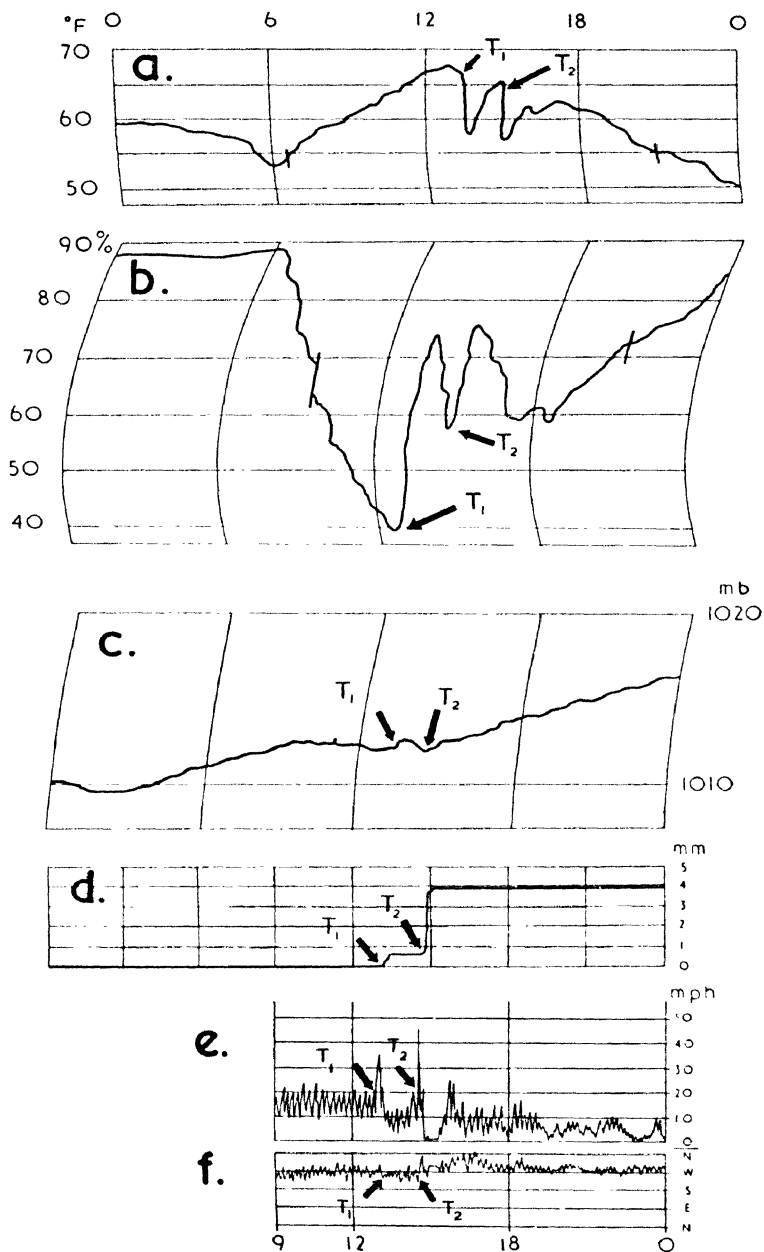


Fig. 23. Thunderstorm weather, illustrated by continuous traces of (a) temperature, (b) relative humidity, (c) pressure, (d) rainfall, (e) wind force and (f) direction. Abingdon, Berks, 9th September, 1955. The arrows marked T_1 and T_2 indicate the onset of two thunderstorms

first, with large drops and an intense rate of fall, and gradually slackens to light rain. Even some distance from the storm, beyond the edge of the rain area, the gusty wind, pressure rise, temperature drop and humidity increase may all be experienced (Fig. 24).

Sometimes the precipitation includes *hail*, the result of continued accretion of water around an ice crystal falling through the cloud. Where the number of water droplets is small, usually in the colder parts of the cloud, freezing is rapid as the latent heat liberated by the process is soon conducted away: the resulting ice layer is white and opaque due to the inclusion of air. With a higher liquid water content, usually at higher temperatures, the latent heat is less quickly lost and freezing is slow, taking the form of clear ice. Thus the onion-like structure of successive layers of opaque and clear ice that we see when we cut a large hailstone in half is explained by the passage of the stone through cloud of alternately high and low liquid content.

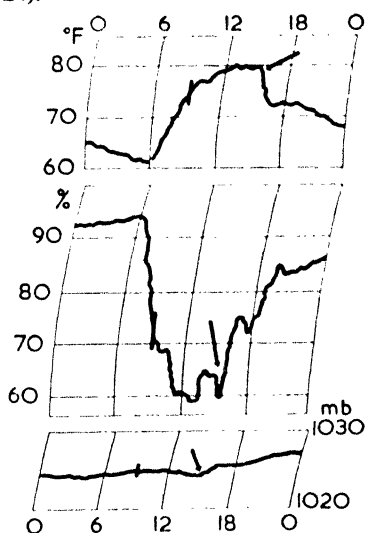


Fig. 24. Thunderstorm weather. The effect of a storm at Tilehurst, Reading, on the recording instruments at Reading University (3 miles away) where, however, no rain fell. (14th July, 1955.) The arrows indicate the arrival of cold downdraught air

Thunderstorm Electricity

But the most characteristic feature of the thunderstorm is the electrical activity leading to thunder and lightning. The mechanism responsible for the building up of electrical tensions is still not satisfactorily explained. But largely through the work of the British meteorologist Sir George Simpson, we know something of the distribution of charges in a typical thundercloud (Fig. 25): the cloud top is positively charged, and everywhere else is negative except for one or more positive patches near the base of the cloud. These often coincide with the downdraught areas, and the heaviest rain is generally positively charged.

As Simpson showed experimentally, when water drops are broken up by strong upward currents, as might occur in a thundercloud, the residual droplets acquire a positive, and the surrounding air a negative charge. Since the droplets descend, or at least remain suspended, while the air is

carried upwards, this may explain the positive areas near the cloud base and the negative area above. It is also suggested that when ice crystals rub together they become negatively charged by friction and the air becomes positive, and that this occurs in the upper regions of the cloud: the heavier crystals would sink, leaving a net positive charge at the top. There is no direct evidence supporting this theory, and more investigation remains to be done.

The building up of electrical tension in some such way is eventually relieved by discharges, giant sparks which leap from cloud to cloud, or cloud to earth, or between different parts of the same cloud. This is what we observe as *lightning*. The popular distinction sometimes drawn

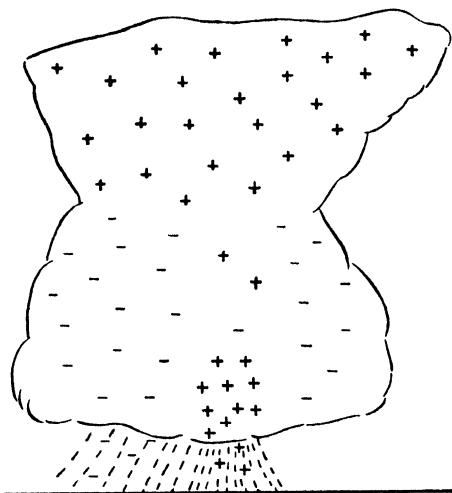


Fig. 25. The distribution of electric charges in a thundercloud

between 'forked lightning' and 'sheet lightning' is quite without basis: the latter is simply the momentary illumination of distant clouds by discharges that are themselves invisible because other clouds are in the way. A kind of discharge around a central nucleus is occasionally observed and given the name *ball lightning* but very little is known about this phenomenon. The heating of the air in the path of the lightning discharge results in a sudden expansion followed by contraction which sets up the sound waves, augmented by echoing effects, that assail our ears as *thunder*.

Not all thunderstorms develop within the air mass: some occur in association with fronts (Chapter 6). The air-mass thunderstorms may result from local over-heating or may be 'triggered off' in various ways which involve some consideration of local factors (Chapter 9). The effects

of thunderstorms and the problem of forecasting them are dealt with in Chapters 10 and 11.

Conservative and Non-Conservative Properties

Some air-mass properties remain fairly constant with the passage of time, others are very subject to change. Clearly, an absolutely constant property would be a boon to the forecaster who could use it as a kind of identification label enabling him to trace the air mass from chart to chart. In practice no property quite fulfils this requirement, but some are more *conservative* than others.

The dry-bulb temperature is a good example of a *non-conservative* property: it undergoes a diurnal variation, it changes adiabatically when lifted and it is much affected by local factors. Although the dry-bulb temperature is plotted on synoptic charts and provides valuable information in other ways, it is often too fickle as a means of air mass identification. Among the humidity elements, relative humidity has the same defects.

The most conservative of the properties we have considered is the dew-point, which hardly varies except when the actual moisture content of the air is altered. There is a diurnal variation—the dew-point is lower at night after the deposition of dew than during the day when this moisture is re-evaporated—but the range is slight, a few degrees at most. It is subject to small local variations: it will increase after a shower (again due to evaporation), and is a little higher over coastal areas than inland. But its change with height in ascending air is negligible, as we have seen. In general the dew-point is the most valuable air-mass label, and is easily determined and plotted on synoptic charts or upper air diagrams.

The secondary characteristics, like cloud type or visibility, are liable to variation but often give useful supplementary clues to the identification of air masses.

AIR-MASS TYPES

A knowledge of the characteristic properties of air masses enables us to condense a reasonably full description of the weather they bring into a few pertinent words. If, for example, we speak of 'cool moist unstable air,' this conjures up a picture of convectional activity, with towering cumulus clouds and showers separated by brighter intervals, or perhaps even thunderstorms: while a 'warm stable air mass' implies warm, probably muggy, weather, with more or less cloud of the stratus or stratocumulus type. In the present chapter we consider the origin of air masses, how they acquire the properties by which we distinguish them, the circumstances in which these may be modified, and we suggest a convenient classification of air masses that will help us to understand and to describe regional weather, with particular emphasis on the British Isles.

Source Regions

All air-mass characteristics depend on the primary properties of temperature and humidity which are acquired from underlying surfaces by various processes of transference—radiation, conduction, turbulence, evaporation—already considered. An air mass being defined as a large and fairly homogeneous body of air, it follows that its uniform properties must be derived from a large and fairly homogeneous area of the earth's surface. Such an area is called a *source region*.

Physical uniformity is clearly the first essential feature of source regions. They are entirely sea, or entirely land regions of uniform surface, for example, desert or snow-covered lowland. The second requirement is that the air should remain in residence over the source region long enough to allow the processes of transference of heat and moisture to operate. From the meteorological viewpoint, source regions can be effective only in the relatively stagnant parts of the general circulation: these, we shall see, are the great permanent and semi-permanent high pressure centres of the earth, areas of settled weather and stable conditions where air descends slowly and spreads outwards, so that influences from outside the source regions cannot penetrate. The air mass eventually moves away from the source region, according to the immediate dictates of the pressure pattern, and ultimately towards adjacent low pressure areas.

The major source regions of the world, fulfilling both geographical and meteorological requirements, are the Polar regions of ice- and snow-covered land or water, the sub-tropical deserts, the sea areas under the sub-tropical high pressure systems, and the great Continental areas when they are in the grip of winter snow. Thus air masses are initially warm or cold, wet or dry, according to the nature of the source region.

The Modification of Air Masses

Away from the source regions that gave them birth, air masses move often over surfaces of different character and undergo significant *modification* of their initially acquired properties. Consider for example an air mass formed over the North Atlantic in the so-called 'Azores High' (one of the sub-tropical high pressure centres): this is a warm air mass originating over water of surface temperature 60° – 70° F. Some of this air spreads northwards, thus passing over progressively colder seas: the air mass is then cooled from below and its initial stability strengthened, with all this implies for its associated weather. On the other hand, air of the same kind also moves towards the equator, eventually to regions where the sea temperature is over 80° F.: this air is warmed from below, its initial stability is destroyed and it becomes an unstable airstream associated with a very different type of weather. Thus the loss or gain of heat from underlying surfaces is a most important air-mass modification.

Equally significant is the loss or gain of moisture. An air mass originating over land is dry but once it moves over water it greedily takes up moisture, its very dryness encouraging evaporation at a rapid rate (especially if it is also warm). On the other hand, an air mass born over the sea is initially moist: if it moves over land it tends to lose moisture by precipitation but this process may be slow—especially if it is little aided by orographic uplift—and the air may remain quite humid even after deep penetration into the heart of a continent.

The depth to which air-mass properties may be modified varies according to circumstances. In the case of an air mass becoming more stable on account of its movement, both the surface cooling and any further acquisition of moisture are restricted to the lowest (turbulent) layer, usually under a marked temperature inversion. In the case of an air mass becoming more unstable by virtue of its movement, both the heat and moisture gained are distributed upwards to greater heights by the convection that is inseparable from instability.

Another order of air-mass modification is related not to surface influences but to large scale vertical movement within the air mass as a whole which are imposed by changes in the associated pressure pattern. The modifications take the form of general uplift, making for a lessening of

stability, or general sinking which tends to stabilise the air. These dynamic effects are related to the depressions and anticyclones with which air masses often become involved and are best understood in connection with these pressure systems (see Chapter 7).

By one or other of these means, 'old' air masses that have wandered far from their source regions may eventually become modified out of all recognition. The immediate path of the air is then much more significant than the original source. Again, if an air mass moves over a region of diverse surface properties (for example, part land, part sea), different portions become modified in different ways, especially as regards the stability and moisture content of the lower layers. In this way variations may appear within a single air mass.

The Classification of Air Masses

The most generally accepted air-mass classification is based on the two considerations outlined above—the source region and the subsequent modification during movement elsewhere. Most source regions are located either in high latitudes or in sub-tropical regions: this gives an immediate division into *Polar* (P) and *Tropical* (T) air masses, from which we know the temperature characteristics of the air at least at its source. Cutting across the division based on latitude is a further distinction dependent on whether the source region is land or sea: thus air may be *Continental* (c) or *Maritime* (m), which tells us something about its initial endowment with moisture. There are then four combinations based on the source of the air—Polar Maritime (Pm), Polar Continental (Pc), Tropical Maritime (Tm), and Tropical Continental (Tc).

A further designation is based on the modification of initial properties by surface heating or cooling: an air mass is *warm* (W) if its lower layers are warmer than the underlying surface, *i.e.* a stable air mass. A *cold* air mass (K for *kalt*, German for cold) is colder than the underlying surface, *i.e.* it has a tendency to instability. There are, of course, occasions when air is moving along rather than across the isotherms representing the current distribution of temperature, and neither W nor K can then apply.

The basic classification is as follows

				<i>Polar</i>	<i>Tropical</i>
<i>Maritime</i>	..	..	..	Pm W	Tm W
				Pm K	Tm K
<i>Continental</i>	..	..	..	Pc	Tc W
					Tc K

It must be emphasised that the suffixes W and K refer purely to the temperature contrast between air and underlying surface. From the

'Azores High,' Tm air spreads both polewards, as TmW, and equatorwards as TmK: but TmK is eventually warmer than TmW. There can be no distinction of this kind in Pc air, for the snowbound continents where this air originates in winter are the coldest places on earth, and thus, in whatever direction it moves, Pc air is always PcK. Over land areas there is often a diurnal change in the degree of modification of the lower air layers: thus, due to the diurnal temperature variation, Pm air is PmW at night and PmK during the day, but it is not possible to allow for this kind of change in any convenient classification.

Sometimes the category *Equatorial Air* is added to the basic classification. It refers to Tropical air that has become relatively stagnant in the equatorial zone: it is not therefore a true air mass but the result of much modification under calm conditions. Also sometimes used is the class *Superior Air* describing high level air from the upper troposphere which may be brought down near, or occasionally to, the surface as a result of general subsiding movement. Initially dry, it is warmed adiabatically and further dried through its descent: at the surface it is similar in properties to Tc air. Superior Air is not an air mass in the strict sense. The term is used mainly in American practice.

Air Masses of the British Isles and their Characteristic Weather

While the classification of air masses given above is of universal application, in some parts of the world regional peculiarities may introduce modifications into the general scheme. The air masses that directly affect the British Isles are conveniently classified as in Table 3, which also summarises very briefly some of the details which now follow. The more usual routes by which these air masses reach the British Isles are shown in Fig. 26.

Each of the major types is illustrated by an actual synoptic chart (Figs. 28–33), chosen to show the characteristic distribution of weather in the air mass over the British Isles. In these charts the station data have been plotted as described in Chapter 2: readers will soon become familiar with the meanings of code figures and symbols by reference to the Appendix. On each chart the isobar pattern has been drawn in: for the moment we can accept isobars (see Chapter 7) as broadly depicting the movement of the lower air layers undisturbed by frictional effects at the surface. The charts are all for 0600 G.M.T.: that is, they illustrate conditions before day-time heating becomes effective.

Typical (but not necessarily average) upper air conditions in each air mass are shown by selected temperature-height diagrams (Fig. 27) of summer and winter examples. The dew-point curves are also plotted. These are early morning ascents illustrating nocturnal conditions: the

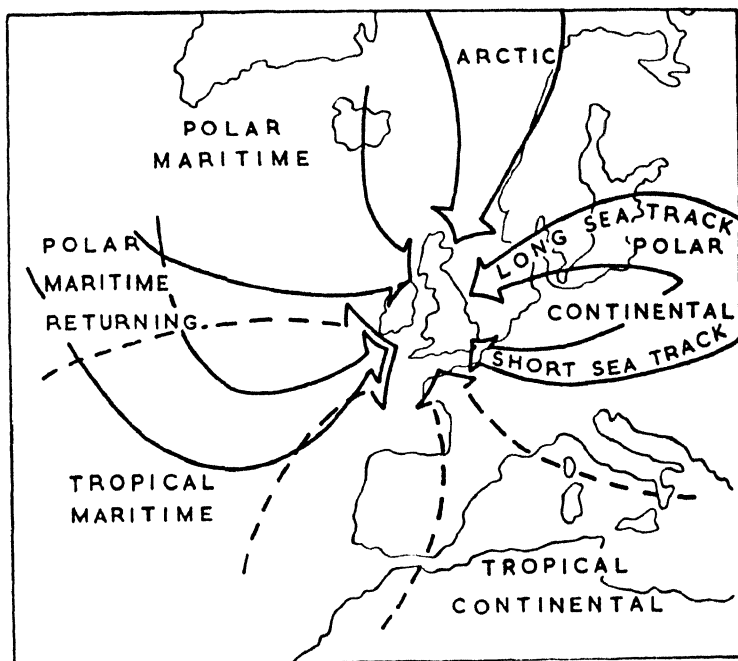


Fig. 26. Air masses affecting the British Isles

subsequent day-time modification of lapse rate near the ground is suggested by the dotted curves.

A detailed study of the characteristics and occurrence of air masses over the British Isles was made by the late Dr. J. E. Belasco, of the Meteorological Office, and we have generalised his data in Table 4, which gives some idea of the surface temperatures to be expected in each type in south-east England. These are average figures about which a variation of a few degrees either way is likely, and they leave out of account the temperature extremes that often occur under peculiar local conditions in quiet, anticyclonic weather (see Chapter 9).

Arctic Air (Direct Polar Maritime)

As its name suggests this air mass originates in the more or less permanent high-pressure area of the Arctic Ocean, together with Greenland and the northern fringe of Canada: but in fact this is also the source region of most of the Polar Maritime types we experience. The distinctive quality of Arctic air lies in the directness of its track from source to British Isles and the rapidity of its modification. Developed over surfaces of ice and snow, it is an extremely cold air mass, of low moisture content

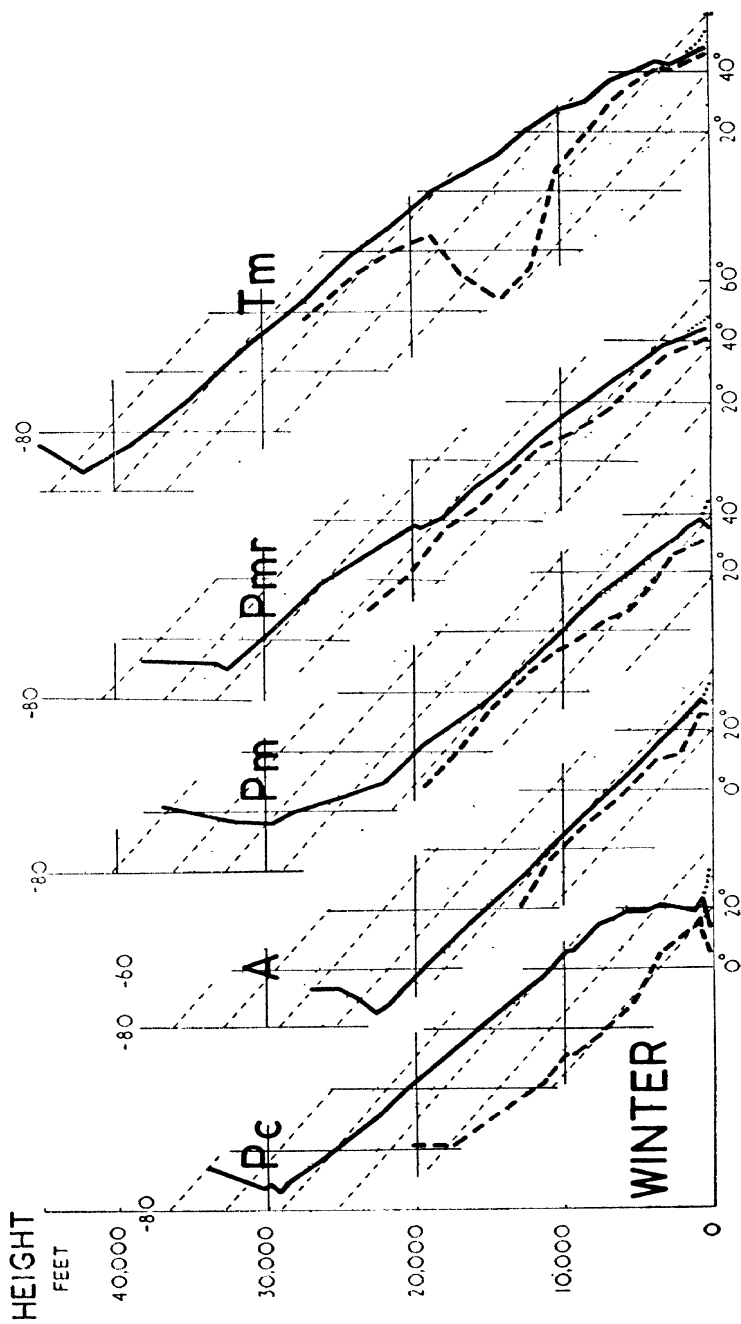


Fig. 27 A. Typical upper-air soundings in different air masses, winter

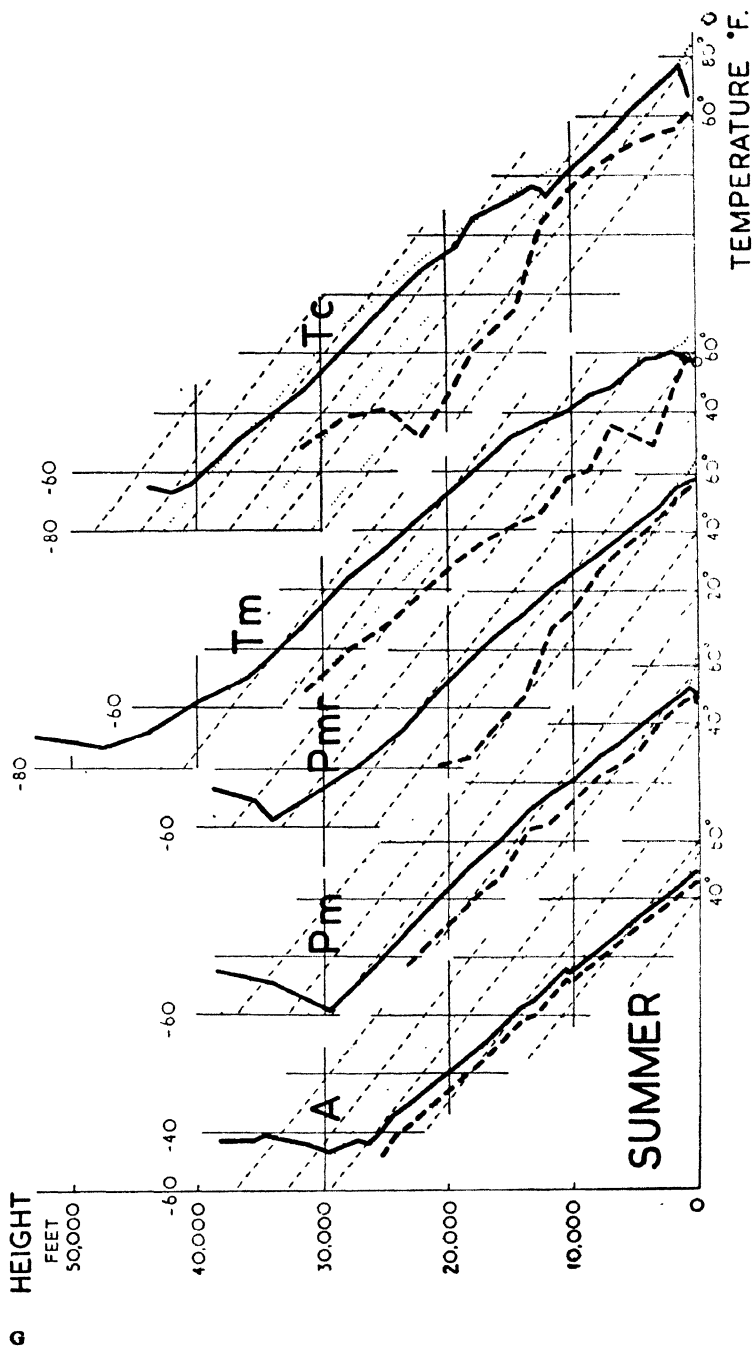


Fig. 27. B. Typical upper-air soundings in different air masses, summer

TABLE 3

Air Masses of the British Isles

<i>Air Mass</i>	<i>Source Region</i>	<i>Path Surface</i>	<i>Direction of Approach</i>	<i>Occurrence</i>	<i>Typical Cloud</i>	<i>Typical Weather</i>
ARCTIC (Direct Polar A maritime) (PmK)	Arctic Ocean	Ocean Short Track	N., N.E.	All seasons not common	Large Cu, Cb	Cold, showery, sometimes thundery
POLAR MARITIME (PmK)	Arctic Ocean, or modified Canadian Pc	Ocean Long Track	N.W., W.	All seasons common	Cu, Cb	Cool or cold. Showers and bright intervals Radiation Fog
POLAR MARITIME RETURNING Pm(τ) (PmW)	As for Pm	Ocean	S., S.W.	All seasons	St, Sc	As for Tm but cooler
POLAR CONTINENTAL Pc	Eurasian Continental Interior	(1) Mainly Land (2) Land and North Sea	E., S.E. E., N.E.	Winter not common	Little cloud Small Cu Often St. or as for Pm	Very cold and dry Cold and cloudy. Sometimes like Pm
TROPICAL MARITIME (TmW)	North Atlantic Azores region	Ocean	S.W., S.	All seasons common	St, Sc	Warm, humid, cloudy. Advection Fog
TROPICAL CONTINENTAL Tc (TcW)	N. Africa, Spain	Land, Med. Sea, Channel	S., S.E.	Summer Rare	Perhaps Cu, Cb	Hot, dry, dusty. Sometimes thundery

TABLE 4

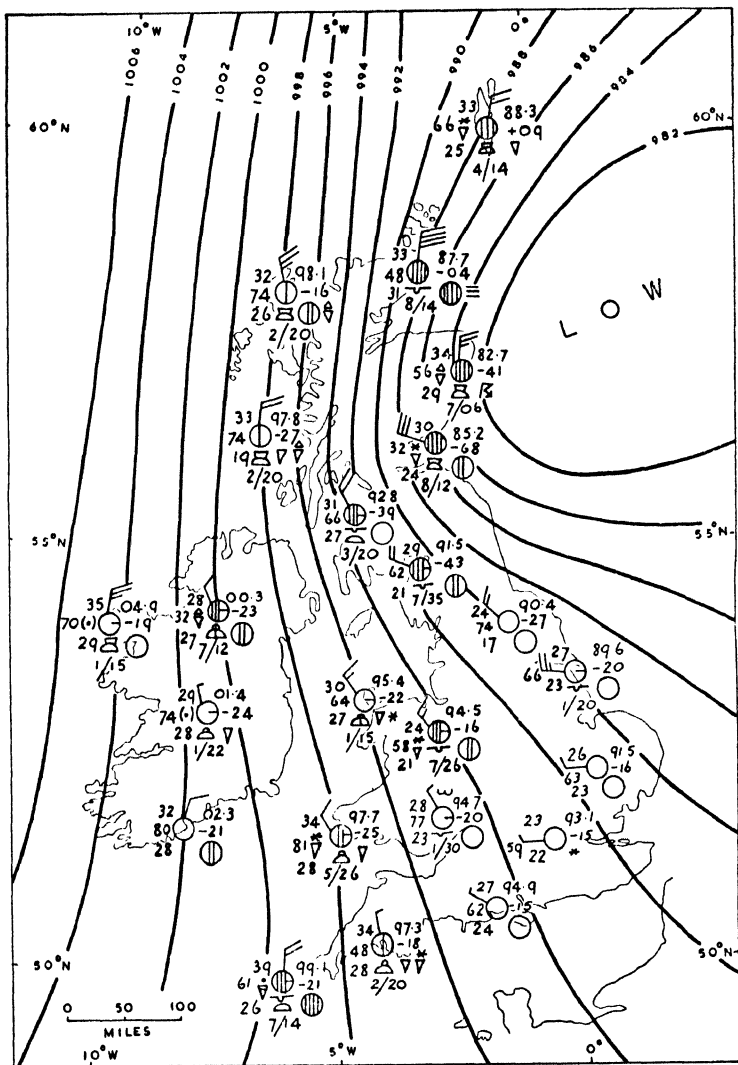
Average Maximum and Minimum Temperatures at Kew in Different Air Masses (°F.).

	J.	F.	M.	A.	M.	J.	J.	A.	S.	O.	N.	D.
A	35 28	39 29	43 32	49 37	53 39	60 47	65 51	63 50	58 45	47 35	43 32	38 29
Pm	44 35	46 36	50 39	56 43	61 47	66 52	69 55	70 54	64 51	55 44	50 40	46 37
Pmr	50 44	51 43	54 44	59 47	65 51	69 55	72 59	70 58	69 56	59 49	54 46	52 45
Pc	31 25	34 27	39 30									35 28
Tm	52 46	53 47	57 48	59 50	67 53	70 58	72 60	72 60	70 58	62 52	57 49	54 46
Tc	49 40	50 40	59 44	68 49	76 55	78 59	82 61	82 61	74 58	64 51	57 48	52 45
All Air Masses	44 35	45 35	49 36	55 40	62 45	68 51	71 55	70 54	65 50	56 43	49 39	45 36

because of its low temperature, and quite stable: an upper-air sounding usually shows a pronounced temperature inversion. On occasion this air streams southward towards our islands, passing over progressively warmer seas whose temperature is unduly high for the latitude (because of the North Atlantic Drift of warm water), exceeding 40° F. in winter and 50° F. in summer in the sea area north of Scotland.

The Arctic airstream is thus subjected to rapid warming from below: it is PmK *par excellence*. The lower layers are soon transformed from a stable to an extremely unstable state. As it gathers moisture, conditions become ripe for intense convectional activity, the building up of large cumulus and cumulonimbus cloud, towering often to 25,000 or 30,000 feet, and the release of stormy showers of rain or hail.

Fig. 27 shows typical upper-air conditions over a northern (Shetlands) station where Arctic air is experienced, so to speak, near its best. Apart from surface effects no trace remains of the original stability for by now the air has been thoroughly stirred by convection: in fact, conditional instability prevails to considerable heights. The air is very moist throughout, as shown by the closeness of dew-point and temperature curves, due to the upward transport of water vapour by convection currents plus evaporation from falling precipitation. The summer ascent is exactly similar, except that the air is warmer at all heights. In both cases the surface air is stable (the summer curve only just so), but relatively little



heating is required before convection cloud can again develop. Often enough the air is as unstable during the night as during the day.

Fig. 28 shows what a northerly airstream looks like on a synoptic chart. Sometimes the direction of approach is more north-easterly. In any case, Scotland and north-eastern England receive this air in its most characteristic form. Notwithstanding the time of the chart, cumulonimbus clouds and showers are already much in evidence in northern coastal districts, for

here, as in the neighbouring sea areas, nocturnal cooling and stabilisation are negligible and convectional activity persists day and night. Further south, especially where the air has had a long land track, the stabilising process is more effective and skies are mainly cloudless: but even here the occasional snow or shower symbol in the past weather shows that scattered showers may still happen, even at night. With a more north-easterly track, the coastal areas of East Anglia and the north of Kent may suffer frequent showers from air well supplied with energy and moisture from its passage over the North Sea. During the day showers are everywhere prevalent.

Arctic weather is far from pleasant in winter. Frost is inevitable at night everywhere except in some coastal areas and temperatures do not much exceed freezing-point, if at all, during the day. High relative humidity combines with cold to give the air a raw chilliness and, frequently, strong gusty winds make outdoor conditions bitter and harassing. Not the coldest air we experience (Polar Continental has that distinction), Arctic is yet probably the most conducive to long attendance by the fireside for those who have the chance. It provides bad weather for outdoor workers, for men of the sea working in the teeth of squalls and showers, and for aircraft pilots contending with bumpiness due to strong up-and-down currents and the danger of icing (see Chapter 11) in big clouds.

The unwelcome, wintry showers are frequent and few areas escape them. Precipitation may be of water or ice, according to the surface temperature: if this exceeds 37° F. the most probable form is rain, if it is below 32° F. snow showers are likely, and with intermediate temperatures, either, or both combined as *sleet*, may occur. In Fig. 28 most showers, present or past, are of snow, except in the warmer Scilly Isles and south-west Ireland. The depth of unstable air is often sufficient to give thunderstorms: one is reported in the past weather at Dyce (Aberdeen), and hail is falling there—and also in Northern Ireland—at the time.

In summer Arctic air provides somewhat cooler interludes: it is quite common in the north but rare in the south, where, when it does occur, with maximum temperatures in the 60's, less cloud and less frequent showers than in winter, it gives disappointing rather than unpleasant weather.

Polar Maritime

Ordinary Polar Maritime, the most frequent of our visiting air masses, often has its source also in the Arctic regions, but sometimes it is modified Polar Continental air that has moved eastwards from Canada. In either case, the air spends a considerable time in the North Atlantic and finally

approaches these islands from the north-west or west. In general it brings a type of weather similar to that of Arctic air but less intense in character. It is older air, no longer fresh from its source, more modified by its long sojourn over relatively warm waters. Moving towards the British Isles, it travels over still warmer seas and is a true PmK type, characterised by instability and convectional activity, but the temperature contrast between air and water is less marked and these characteristics less pronounced than in the case of Arctic air. Comparison of the temperature-height diagrams (Fig. 27) shows that the upper air conditions are similar except that the Pm curves are warmer, particularly in winter, though only slightly in summer.

Showers separated by bright intervals, more or less prolonged, sums up the typical weather of Pm air in contrast with the frequent showers of Arctic air. But an important distinction can usually be drawn between the weather of windward coasts and that experienced well inland or on lee-ward coasts. While in northern, north-western and western coasts, instability persists day and night, in the midlands and east, surface cooling at night imposes stable conditions which alternate with instability by day. This does not always occur with Arctic air. Both summer and winter ascents in Pm air (Fig. 27) were made from eastern stations (Leuchars in Fifeshire, Hemsby in Norfolk) and both show the characteristic nocturnal inversion. Day-time heating soon replaces this by a super-adiabatic layer.

In coastal areas towering cumulus or cumulonimbus are likely at any time, with cloud base becoming low in the accompanying showers, which may be of snow or sleet in winter. Thunderstorms and hail may occur in very unstable air, again at any time of day or night. Inland, however, Pm weather is often deceptive. The typical day dawns bright and clear, and the unwary go coatless or umbrella-less to work, while the un-weather-conscious housewife hangs up her washing. During the morning, usually earlier in winter, later in summer, the cumulus clouds appear, small and innocent at first, large and threatening later (Plate III). The showers, which may be thundery in summer, usually fall at the least opportune moment and emphasise the lesson to be learnt (Plate IV). In the evening the showers die down and the clouds disperse to give fine starlit nights.

Fig. 29 illustrates weather in a north-westerly stream of Pm air in autumn. Showery activity is widespread over western districts, further encouraged by orographic uplift over coasts and adjacent high ground. Notice the cloud types at Ronaldsway, Isle of Man, where it is showering at the time of the chart: this station reports low cloud type 9 (cumulonimbus), medium cloud type 6 (altocumulus formed by the spreading out of cumulus), and high cloud type 3 (anvil cirrus): these observations epitomise the showery sky. In detail, the configuration of relief influences the distribution of showers: for example, in a north-westerly stream such as

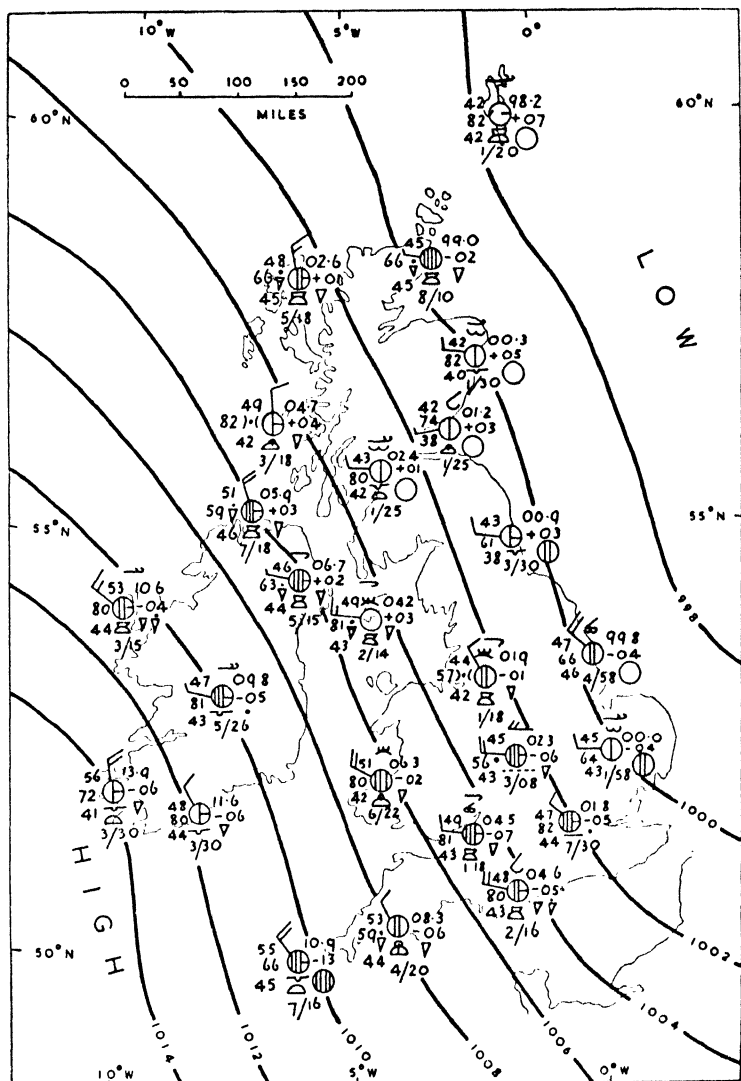


Fig. 29. Synoptic chart: Polar Maritime air, autumn (14th September, 1955, 0600 G.M.T.)

this, showers tend to penetrate, even at night, through the Midland Gate well into central districts. In the east cloud amounts are small, and what is reported consists mainly of remnants of convective cloud that has deposited the bulk of its moisture further west.

Pm weather is much less daunting than Arctic weather. In between showers, conditions are bright and sunny, and the rainbow often seen

against the departing rain-cloud opposite the sun adds colour and interest to the sky. In winter the air is cold but not necessarily freezing, in summer it is cool and bracing (see Table 4). With the constant stirring of air by convection and turbulence, dust and smoke are diffused upwards to great heights and the atmosphere has a sparkling clarity. Visibility is never better than during the bright intervals of Pm air, but it may deteriorate rapidly enough in heavier precipitation. Given warm clothing and some protection against showers, Pm provides good brisk walking weather even in winter, and quite reasonable weather for our outdoor sporting spectacles.

On occasions, however, Pm air becomes relatively stagnant in the grip of an anticyclone. With clear skies a large diurnal temperature range is experienced inland, and in autumn, winter, or spring, night frost may occur, aided by local topographical influences. Such conditions are favourable also to the development of radiation fog in this moist air. Fig. 30 illustrates such a situation: this anticyclone has developed quite typically between the departing cold front and advancing warm front of successive depressions (see Chapter 7). Temperatures have dropped to the twenties in some sheltered inland areas: skies are clear where they are not obscured by fog: the fog symbol is the dominant feature of the chart. Under these circumstances Pm air can be a nuisance as well as unpleasant. Movement by road and rail may be seriously hampered if not totally paralysed and aircraft grounded, while persistent fog, especially when aggravated by pollution, is physiologically distressing to many people and may be lethal to some (see Chapter 11). From the meteorological point of view, such a development illustrates the complete transformation of an air mass from the unstable to the stable state.

Polar Continental

Leaving aside Polar Maritime Returning air to be more conveniently dealt with later, we turn our attention to another source region, that of the interior plains and plateaux of Europe and Asia in winter. Here under the dominance of the great continental anticyclone there develops over surfaces of snow and ice the coldest air mass encountered on the face of the earth. In Polar Continental air over interior Europe the surface temperature may be 20° F. below zero (*i.e.* 52° of frost) or lower, but above this occurs a strong inversion and the air a few thousand feet above the surface may be 10° or 15° warmer. Due again to its continental origin, the air is extremely dry. Very cold, clear, dry weather is typical of Pc air over its source region.

Occasionally, rarely more than once in a winter, the 'Eurasian High' sends out a tongue westwards and Pc air may then invade the British

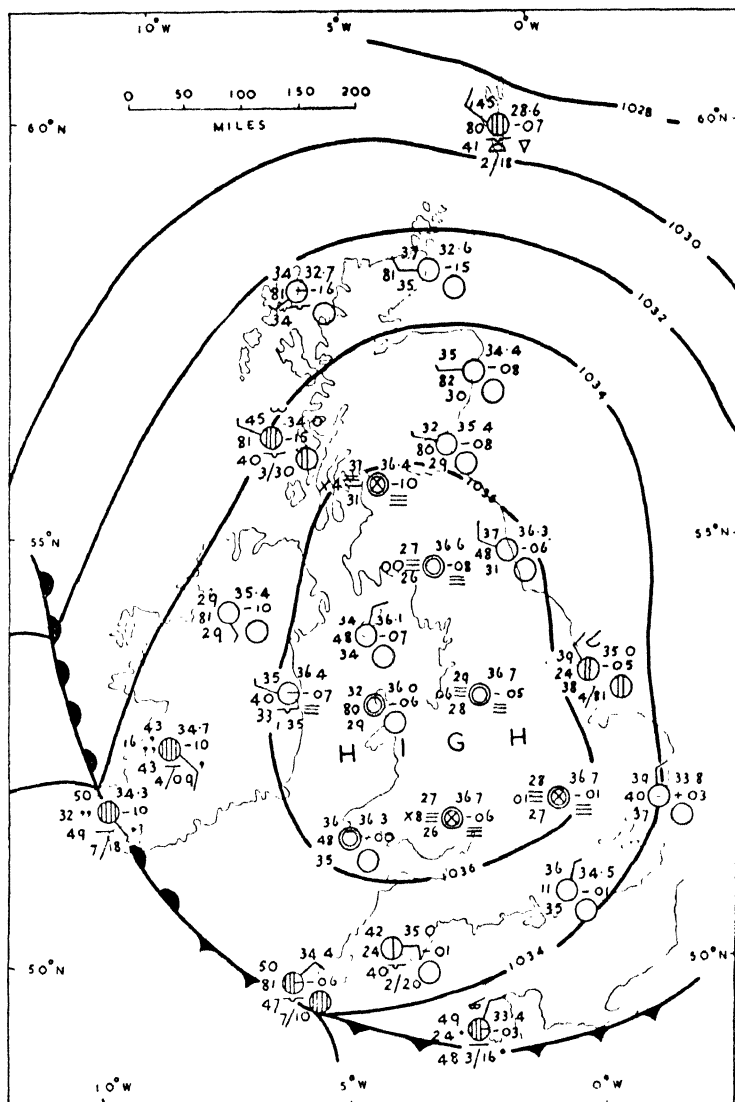


Fig. 30. Synoptic chart: Polar Maritime air, winter anticyclone (17th November, 1954, 0600 G.M.T.)

Isles. The weather it brings depends largely on the track it follows and the opportunities this offers for the modification of its original properties. Sometimes the air approaches England from east or south-east so that its path is mainly overland, including only the English Channel or the narrower southern part of the North Sea. In this case it arrives relatively

unmodified. Sometimes, however, it reaches us from north of east, having crossed a wide stretch of North Sea, in which case the degree of modification is considerable.

The 'short sea-track' variety of Pc air is illustrated by the upper air ascent (Fig. 27). While the air as a whole is no longer as cold as at its source, surface temperatures remain intensely low and the inversion aloft is maintained. Although some moisture has been acquired in the very lowest layers, no stirring is possible under these extremely stable conditions and the air above is dry. With day-time warming, a little shallow cumulus cloud can form beneath the inversion, but frequently skies are perfectly clear night and day.

The synoptic chart (Fig. 31) catches a moment during the great invasion of Pc air in 1947. The isobars tongue north-westwards over England like an icy finger, and over most of the chart the air is effectively of the short sea-track type. Temperatures are as low as 10° – 12° in places, and skies are cloudless, except for small patches of high cloud (perhaps remnants of some long-decayed frontal cloud). Visibility is poor in places and mist is widely reported, but only one station actually has fog: extensive radiation fog is not usual in this air—unlike Pm—because of the low humidity, but urban and industrial smoke sometimes contribute to local fog formation.

In the south-west and extreme north of the British Isles, the air in this example reaches land having crossed a considerable stretch of water, and the higher temperatures, increased cloudiness and even occasional showers in these areas give a hint of the rather different weather that prevails with Pc of the long sea-track type. More characteristically this air leaves the Continent to approach Britain from east-north-east or north-east, and thus crosses water as warm as 40° F. or more over large areas. Considerable acquisition of warmth and moisture must occur. If the warming from below is insufficient to destroy the original inversion, turbulence and convection may combine to produce a layer of stratus cloud below it. This North Sea Stratus, as it is called, often spreads inland into eastern districts. On the other hand, the warming process may entirely eliminate the inversion, in which case the air is transformed into a kind of Pm bringing higher temperatures, convection cloud and snow showers to east coast districts.

Most of the lowest temperatures recorded in southern England at least have occurred in Pc air, especially of the mainly overland track type. Although this kind of air is a rare visitor, once arrived it tends to persist for longish spells during which even maximum temperatures may remain down in the twenties, and we experience briefly the type of winter weather that is normal to the citizen of Warsaw or Berlin. In more modified Pc day-time temperatures often climb into the thirties, giving higher average

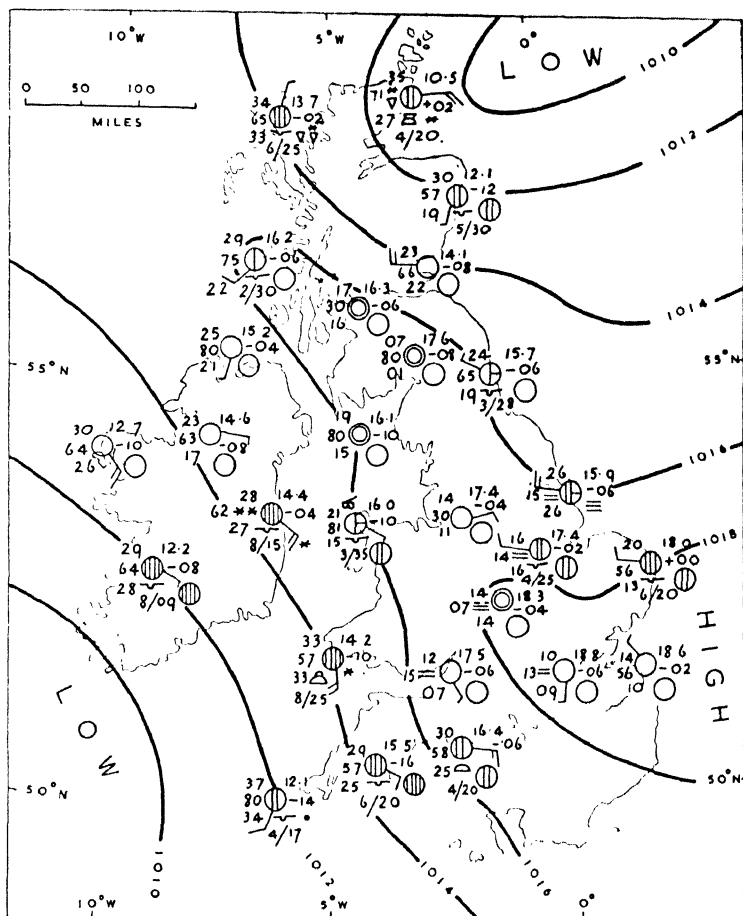


Fig. 31. Synoptic chart: Polar Continental air (24th February, 1947, 0600 G.M.T.)

values for the air mass as a whole (Table 4). Even the intense cold is not necessarily unpleasant as long as the air is still: indeed many people find this clear crisp weather stimulating (Plate V) though sensitive skins tend to become chapped in the dry cold. With strong winds, even if temperatures are higher, the fireside seat becomes definitely more attractive.

Ice deposits on the ground are often part of the scene in Pc air. Frost (*i.e.* a screen temperature of less than 32° F.) may occur in any Polar air mass, but, of course, actual condensation in the form of ice crystals (white frost or *hoar-frost*) requires also sufficiently moist air. If, as with a little-modified Pc type, the air is very dry, *i.e.* has a very low dew-point, the temperature may never fall low enough to reach it, even if below 32° F.

The result is then a statistical frost that leaves no sign on the ground (a so-called *black frost*). In moister Pc air fog may develop under freezing conditions, in which case the water droplets composing the fog are super-cooled (Chapter 3): they tend to freeze on contact with objects like trees, fences and telegraph wires, covering them with deposits of opaque ice or *rime* which often grow out on the windward side (Plate VI).

Less picturesque and pleasant is *glazed frost* or *glaze*, which occurs when a long residence of Pc air has left surfaces with temperatures well below 32°: if rain then falls (*e.g.* from advancing frontal cloud) it freezes instantaneously on falling to give a coating of smooth transparent ice on the ground and other objects. The heavy weight of these ice deposits may be sufficient to snap branches off trees and can play havoc with overhead wires, while the points on railway lines ice up and slippery roads cause a crop of accidents. This is happily a rare occurrence in Britain though many people in the west midlands will remember the glazed frost at the end of January 1940.

True Pc air can develop only in winter for only then are the seasonal anticyclones strongly established over the continental interiors. This is not to suggest that air may not reach the British Isles by easterly routes at other seasons. We may have modified Arctic or Pm air bringing unseasonable cold in spring, summer or autumn, or air of diverse origins may warm up considerably in summer over southern Europe and reach us from the south-east as a warm air mass akin to Tropical Continental.

Tropical Maritime

Originating over North Atlantic waters under the influence of the Azores anticyclone, Tropical Maritime air is initially warm, moist and stable. In its source region sea temperatures are around 70° F. in summer and 60° F. in winter, but in the waters off our south-western coasts they are some 10° lower, so that during its travel north-eastwards the air is cooled from below and is properly designated TmW. This modification emphasises further the stability of the lower layers and often creates, or strengthens an existing, temperature inversion. Because of their warmth these lower layers can acquire a great deal of moisture and the total water content is on average higher than in any other air mass: but the general stability discourages upward transport of moisture and the middle and upper layers are relatively dry. This is shown particularly by the summer ascent in Fig. 27.

Sea (advection) fog or low stratus cloud characteristically form in this stable air as its bottom layers are chilled to or near dew-point by the cooler waters off our south-western coasts. The base of this turbulence cloud may be anything up to a thousand feet up but is often a mere hundred or two feet, and sometimes right down on the surface (*i.e.* fog—

see Plate VII). Cornishmen are familiar enough with the fog banks rolling in towards their shores, and early seaside visitors too, for the frequency of sea fog is highest in late spring and early summer (when the contrast between air and sea temperature is greatest). The summer upper-air example, an ascent from Camborne in Cornwall, shows saturated air in the lowest layers, suggesting a fog sheet about 500 feet thick.

The characteristics of Tm air on crossing the coast depends on the season, the time of day and the form of the land. In winter, especially at night, the land is colder than the sea, and the low cloud persists and even thickens. Advection fog may spread inland, especially over very cold surfaces, but often it is lifted by increased turbulence over land and converted into low stratus. The typical winter pattern of Tm weather, as shown by the synoptic chart (Fig. 32), is one of much low cloud everywhere, shrouding high ground in 'hill fog,' with a tendency for stratus to break and lift to stratocumulus in the north and east, that is, after a long land track (Plate VIII). The only form of precipitation is usually orographic drizzle or light rain induced by the local thickening of cloud over cliff coasts and high ground inland. In winter, the cloud sheet may remain unbroken over wide areas and little diurnal temperature variation is possible beneath it.

The warmth of summer radically alters the potentialities of Tm weather for during the day surface heating is sufficient to eliminate the 'W' nature of the air mass everywhere away from the coastal strips. Because of this, summer sea fog is rarely able to penetrate more than a mile or so inland in day-time, while well inland the low cloud sheets are well broken by warming from below and may disperse entirely. Further insolation transforms the initial stability of the lower layers into instability, and turbulence cloud is often replaced by convection cloud. This may be only shallow fair-weather cumulus, but there is ample moisture available for considerable cloud development, and continued heating may give rise to strong convection and even to air-mass thunderstorms. In the evening convection cloud disperses and if night cooling is sufficient the turbulence cloud may well re-form. Much depends on the age and track of the air: the reversion to stratus is less likely well inland where nights may be fine and clear.

Tm air brings our milder winter weather, when even in January maximum temperature may be high in the fifties and minima too may be over 50°, though averages are rather lower than this (Table 4). The high relative humidity, however, gives the air a mugginess that is less welcome. Most of us find such conditions enervating and sigh for the freshness of Pm air at its best. Many people claim that aching corns, rheumatic pains and 'twinges' of various kinds accompany—or even presage—the change from Polar air to Tm (although the actual evidence available is confusing). Houses that are not kept continuously warm in some way (which means

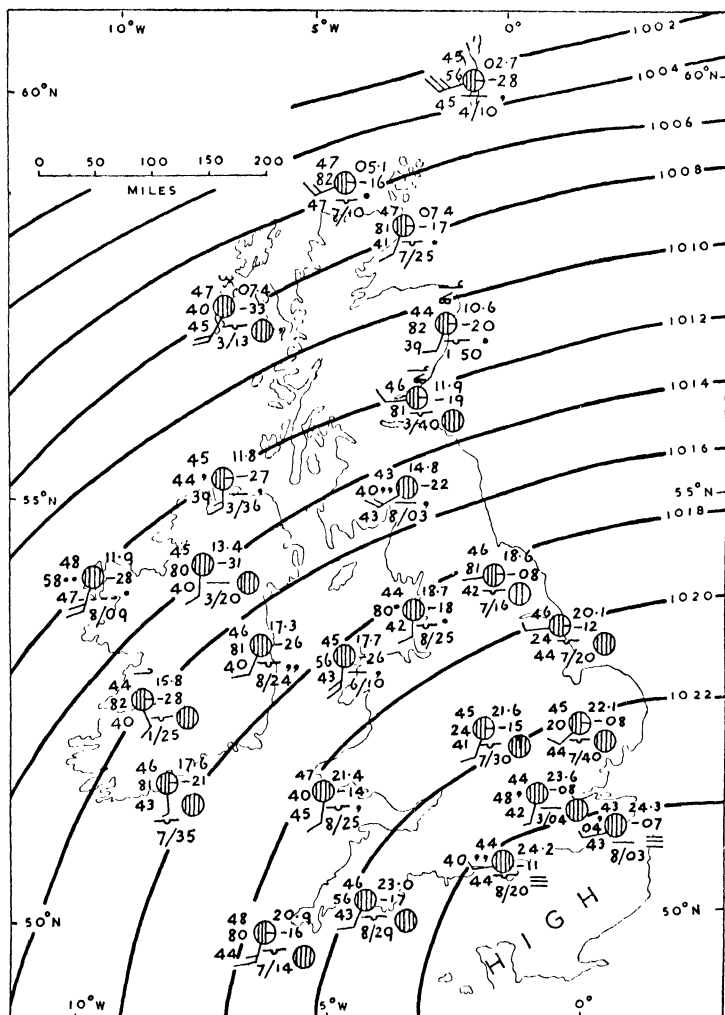


Fig. 32. Synoptic chart: Tropical Maritime air, winter (30th January, 1953, 0600 G.M.T.)

most houses in this country) also react to such a change, as the arrival of warm moist air in contact with previously chilled walls gives rise to copious condensation.

In summer, T_m air can again be oppressive especially at night when, with temperatures often over 60° F., sleep may well be difficult. The combination of high temperature and high humidity creates difficulties of various kinds. Bacteria become highly active: milk goes sour and food-

stuffs rapidly deteriorate. On the farm, pests thrive and fungal diseases quickly spread. For example, a period of two or more days with temperatures always above 50° F. and relative humidities above 75 per cent is generally followed by an outbreak of potato blight two to three weeks later. At such times the farmer is busy spraying or dusting with insecticides and fungicides.

At its best, however, Tm air provides the kind of weather we always demand of our summers and only rarely obtain. When anticyclonic conditions establish themselves in this kind of air, most parts of the country enjoy long spells of sun and warmth—the ‘heat waves’ of popular parlance—with day-time temperatures soaring into the seventies and eighties and the sales of ice-cream and cold drinks mounting accordingly. Relative humidity may be high during the night but is much reduced by the heat of the day. Fig. 33 illustrates such a situation with a ridge (or extension) of high pressure from a centre over Scandinavia covering this country. Even at 0600 G.M.T. temperatures are around 60° F. generally, with the notable exception of east coast areas. Here moister air off the North Sea has brought low cloud or fog which has kept temperatures near their minimum values and relative humidities correspondingly high. From the same cause low stratus has crept patchily inland into eastern and southern districts, while elsewhere cloud amounts are negligible. During the day most of this cloud will disperse except in some east coastal areas which may not share the fine weather enjoyed by the rest of the country. When such fine Tm spells coincide with Whitsun or August holiday week-ends, many townsmen find the call irresistible and there is the traditional exodus to the country or the coast and record numbers of cars are counted on the roads. Sometimes, depending on the synoptic situation, as Fig. 33 reminds us, it is possible to select the wrong coast.

These fine spells apart, visibility is generally poorer in Tm air than in most Polar air and familiar landscapes take on a hazier, less sharply defined appearance. This is understandable in moister air, seldom far from saturation, and in which impurities are confined by stability to the lower layers. Tm air usually needs little enough inducement to give fog (or stratus cloud with more wind and turbulence): advection fog, ‘hill fog,’ or radiation fog (occurring inland at night if a clearance of cloud during the day allows sufficient radiative cooling) are all common, sometimes seriously hindering traffic by sea, air or land. Low cloud ceilings are an additional hazard to aviation. Otherwise, Tm gives generally dry if dull weather, apart from convectional thunderstorm rains occurring in summer under highly modified conditions and with local orographic effects. Investigations suggest that some exceptionally heavy falls (mainly in summer) must be attributed to forced uplift of very moist Tropical air over mountainous regions.

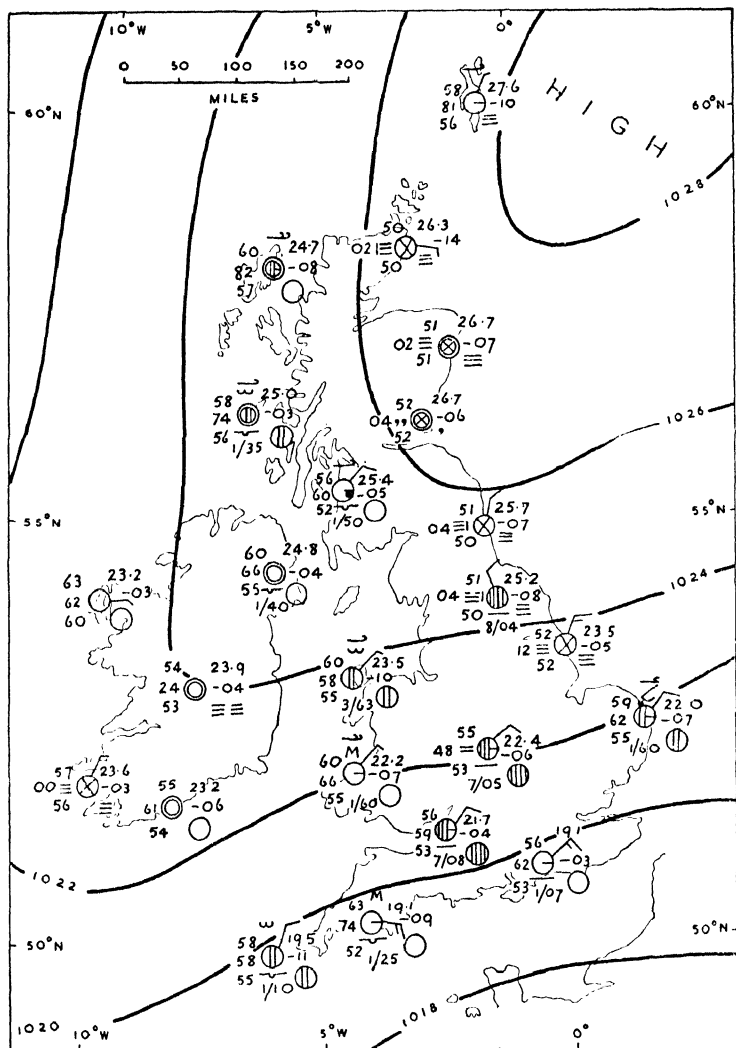


Fig. 33. Synoptic chart: Tropical Maritime air, ridge of high pressure in summer (10th July, 1955, 0600 G.M.T.)

Polar Maritime Returning

Consideration of this air mass has been left until this point because of its general similarity to Tm once it reaches the British Isles. It is Pm in origin and initial properties, but due to the vagaries of the pressure pattern it has moved further south than usual, south of latitude 45°. Here over warm waters it acquires some of the characteristics of Tm air and returns



PLATE IX. Frontal weather: fine tufted cirrus clouds, possibly the forerunner of a warm front



PLATE X. Frontal weather: skies clearing over the Brecon Beacons after the passage of a fast-moving cold front

north-eastwards towards the British Isles as a warm air mass, becoming more stable in its lower layers. It arrives as a somewhat cooler version of Tm, giving much the same weather. Its true nature as a Polar air mass, however old, is often betrayed well inland on summer days when the turbulence cloud disperses and, once convection is initiated, large cumulus readily grow into the colder upper layers and showers may be expected: the air then resumes the characteristics of ordinary Pm.

Tropical Continental

True Tropical Continental air only rarely invades these islands. Over its source region in the North African desert (and secondarily the Spanish Plateau in summer) this air is extremely hot and dry and often dust-laden during the day due to strong convection, although stability returns at night. If a southerly flow moves this air towards the British Isles over a mainly land track, we may scorch under arid skies made pallid by a characteristic dust haze. Or the air may spend a considerable time over the Mediterranean Sea, when much moisture is picked up, and under these conditions (illustrated by the upper-air sounding in Fig. 27) Tc may contribute to summer thunderstorms in Southern England. Because of its origin, Tc is warmer in its lower layers than Tm and gives us our hottest summer days, with maxima near or over 90°, and nights, with minima over 65°, that may be too warm for comfort. Probably, authentic Tc air reaches us in summer only, and warm air with a similar southerly or south-easterly track in winter may be much-modified Tm.

The Frequency of the Different Air Masses over the British Isles

Some idea of how often and at what time of year these air masses and their characteristic weather are likely to occur may be gleaned from Table 5, which is a simplified version of data compiled by Belasco. The Table gives, for each month, the average percentage frequency of occurrence of each air mass at three widely separated stations, Kew, Scilly and Stornoway (Hebrides). A comparison of the figures for these stations, situated approximately at the south-eastern, south-western and northern corners of Britain, shows that significant differences in the length of occupation by different air masses may be found even in a relatively small country. Here lies the reason for regional differences of weather and climate within the British Isles. Readers living between the far-flung points represented in the Table can make appropriate interpolation for their own areas.

The values have been derived from an examination of 12-years' records, which is not long enough to yield reliable averages, but they may be taken as a useful guide. Not all the air reaching the British Isles could be classi-

TABLE 5

Percentage Frequency of Different Air Masses at Kew, Scilly and Stornoway

KEW

Air	J.	F.	M.	A.	M.	J.	J.	A.	S.	O.	N.	D.	Year
A	7	6	7	7	14	6	5	5	6	4	6	5	6.5
Pm	20	24	16	25	17	32	33	31	25	21	28	24	24.7
Pmr	14	10	7	8	14	6	10	8	8	10	10	14	10.0
Pc	7	6	1	0	0	0	0	0	0	0	0	3	1.4
Tm	11	13	12	6	6	9	11	7	7	8	12	12	9.5
Tc	3	4	5	5	6	3	4	5	3	9	7	2	4.7
F	12	8	10	13	11	11	12	13	13	9	12	11	11.3
H	15	19	31	31	19	30	23	24	35	25	17	22	24.3

SCILLY

A	4	3	3	4	11	4	3	6	4	2	5	1	4.2
Pm	29	15	18	28	25	34	38	34	25	26	29	29	27.5
Pmr	16	6	12	9	9	5	8	10	11	13	9	11	10.0
Pc	3	6	0	0	0	0	0	0	0	0	0	2	0.9
Tm	17	20	13	9	8	11	15	11	9	12	19	18	13.5
Tc	2	2	4	3	2	1	1	2	3	5	6	1	2.7
F	9	12	11	11	15	13	13	13	12	10	12	11	11.8
H	11	15	30	30	20	29	20	19	34	21	14	22	22.1

STORNOWAY

A	13	8	10	13	16	14	12	16	9	7	12	5	11.3
Pm	31	30	27	31	23	32	35	31	41	31	31	35	31.5
Pmr	22	20	12	11	15	14	15	14	15	16	16	22	16.0
Pc	1	5	1	0	0	0	0	0	0	0	0	1	0.7
Tm	6	10	13	11	5	6	6	6	7	11	13	10	8.7
Tc	0	1	2	2	1	0	1	0	1	3	3	1	1.3
F	10	9	10	11	15	14	20	14	8	9	12	10	11.8
H	11	11	15	16	19	18	10	15	15	17	7	12	13.8

fied, which explains why the percentages in any one month do not add up to 100: most of this indeterminate air came from the Continent, without, however, fitting readily into either the true Continental types. To the air masses as classified earlier in this chapter have been added two further categories. F denotes air in the vicinity of fronts, *i.e.* essentially conditions transitional between one air mass and another: H represents anticyclonic air, *i.e.* air in or near the central regions of an anticyclone where the weather characteristics depend less on the origin of the air than on the dynamics of this particular pressure system. Discussion of fronts and anticyclones belongs to later chapters but in passing we may note from Table 5 that the frequency (though not necessarily the intensity) of frontal weather seems practically the same everywhere in the British Isles, but that anticyclonic weather is much less frequent in the extreme north than in the south.

The tables show clearly enough the extent to which our weather is dominated by maritime air of one kind or another. Even excluding that involved in anticyclones, maritime air is with us for between one-half and two-thirds of the year, the frequency increasing westwards and, especially, northwards. On the other hand, well-defined continental air is relatively rare. Kew and south-east England generally are most likely, for obvious reasons, to experience both the cold Pc of winter and the hot Tc of summer, but even here a year may pass without a visitation from one or other of these air masses. Continental air is even less common at Scilly and hardly ever reaches Stornoway. This is the basis of the distinction between the more maritime west and the more continental (but still not very continental) east.

Of the maritime air masses, polar air is more important than tropical, and ordinary Pm is the most frequent single air mass: if Table 4 is re-examined it will be seen that the average temperatures for all air masses (*i.e.* the ordinary long-period averages) are closer to those in Pm air than in any other air mass. If we combine Pm and A as giving cool or cold unstable weather, then these persist for about a third of the time in the south (Kew and Scilly) but for over 40 per cent of the time at Stornoway, where the frequency of Arctic air is understandably greater than elsewhere. If Pmr air is added to Tm, both bringing mild or warm stable conditions, the differences among the three stations are small: Kew has the lowest quota (just under 20 per cent) and Stornoway the highest (nearly 25 per cent) though here Pmr is nearly twice as frequent as Tm and there are years when no Tm is experienced at all.

Generally speaking, Pm and A air are more common in the summer than in the winter six months, while the reverse is true of Tm and Pmr. Nevertheless, each of these is an all-seasons air mass and only Pc is really confined to one particular season. This air is a very rare occurrence in December, more common in January and February (in some areas February more than January) and is even apt to provide a last snarl of winter in March. Tc is mainly a summer visitor in its most characteristic form.

FRONTS AND FRONTAL DISTURBANCES

When the air mass in occupation of our region moves away to be replaced by another, we observe the transition from the properties of the first air mass to those of the second. This relatively rapid change of temperature, humidity and other characteristics occurs within a more or less sharply defined boundary zone which, in modern meteorological parlance, is called a *front*. But more is involved than a discontinuity in air-mass properties. If the word 'front' conjures up a picture of a fluctuating battle line between opposing forces, then this is to some extent an apt analogy. Fronts are the battle lines of the atmosphere, along which contrasting air masses contend for possession of the underlying region. And just as military fronts are zones that periodically erupt into battle after periods of 'all quiet,' so atmospheric fronts, while sometimes quiescent, are more often belts of disturbed weather. Most of the really bad weather we experience in this country we may rightly blame on to fronts and, as was shown in Chapter 5, we suffer frontal weather for something like one-eighth of the time.

Air Mass Boundaries

The military analogy helps us no longer when we examine in detail the nature of air-mass boundaries and consider exactly how neighbouring air masses lie side by side. If two liquids of different densities which do not mix (like oil and water) are poured simultaneously into one vessel the lighter liquid floats above the heavier: in other words, there is a horizontal boundary of separation. Colder and warmer air masses are, from this point of view, heavier and lighter fluids respectively and, were the earth a stationary body, the colder air would completely underlie the warmer. But the earth is rotating and other forces in addition to gravity are involved. In fact, adjacent air masses are separated by sloping boundaries (see Fig. 34). We could obtain a somewhat analogous effect by gently swishing round the oil and water in the vessel mentioned above (the point is explained more fully in Chapter 7).

An air-mass boundary thus takes the form of an inclined *frontal surface* meeting the ground along a horizontal line, which is how we see the front

in plan, on a synoptic chart: in cross-section we see it as a sloping line. In all diagrams such as Fig. 34 we are obliged to exaggerate this slope many times for in reality the angle of inclination is always very small. A mathematical formula has been devised, based on certain assumed conditions, which shows that the angle will be steeper (1) the higher the latitude (2) the smaller the temperature difference between the two air masses and (3) the greater the difference in speed of movement. In actual examples the slopes of recognisable fronts vary from about 1 in 25 to about 1 in 300, with 1 in 100 as something like an average value in our latitudes.

While we commonly speak of frontal surfaces, it would be wrong to visualise these as sharply defined planes of discontinuity in the atmosphere. A balloon sent up from point A in Fig. 34, c, passes first through the cold air, then through a turbulent transition layer or *mixing zone*, some thousands of feet thick, and finally into the warm air above. In Fig. 35 an actual ascent through a well-marked front (not all examples are as

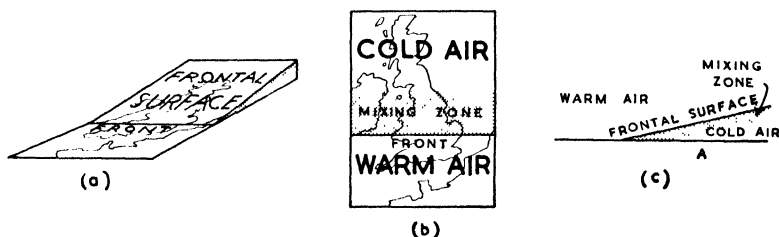


Fig. 34. The structure of a front seen (a) in block diagram, (b) in plan, (c) in cross-section

clear-cut) shows the characteristic composite temperature curve. The mixing zone shows, in this case, an increase of temperature with height (frontal inversion) but in many others is represented merely by an isothermal layer or one with a much reduced lapse rate. The characteristic moistness of the air is due, as we shall show, to the associated cloud and rain.

What we rather loosely call the 'line' of the surface front thus represents in reality a belt of some width, anything from a few miles to about 50, depending on the slope of the frontal layer. For this reason, abrupt changes in air-mass properties at fronts are not to be expected either on the weather map or in our personal experience of frontal weather. However, on the scale of the usual synoptic charts it is justifiable to represent fronts as lines rather than belts, and it is certainly convenient to do so for practical purposes of analysis and forecasting. Usually the line is placed at the edge of what is beyond doubt the warm air, *i.e.* the mixing zone is conventionally regarded as belonging to the cold air. In fact some fronts are fairly sharply defined while others are so diffuse that they are scarcely recognisable either on the map or in the atmosphere.

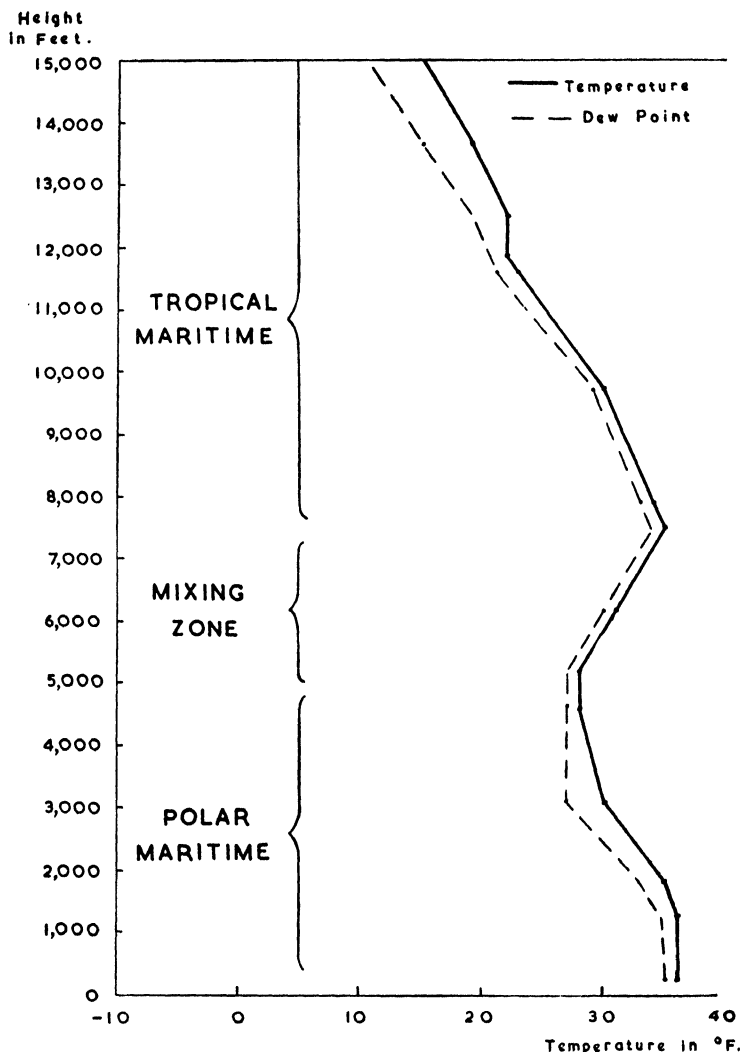


Fig. 35. Upper-air sounding through a front. Aldergrove, 22nd December, 1952, 0300 G.M.T.

Active Fronts

Nothing that has been said so far explains why fronts are frequently belts of bad weather. Indeed in the simplest situation, the front has no movement, warm and cold air masses flow side by side in parallel streams and there is no reason for cloud or rain to occur. Quite passive fronts of

this kind are sometimes observed on the synoptic chart, representing a transition between air masses but giving no frontal weather. More often, however, this balance is disturbed and there is a movement of either air mass, or of both, across the front causing a congestion of air which is relieved only by a removal of some of it in the upward direction. This implies in effect the ascent of warm (relatively light) air above the wedge of cold (relatively dense) air. Rising air at an active front, like rising air due to convection or to turbulence or to orographic influence, is subjected to adiabatic cooling with all its consequences for the formation of cloud and the release of precipitation.

Fronts may themselves move, in response to the push of the air mass on one side or the other, and according to their direction of movement are described as warm or cold. At a *warm front* the warmer air mass is advancing and the wedge of colder air retreating, so that an observer standing initially in cold air later finds himself in warm. With an active warm front, the warm air advances more rapidly than the cold air recedes so that warm air must climb above the wedge of heavier air, giving rise to characteristic warm front cloud and weather.

At a *cold front* it is the cold air that advances, covering ground lately occupied by warm. If the cold front is active, the cold air wedge undercuts the warm air and compels it to rise: this results in typical cloud and precipitation forms which may differ from those associated with warm fronts. Fronts with no movement cannot be classed as warm or cold: they are simply stationary or, in practice, nearly so (*quasi-stationary*). It is also possible for a front to reverse its direction of movement and thus change from warm to cold or cold to warm.

A moving front is not necessarily an active one. If there is no relative motion between the two air masses across the front, so that one does not gain on the other, then there is no reason for the ascent of warm air and consequently no cloud or rain. In practice, as we should expect, some fronts are more active than others. A strong front will be marked by thick cloud development and much precipitation, a weak front may yield little precipitation or none and may be marked merely by a belt of frontal cloud.

The Major Frontal Belts

The pattern of fronts over a large area at any one time may be most complex, as a glance at almost any Northern Hemisphere chart of the Daily Weather Report will confirm (see, for example, Fig. 49). The average positions of the major fronts during mid-winter and mid-summer respectively are shown in Figs. 60, *a* and *b*. We shall return to these maps (Chapter 7) in connection with the general circulation of the atmosphere

and we make use of them at present only to note briefly some relationships between fronts and air masses, bearing in mind that they represent mean conditions and are unlikely to resemble any single synoptic chart.

Most significant for our own weather is the *Polar Front*, a constant feature of the North Atlantic weather map, though far from constant in position and intensity. With its continuation in the North Pacific Ocean this major air-mass boundary may at any time be traced more than halfway round the earth in middle latitudes. It demarcates Polar air (which may be maritime or continental in origin) from Tropical air (generally maritime) and, thus separating air masses of different mean temperatures, is well defined and often vigorous. There is a corresponding Southern Hemisphere Polar Front.

In high latitudes are found the *Arctic Front* and its Southern Hemisphere counterpart, the *Antarctic Front*. The Arctic Front separates Arctic (or direct Pm) air from less fresh Pm, that is to say, air masses of similar origin distinguished only by different degrees of modification. The air mass contrast is not therefore usually as strong as in the case of the Polar Front, and the Arctic Front is less well marked a feature. The Antarctic Front, however, divides very cold Pc air originating over the Antarctic ice-cap from much warmer Pm air further to the north.

The remaining major front, which more or less encircles the globe in very low latitudes, is the *Intertropical Front* (ITF), sometimes referred to as the Equatorial Front. The ITF is not always a front in the normal sense. Over the oceans the air masses on either side are almost identical in properties (TmK) and the necessary contrast is lacking. Over or near land, however, one of the air masses may have Tc characteristics, and more of a contrast then develops. Again, at or very near the Equator there can be no frontal slope (see page 118) but when, as does happen, the ITF is drawn into higher latitudes (as part of monsoon circulations—see Chapter 8) a more typical frontal structure can form. In any case, notwithstanding these differences, the ITF (concerning which there is much more to be learned) represents a zone, more or less broad, into which air flows from both north and south, creating an excess that must be relieved by upward movement, with well-known consequences for the weather of low latitudes.

The mid-winter and mid-summer pictures differ mainly in the position of the fronts which tend to shift with the sun, moving polewards in summer and equatorwards in winter. They also generally weaken in the summer season. Otherwise, the only striking difference between the two maps is the existence of a kind of subsidiary Polar Front in the Mediterranean area in winter. This *Mediterranean Front* forms between Pc air from the European interior and Tc air from North Africa. In summer this clash of air masses does not occur.

The Development and Decay of Fronts

Comparison of a few consecutive weather maps will demonstrate how widely these fronts may fluctuate from their mean positions, and closer examination will show too that they vary considerably in intensity both from time to time and from place to place. Here a front may be vigorous and the contrast in air-mass properties sharp, while further along not only is there little frontal weather but the air-mass contrast is so gradual and blurred that the front is difficult to trace. In fact, fronts frequently fade out (Fig. 49) and, while it may seem easy to join up the loose ends in some instances, the forecaster has obviously chosen not to do so because it would be meaningless in terms of the resulting weather. On the other hand, fronts may form other than at the major air-mass boundaries: there are, in fact, both general and local frontal belts. The processes by which fronts develop where none existed before or by which existing fronts are

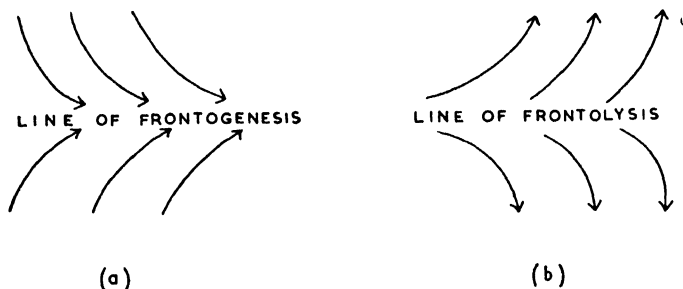


Fig. 36. Air flow patterns favouring (a) frontogenesis, (b) frontolysis

intensified are summed up in the term *frontogenesis*: conversely, the decay of fronts is known as *frontolysis*.

The conditions favouring frontogenesis have been closely studied by the Scandinavian meteorologists, especially by the Norwegian Sverre Petterssen. In the first place there are involved two air masses of different properties, of which temperature is by far the most important. Secondly, the pattern of air flow must transport these air masses towards each other so as to sharpen the transition zone between them. As we have seen, this entails a net accumulation of air, relieved by ascent above the front. An air flow that would encourage frontogenesis is suggested in Fig. 36, a.

As far as the major fronts are concerned, the requisite conditions stem from the operation of the general circulation (see Chapter 7) which provides regions of low pressure towards which air of different origin is drawn. If these fronts move over parts of the earth's surface where temperatures change rapidly within relatively small distances, frontal contrasts are apt to be intensified (frontogenesis). Such favoured areas include the boundary zones between warm and cold ocean currents (e.g. the Gulf

Stream and Labrador Current off the East coast of North America), the fringe of ice-caps or snow-covered land (e.g. the Greenland and Antarctic plateaux), or even coastlines separating warmer land from cooler water or vice versa. Local frontogenesis, in the sense of the creation of new subsidiary fronts, may also occur over such areas, provided the right pattern of air flow is established.

In contrast, frontolysis occurs either when the temperature contrast between adjacent air masses is eliminated by modification from below or if the air flow acts so as to diffuse the transition so widely that ultimately no front is recognisable. Fig. 36, *b*, shows a type of air flow that would entail frontolysis. Instead of an excess of surface air and resulting ascent, the outflowing pattern leaves a deficiency that can be made up only by the descent of upper air into the frontal zone: this inevitably blurs the frontal contrast. This type of flow involving descent of air is characteristic of anticyclones and it can be seen from Fig. 49 that the breaks in the fronts, *i.e.* the frontolysed portions, occur in the regions of high pressure. As descending air is necessarily warming and drying air, the frontal cloud tends to disperse under such conditions. If, in addition, the two air masses lie stagnant side by side over a uniform surface they will both be subjected to like influences from below which will further reduce the contrast between them.

The Weather of Warm and Cold Fronts

The weather characteristics of active fronts, warm or cold, depend on large-scale ascent of air: this may give rise to extensive cloud with its base not far above the ground and its top limited sometimes only by the tropopause, and in this thick frontal cloud conditions are generally ripe for the release of considerable precipitation. Since it is in all cases the warmer air that is lifted, the thickness and type of cloud and in fact the intensity of frontal weather, are dictated partly by the nature of the warm air mass, that is to say, its moisture content and stability conditions. Apart from this, it might be thought that, because warm and cold fronts differ basically only in their direction of movement, the warm front cloud and weather system is identical with that of the cold front but, so to speak, in reverse. On some occasions this is not far from the truth, but on many others there are certain differences in air flow which are reflected in differences in cloud structure and resulting precipitation. These differences we shall now consider in more detail.

The Warm Front System

At the warm front the ascent takes the form of a relatively gentle up-gliding of warm air above cold (see Fig. 37, *a*) which gives rise to stratiform or sheet clouds. The inclination of the warm frontal surface is

generally small, between 1 in 100 and 1 in 300 being typical: this means that the warm front cloud is often very extensive. The lower portion of an advancing warm frontal surface is held back by friction with the ground: this further flattens the already gentle slope of the front (Fig. 37, *a*) and, incidentally, makes the identification of the surface front both on the map and on the ground even more difficult.

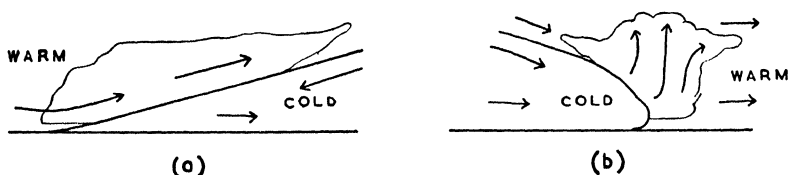


Fig. 37. Air movement at fronts: (*a*) warm front, (*b*) cold front

The warm front cloud extends as a wedge-shaped mass thickest near the surface front and lying mainly ahead of it. The base of the cloud coincides roughly with the lower edge of the warm air. Although the frontal cloud is conventionally portrayed as a continuous mass (as in Fig. 37, *a*), in practice it is very commonly a well-layered cloud, corresponding to variations in the moisture content or the stability conditions vertically in the warm air mass: these are generally apparent in the radio-sonde observations. The clear 'lanes' between the cloud layers lessen the problems of aircraft pilots. However, with very moist or unstable warm air, the frontal cloud may be solid to great heights. At the other extreme, weak warm fronts may be represented only by vestigial cloud layers at various levels.

The topmost clouds form high up on the frontal slope and therefore well ahead (commonly 5–600 miles) of the surface front. At the extremely low temperature of these levels (of the order of 25,000 or 30,000 feet) the air contains little moisture, and condensation takes the form of thin clouds composed entirely of ice particles. These are the wispy or feathery structures known as *cirrus* (Plate IX). Below these delicate and tenuous forms, sometimes merging with them, sometimes separated by clear sky, lies a denser and more continuous sheet of ice cloud, *cirrostratus* (occasionally mixed with *cirrocumulus* if the air at these heights is unstable). Lower still come clouds of the medium levels (prefix *alto*), most typically of the sheet type (*altostratus*) but often also the less stable form (*altocumulus*), thickening progressively in the direction of the surface front.

These medium clouds contain both ice particles and super-cooled water droplets, and so it is from this part of the system that the first rain (or snow) will fall. This is generally slight and evaporates in the drier air below the front: eventually, however, the precipitation is heavy enough to

reach the ground and this may occur anything up to 200 miles ahead of the surface front. The increasingly dense rain-bearing cloud near the surface front is called *nimbostratus*: this represents a further stage of thickening from *altostratus* and is essentially the same kind of cloud with a lower base (*nimbostratus* and thick *altostratus* are denoted by the same symbol on synoptic charts).

Warm frontal precipitation is usually persistent rather than heavy. Beginning intermittently from medium cloud, it later becomes continuous and reaches moderate intensity before it finally ceases with the passage of the front. A warm front rain belt 150 miles wide advancing at 20 m.p.h. (a quite usual speed) will provide a good few hours of steady rain measuring perhaps quarter of an inch in the rain gauge. Moreover it is generally effective rain, steadily soaking the soil and, falling from cloud-laden skies, only little reduced by evaporation: it is welcome rain for farmers and gardeners after a dry spell. Weak fronts, however, yield little or no precipitation. On the other hand, where the relief of the land reinforces frontal lifting quite heavy rain may be experienced. If the warm air mass happens to be highly unstable, the resulting cloud shows considerable vertical development, with *altocumulus* replacing *altostratus*, and there may even be *cumulonimbus* embedded within the main cloud mass; these clouds, which introduce new hazards for aviation, not only yield locally intense precipitation but may give rise to frontal thunderstorms. Warm front thunderstorms are not however as common as those associated with cold fronts.

Sometimes the continuous rain near the surface front is accompanied by fog, thus adding further to the damp gloom of warm front weather. *Frontal fog*, as it is called, forms in the shallow cold air which becomes saturated as a result of continued evaporation from warm raindrops falling from the overlying *nimbostratus*. It is effectively a building down of the main cloud base to the surface. Frontal fog forms only if the temperature of the falling rain is substantially higher than that of the cold air (*i.e.* with a well marked front) and only if the winds ahead of the front are light. With moderate or fresh winds, turbulence lifts the saturation level and condensation then takes the form of ragged patches of *stratus-fractostratus* or *scud*—which appear to race across the sky at high speed because they are only a few hundred feet above our heads. Frontal fog may last several hours. Even if actual fog does not occur, visibility at and near the warm front is generally poor due to precipitation, and *scud* clouds or even the main *nimbostratus* base may shroud high ground in 'hill fog.'

Portrait of a Warm Front

In Fig. 38 a fairly typical warm front situation is illustrated both in plan and in cross-section. The synoptic chart shows a warm front, separat-

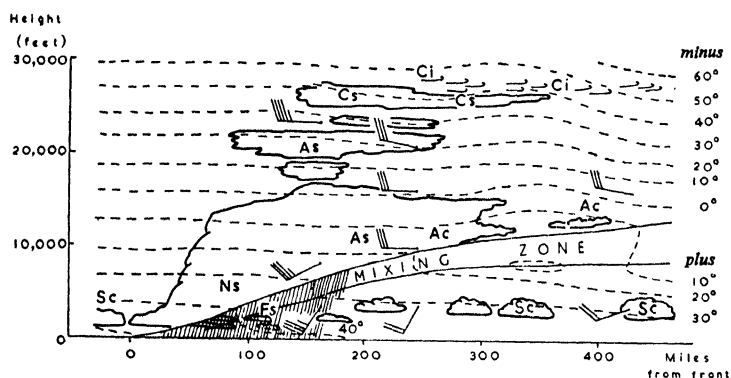
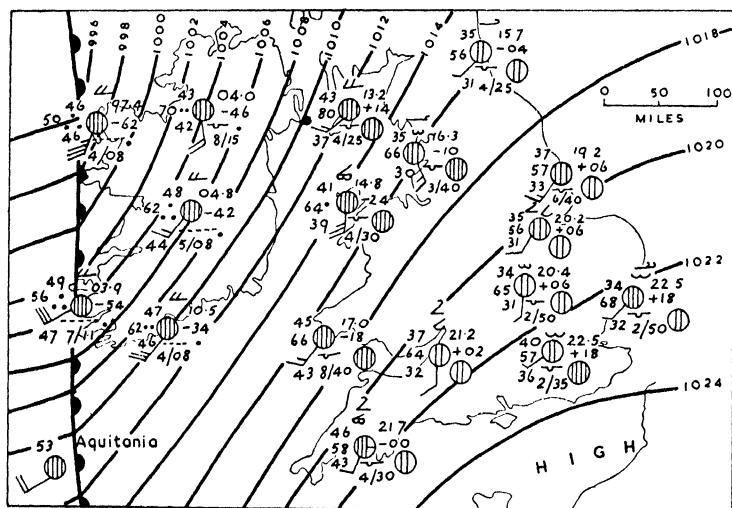


Fig. 38. Portrait of a warm front, 5th April, 1947, 0600 G.M.T. In the cross-section, broken lines are isotherms, and wind arrows give strength (see Appendix p. 258) and compass direction (assuming north at top of page) at various heights

ing Pm air from Tm, approaching the west coast of Ireland: the frontal cloud, measuring over 400 miles in width, and the rain belt, varying from 150 to 200 miles wide, extend eastwards over most of the British Isles. Comparison of temperatures and dew-points at the eastern and western edges of the map demonstrates the transition in air-mass properties. The characteristic cloud types appear in successive bands (see the Appendix for the meaning of the symbols), the high cirrus and cirrostratus over eastern England, the medium layers (altostratus and altocumulus) appearing below at first patchily and then more continuously further west, while over Ireland, where rain is falling, the main cloud forms are nimbostratus above and fractostratus below. The variable cloud (mainly stratocumulus)

occurring ahead of the frontal cirrus in East Anglia and to some extent further west, is decaying cloud (probably convectional) belonging to the cold air mass: frequently, this residual cloud though thin, is widespread and may mask the true frontal cloud above.

The cross-section, built up from both upper air and surface data (although it is partly diagrammatic as regards the detail of clouds) helps to complete a three-dimensional picture. The frontal slope, determined from the available radio-sonde observations, is on average about 1 in 200. The humidity data indicate that the frontal cloud is layered rather than solid. The thick medium cloud extends upwards only to about 15,000 feet and many warm fronts have relatively little dense cloud above this level. The run of the isotherms shows clearly the temperature change between warm and cold air: the downward dip of each isotherm as it crosses from warm air to cold denotes the transition or frontal zone. The wind arrows, plotted from the upper winds data, show that each of the two air masses has its own direction of travel and that the frontal zone is a region of wind change (from south-westerly to southerly in the cold air to nearer westerly in the warm air above). This wind change is also a feature of the surface synoptic chart. We shall refer in Chapter 7 to the behaviour of winds at fronts and also to the pressure changes associated with them.

Observations of the Passage of a Warm Front

As the frontal system moves it is clear that a suitably placed observer will experience a characteristic sequence of weather events. This enables us frequently to recognise the approach of a warm front some hours ahead. It is heralded first by the high cirrus wisps and plumes, visible above the residual clouds of the cold air. Not all cirrus is frontal. If, however, it appears to thicken into denser more uniform cirrostratus, the warning is not to be ignored. Cirrostratus is easily recognised. Only a filmy sheet, it allows the sun or moon to shine through quite brightly but characteristically surrounded by a *halo*. Solar and lunar haloes are wide luminous rings far out (22°) from the sun or moon, generally appearing white but sometimes showing the spectral colours through dark glasses: they are caused by the bending (*refraction*) of light rays by ice particles. The halo should be distinguished from the *corona*, a much smaller, vaguely coloured ring sometimes observed fringing sun or moon but always through water droplet clouds. Only thin continuous ice clouds can produce a halo.

Of the medium clouds that next appear, altostratus is a quite featureless sheet but altocumulus is distinguished by its more definite form. Thin altostratus allows the sun to shine through in vague watery fashion but as the cloud thickens further the sun is completely blotted out. By this

time it is usually raining and, as for the rest, the murky wet nimbostratus and fractostratus sky is too familiar to require description. The rain (or snow, or snow turning through sleet to rain, depending on the temperatures) is heaviest near the front but ceases when the front is through. The observer is now in the warm air and air-mass weather replaces frontal weather. In Fig. 38 these behind-the-front conditions are represented by the limited observation from the liner *Aquitania*, which shows relatively high temperature and cloudy skies.

In addition to these observed changes, the instruments at a station in the path of the warm front record (1) an increase in surface temperature (2) an increase in relative humidity and dew-point (further raised as a result of precipitation) (3) falling pressure and (4) increasing and *backing* wind (*i.e.* changing direction in an anti-clockwise sense). After the passage of the front, temperature and humidity become representative of the warm air, pressure either steadies or falls more slowly and the wind *veers* (changes in a clockwise direction). Continuous traces illustrating these changes are shown in Fig. 45.

Types of Cold Front

Cold fronts are much more variable in their habits than warm fronts. In many cases the wedge of cold air advances vigorously, butting up the warm air ahead so that the frontal clouds show considerable vertical development. In further contrast to the warm front, the effect of surface friction is to steepen the inclination of the frontal surface (see Fig. 37, *b*) so that slopes of 1 in 150 to 1 in 25 are more typical of the cold front. Partly as a consequence of the steeper slope, the cloud and precipitation accompanying the cold front are usually concentrated into a narrower belt than that associated with the warm front. This is further emphasised in many instances because back from the surface front (Fig. 37, *b*) there is a general downward motion, both of warm air moving down the frontal slope and of cold air spreading out and settling below it; this subsidence, with its consequent adiabatic warming and drying out of the air, limits the extent of the cloud in this direction and causes an often rapid clearance of cloud and improvement of weather (Plate X). For these reasons, the entire cold front system may measure only a hundred or two miles in width.

With this type of front, the cloud is cumuliform in character with an admixture of medium cloud, generally altocumulus formed from the spreading out of cumulus. This medium cloud tends often to extend ahead of the main mass with the stronger winds in the warm air at these heights. If the warm air is unstable, clouds may tower up in the form of mighty cumulonimbus which yield considerable precipitation of showery type,



PLATE XI. Thundery weather: towering cumulus and the spreading anvil of a thunder-cloud

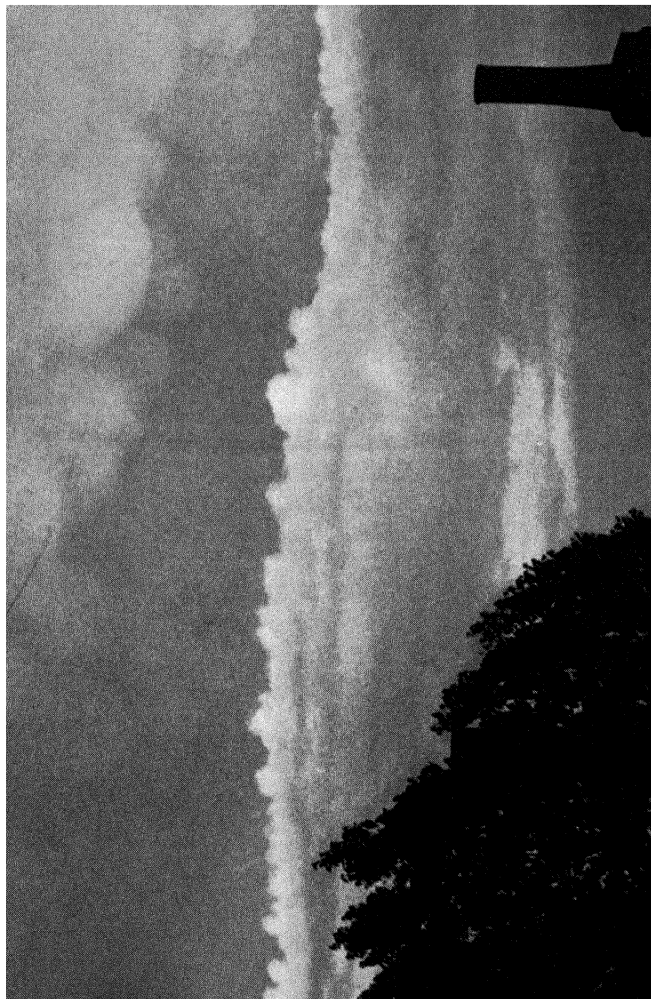


PLATE XII. Thundery weather: mammatus pouches hanging from an overhead cloud and a bank of altocumulus castellatus in the middle distance

often heavy while it lasts but of relatively short duration. Such intense precipitation, much of it effectively wasted by too-rapid run-off or by evaporation when skies clear, is not as useful as warm front rain from the farmers' point of view: indeed, it may be a positive nuisance by flattening ripened crops. Sometimes the vertical currents may be strong enough to give cold front thunderstorms indistinguishable from heat thunderstorms except in their origin and rapid passage.

With a rapidly moving cold front the frictional retardation of the frontal surface near the ground not only steepens the slope but may even reverse it (as in Fig. 37, *b*) so that cold air actually overhangs warm for a distance of perhaps 25 or 50 miles ahead of the surface front. This cold front 'nose,' as it is often called, constitutes a highly unstable arrangement with warm air lying below cold, which may contribute to the intensity of the accompanying weather. All along the front, warm air tends to surge upwards in energetic convectional activity, yielding showers and wind squalls of brief duration but often great severity—the so-called *line-squall* of pre-frontal meteorology.

At the other extreme, many cold fronts bring only small amounts of precipitation and sometimes none at all, although they may nevertheless mark sharp changes in air mass. Still other cold fronts, sometimes those advancing more slowly, show a rather different kind of weather pattern. The slower movement is reflected first of all in the absence of a 'nose' and its accompanying features and, secondly, in a gentler frontal slope, more like that of a warm front. Then the warm air is not forced vertically upwards with so much vigour but tends rather to slide up the frontal surface. There is not the same tendency to subsidence behind the front and the frontal cloud is thus more extensive. In fact the cloud and weather pattern is like a mirror-image of that of the typical warm front. Most of the cloud and precipitation are post-frontal, the warm front sequence in reverse.

Only some cold fronts show a sequence of weather events at all comparable with that of the warm front and even then it is a sequence seen better in retrospect than in advance. The cold front rarely sends high cloud far ahead to announce its coming. The dark seething line of advancing cumulonimbus that accompanies a line-squall is unmistakably menacing. But with many approaching fronts the cloud is hidden by the stratocumulus or stratus characteristic of the warm air ahead. To a pilot the cold front may appear as a wall of cumuliform cloud with attendant problems of bumpiness and aircraft icing. On the ground, falling pressure and backing wind are sometimes indicative of things to come. The precipitation often begins fairly suddenly, and may be short-lived and followed by bright clear skies, or gloomily persistent, according to the type of front. Once the front is through, the wind veers and pressure rises, often sharply. The change in air-mass properties, especially temperature,

humidity, visibility and cloud types, may be very striking to the observer and is often well demonstrated by the traces of continuously-recording instruments (Fig. 45).

Portrait of a Cold Front

A cold front situation is illustrated in Fig. 39 again by chart and cross-section. The front was of the slow-moving variety, accompanied by a substantial, mainly post-frontal, rain belt: it moved southwards over the country, replacing mild Tm weather brought by westerly winds by a cool northerly polar airstream. The wind change is well seen in the section too: the very strong winds at high altitudes above the front catch the eye and will concern us again in Chapter 7. Note how, as the cold air deepens behind the front, the convection clouds typical of this air are able to develop higher until showers become possible. The frontal slope in this example is about 1 in 50.

Waves on a Frontal Surface

We must now consider the processes that disturb the frontal balance, causing these advances of warm air in one region and cold air in another that differentiate the air-mass boundary into warm and cold front respectively. Current ideas on this subject are part of the revolution in meteorological thought created by the Norwegians after the First World War. Long before Bjerknes, in fact as soon as weather maps were being regularly drawn, it was recognised that regions of low pressure—*depressions* or *cyclones*—which could be enclosed by isobars on the chart, tended to be regions of bad weather. The contribution of the Norwegians was to fit a changing pattern of air masses and fronts into the old picture of the depression and to demonstrate that, in most cases, this scheme explained the observed distribution of weather with remarkable success. The Bjerknes model of frontal behaviour, although by no means the last word, remains basic to much of our understanding of the weather of middle latitudes. In this scheme of things warm and cold fronts fall into place as complementary parts of a *frontal depression*.

Frontal disturbances are regarded as *waves* that corrugate the frontal surface much as ripples disturb the surface of water. This analogy is not as remote as it may seem. A water surface can be thought of as a horizontal boundary separating a denser fluid below (water) from a lighter fluid above (air): a frontal surface, not horizontal, it is true, nevertheless also separates a denser fluid below (cold air) from a less dense fluid above (warm air). Waves on a water surface are induced by the wind, *i.e.* by a movement of the air greater than that of the water, so that there is a wind discontinuity or *wind shear* at the separating surface. Similarly a wind

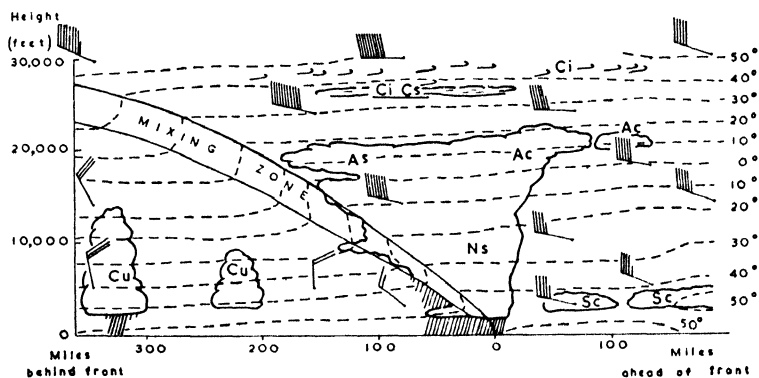
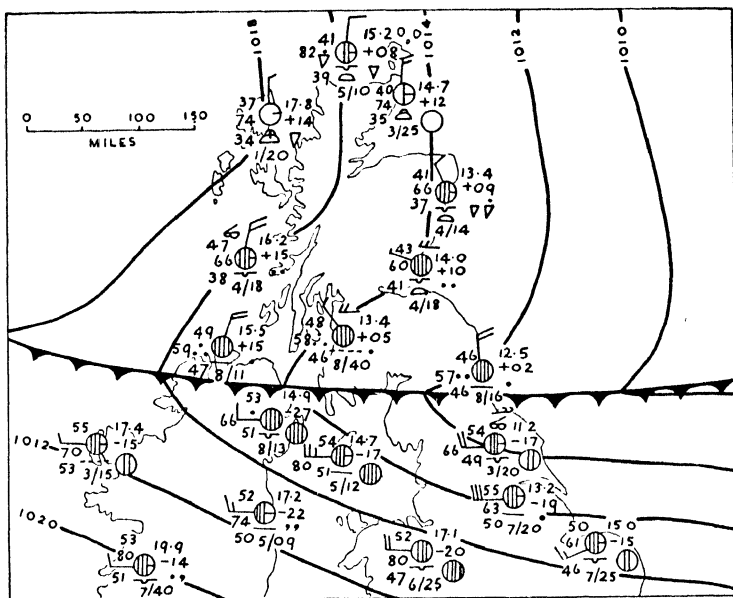


Fig. 39. Portrait of a cold front, 26th October, 1955, 0600 G.M.T. In the cross-section, broken lines are isotherms and wind arrows give strength (see Appendix p. 258) and *compass direction* (assuming north at top of page) at various heights

shear can exist at a frontal surface and, while other forces are also involved, this is the major factor in producing the waves. Since the frontal surface is slightly inclined to the earth's surface, meeting it along a line, these undulations are reflected as wave-like oscillations of the surface front (see Fig. 40). Thus we may legitimately speak of frontal waves on the ground or on the synoptic chart, but must remember that this is merely one view of the three-dimensional picture.

With light winds, water waves are fairly regular, that is to say their amplitude (or height measured from crest to trough) remains roughly the same: when winds strengthen, the wave profile changes, the amplitude increases, the stability of the wave is lost until finally it 'breaks' into foam and spray. Frontal waves behave in a somewhat analogous manner. Some reach a certain stage of development only, maintaining their wave form: these are *stable waves*. *Unstable waves* progress further into a cycle of development in the later stages of which the wave character is lost and replaced by a vortex, a circulation of air with somewhat different weather consequences. This represents the development of a true depression, a process known as *cyclogenesis*.

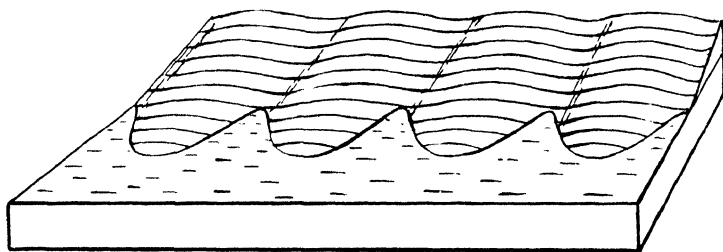


Fig. 40. Waves on a frontal surface

The experience of forecasters, backed by more theoretical justification from the dynamical meteorologists, indicates that the difference between stable and unstable waves is mainly one of size: neither minor ripples nor long shallow undulations in the frontal surface tend to show continued development, but it is the waves of intermediate dimensions that are unstable. However, all frontal waves are carefully watched from chart to chart. We may ourselves follow the life history of a typical frontal depression, 'Low J' (in British synoptic practice, weather systems are named alphabetically) with the aid of some extracts from D.W.R. charts: so common is this development that almost any sequence will do equally well.

The Development of 'Low J'

In Fig. 41, *a*, we see part of the Polar Front out in the western Atlantic. It is shown as a cold front but the movement is very slight: note the indeterminate wind symbols in the cold air mass north of the front. The warm air, in contrast, has quite a strong eastward or north-eastward movement. There is in fact a strong wind shear across the front. Fig. 41, *b*, shows the same scene twelve hours later. The front has now waved slightly so that the surface warm air is pushing into the cold. At the leading edge

of the wave, warm air is advancing, while the cold air is sweeping round to form the back edge: hence the differentiation into warm front J and cold front K. At the tip of the wave, pressure has already fallen and the incipient depression is already labelled 'Low J.'

Twelve hours later (Fig. 41, c) the pattern is well established. The amplitude of the wave has increased, the warm air salient—or *warm sector*—now covers a substantial area. The whole system has progressed eastwards. This is roughly the stage at which the development of a stable wave ceases (there may not even be a closed isobar): the oscillation simply moves along the front until finally it dies out and is no longer traceable on the chart. For 'Low J,' however, this is only a temporary phase.

Fig. 41, d, shows it approaching Ireland. It has now considerably *deepened* (note the central pressure) and the warm sector has reached its maximum amplitude (measuring over 700 miles from north to south). This stage could be described as *mature*. The width of the warm sector is, however, reduced by comparison with the previous chart: this is the beginning of a gradual process of elimination of the surface warm air which is characteristic of the later stages of development.

The reason for this is the tendency for the cold front to advance more rapidly than the warm. The warm air at the warm front is moving not only horizontally but is also ascending the frontal slope, *i.e.* it has an upward component of motion as well as a forward one. In contrast, the cold air progresses only horizontally. As a result, the net forward movement of the warm front is less (by about a third) than that of the cold front, which therefore steadily overtakes it, thus reducing the area of the warm sector and displacing the warm air upwards in increasing quantity. This is the beginning of the process of *occlusion*.

The cold front overtakes the warm front first at the tip of the warm sector and then progressively further away. The cold air mass has now, as it were, caught up with itself (see Fig. 42), thus forcing the warm air up above a thickening continuous layer of cold dense air. But the cold air originally ahead of the warm front and that behind the cold front are seldom of similar temperatures. Although belonging to the same air mass, these two streams of cold air have followed different paths and have therefore undergone modification in different ways. The temperature contrast between them manifests itself in the formation of a new front—an *occluded front* (or, simply, *occlusion*)—which replaces warm and cold front at the surface.

The cold air sweeping into the rear of the depression from a northerly or north-westerly direction is generally fresher and colder than that ahead of the warm front, and consequently undercuts it just as it had previously undercut the warm air: the resulting surface front is then known as a

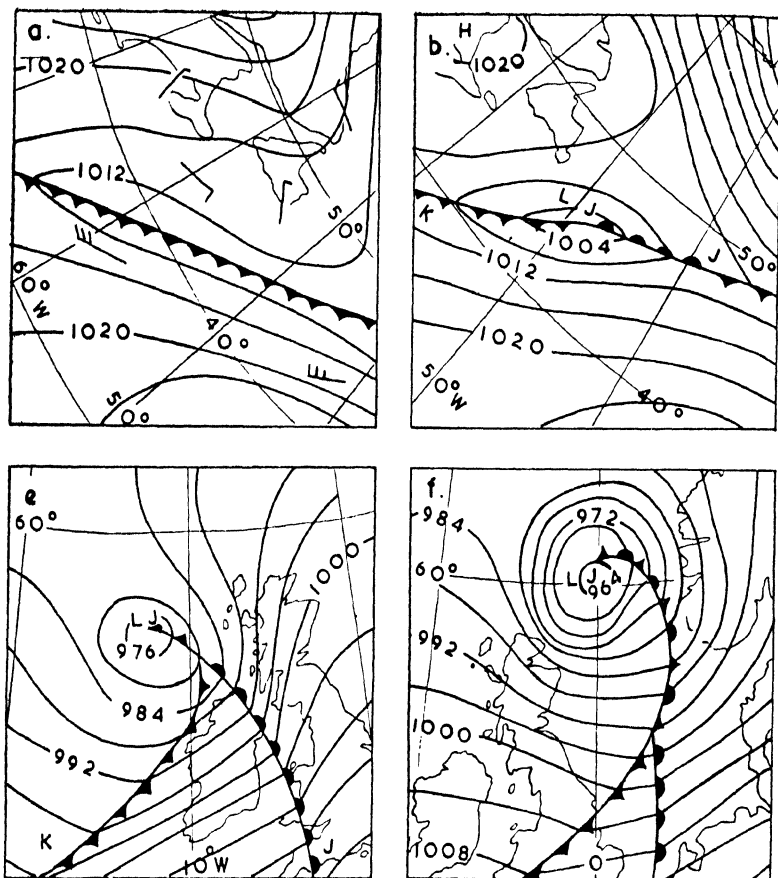
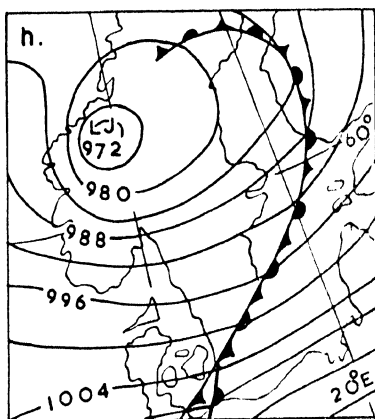
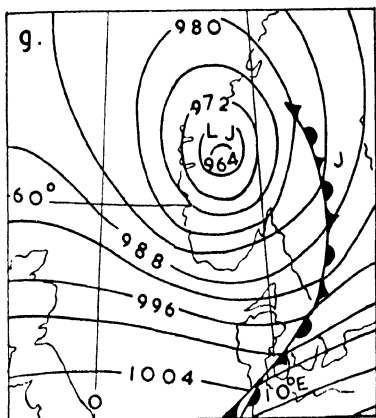
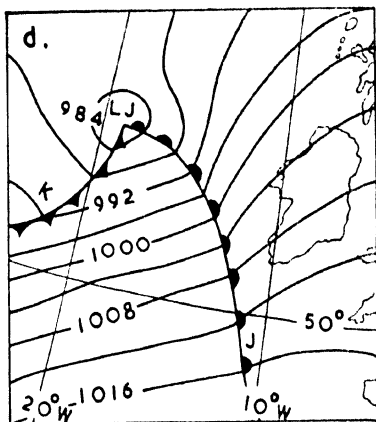
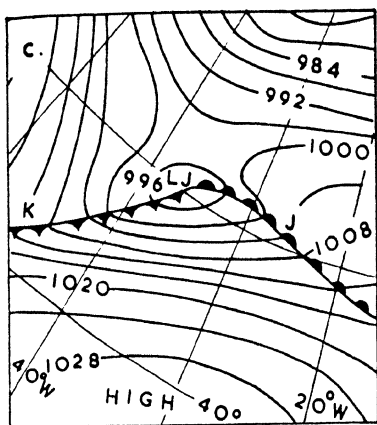


Fig. 41. The history of Low 'J,' (a) 9th November, 1954, 1200 G.M.T., (b) 10th November, 11th November, 0600 G.M.T., (f) 11th November, 1800 G.M.T.,

cold occlusion (Fig. 42, c). Less commonly the air behind the cold front comes from a south-westerly quarter (Polar Maritime Returning) and is then the warmer of the two cold air streams: on occlusion, this warmer air slides up over the colder in warm front fashion (Fig. 42, d): this is the *warm occlusion*.

Returning now to the history of 'Low J,' we see the depression in Fig. 41, e, just after occlusion has begun, and the remaining charts show later stages of the progressive elimination of the warm sector. Occlusion J is mainly of the cold type. As would be expected, a cold occlusion continues the line of the cold front, a warm occlusion that of the warm front. Meanwhile the depression continues to deepen and, now over Norway, reaches its lowest central pressure (964 mb) in Fig. 41, g, when occlusion is not



0000 G.M.T., (c) 10th November, 1200 G.M.T., (d) 11th November, 0000 G.M.T., (e) (g) 12th November, 0000 G.M.T., (h) 12th November, 0600 G.M.T.

far from complete. We may appreciate that while a substantial warm sector remains, energy must be applied to the task of lifting it by occlusion: but once this operation is largely effected, more energy becomes available and this shows itself in increased intensity of circulation and stronger winds. Eventually, however, the energy is spent, the ageing system no longer deepens but instead begins to 'fill' (Fig. 41, *h*). As this happens, the central pressure rises and fewer and fewer closed isobars can be drawn on successive charts. In these dying stages, the occluded front becomes unrecognisable near the centre of the depression where the air is by this time so whirled about as to be virtually homogeneous. Ultimately, all traces of 'Low J' will vanish from the chart.

The sequence of events described above has taken nearly three days

during which time 'Low J' has moved something like 3,000 miles. From the initial waving on the Polar Front to the beginning of occlusion has taken just over a day. This is a fairly typical time schedule. Certainly, a depression that begins life in the western Atlantic, like 'Low J,' is likely to be more or less occluded by the time it reaches north-west Europe: mature depressions are comparatively rare over this country although they do occur. However, not all the synoptic patterns we observe can be explained in terms of the above history: there are further complexities which we shall consider shortly, but it will be convenient first to describe the weather typically associated with the simpler systems.

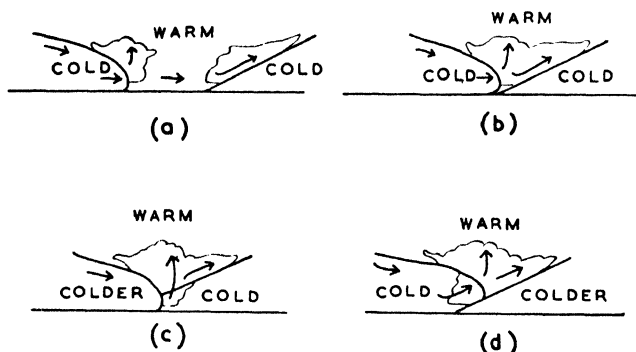


Fig. 42. The process of occlusion. (a) Well before occlusion. (b) Just before occlusion. (c) Cold occlusion. (d) Warm occlusion

Depression Weather

In the broadest terms, depression weather is a combination of the weather of the separate fronts and air masses involved in the system. In detail, however, much depends on the stage of development reached by the depression and, as far as the individual observer is concerned, on his position within it and relative to its path of travel. Before occlusion, the depression consists of two belts of frontal cloud and rain merging near the centre of the system into a general cloudy and rainy area. The warm sector, giving weather typical of the warm air mass (Tm or Pmr usually), is wide in the case of a stable wave or the early stages of a developing depression. As a result of occlusion this part of the weather sequence is removed and the two frontal belts coalesce into one.

Fig. 43 shows a mature depression centred over North Wales: its path of travel is north-eastwards. An observer stationed in East Anglia will experience the weather sequence in its entirety—first, typical warm front weather, then the air-mass weather of the warm sector, followed by cold front weather and finally the weather of the cold air mass in the rear. An

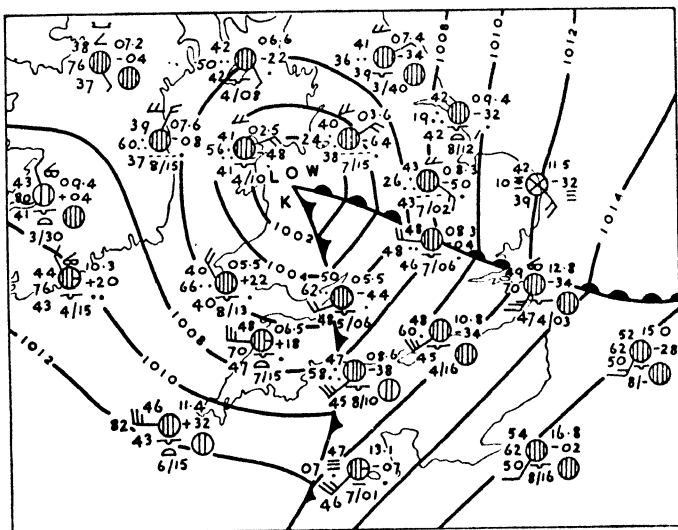


Fig. 43. Mature frontal depression, 8th April, 1947, 0600 G.M.T.

observer in Yorkshire will experience only the general rain area and will not be conscious of the separate facets of the system. At the other extreme, the fronts weaken well south of the centre and an observer in Belgium may report only frontal cloud and an air-mass change, with little or no rain.

Our second example is of a partially occluded depression over western Scotland (Fig. 44) moving rapidly eastwards. This is a deep system (the central pressure is less than 968 mb) with a correspondingly vigorous circulation: surface winds are everywhere strong and reach gale force in the west of the region. Clearly, only observers in northern England and part of southern Scotland will see anything of the warm sector. Further north, occlusion weather will dominate. This is rather variable from case to case but, in general, an occlusion being in origin a composite front, combines warm front and cold front weather though in somewhat telescoped form (Fig. 42). Rain continues to fall from the cloud in the lifted warm air (from what is sometimes called the *upper front*) in addition to that resulting from the frontal lifting of one cold air mass by the other (the surface front).

For this reason occlusion weather has a reputation as the worst frontal weather we experience: a slow moving occlusion may indeed provide many hours of continuous rain and gloom. After a few days, however, continued uplift has so denuded the warm air of its moisture content that it contributes little further to the frontal weather: a cold occlusion then becomes effectively a cold front and a warm occlusion a warm front.

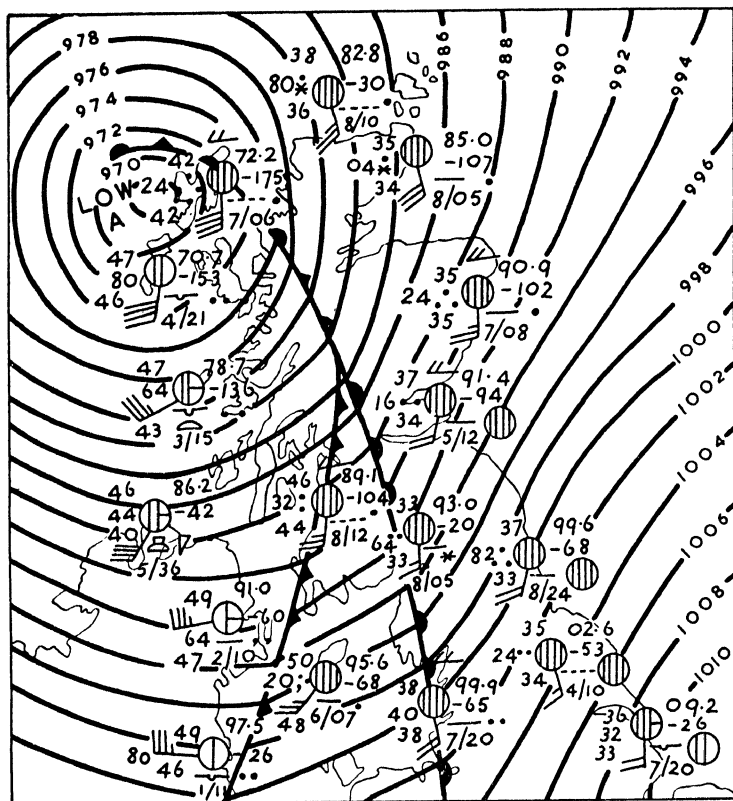


Fig. 44. Partially occluded depression, 30th December, 1951, 0000 G.M.T.

In Fig. 44, incidentally, the occlusion is a warm one. In the final stages of a depression, the cold air is everywhere present in considerable depth and the fronts are no longer significant: the resulting weather is then typical of the cold air mass, that is to say, cold and showery.

The passage of a frontal depression similar to that of Fig. 43 is strikingly shown by continuous traces in Fig. 45. The rapid air-mass changes completely upset the normal diurnal temperature variation. As sometimes occurs with winter depressions—and very illustrative of the temperature changes involved—the warm front precipitation began as sleet which later turned to rain and the cold front first yielded rain which soon turned to snow.

Complex Frontal Patterns

It is common for the sequence of events outlined in the history of 'Low J' to repeat itself again and again along the Polar Front so that, on a

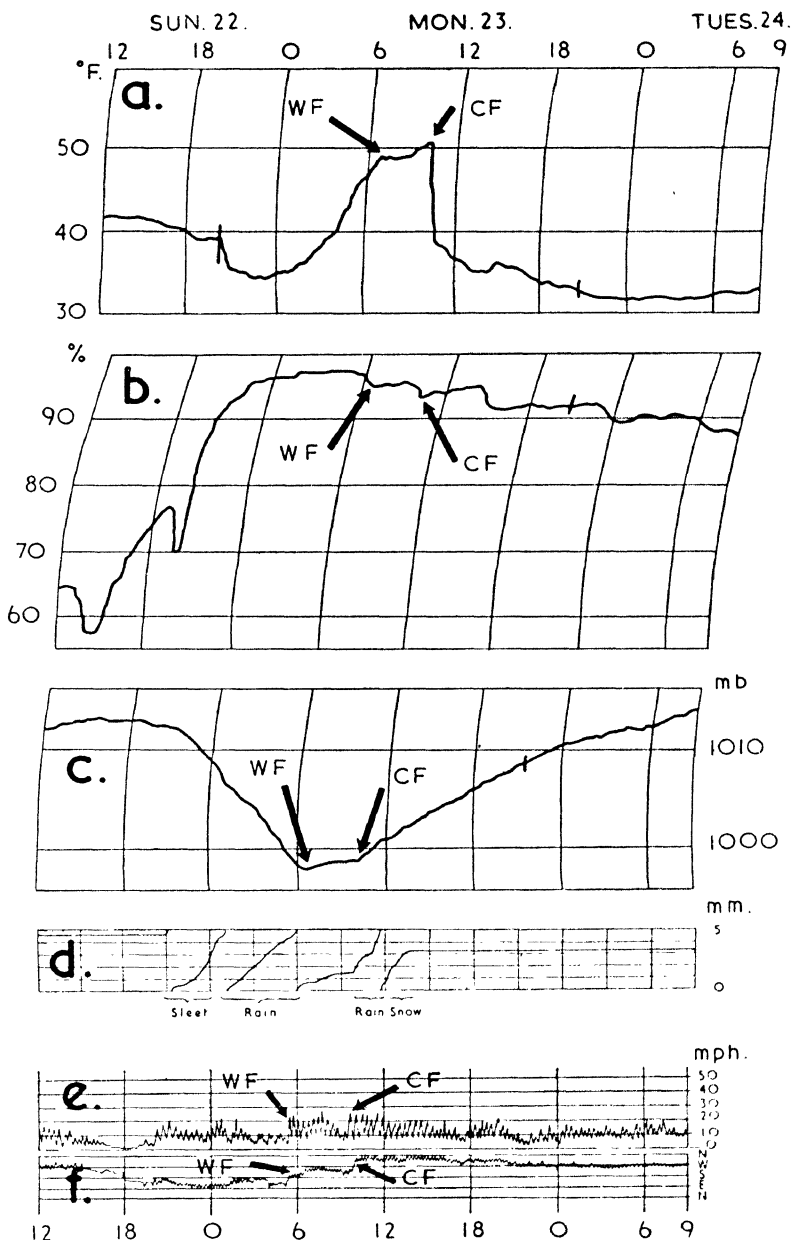


Fig. 45. Weather of a frontal depression, illustrated by continuous traces of (a) temperature, (b) relative humidity, (c) pressure, (d) rain-fall, (e) wind force and (f) direction. Abingdon, Berks, 22nd-24th January, 1956. The arrows marked W.F. and C.F. indicate the passage of the warm front and cold front, respectively

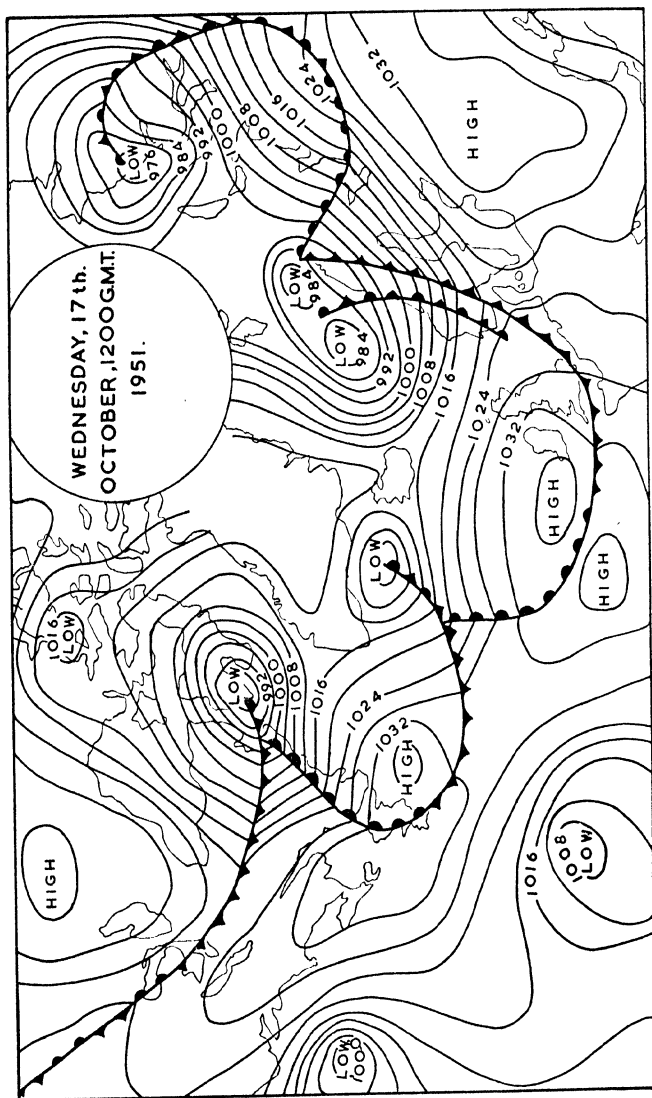


Fig. 46. A depression family

given synoptic chart (e.g. Fig. 46) there may be a string of frontal depressions in middle latitudes, often in different stages of development. Such a series, which usually comprises four or five but occasionally up to seven members, is known as a *depression family*. Each depression follows roughly the same path, usually south-west to north-east, so that a strategically situated region—such as the British Isles with regard to Atlantic depressions—may be visited by each member in turn of the entire family. This is the explanation of our too-familiar spells of unsettled weather that are especially common in late autumn and winter: periods of cloud and rain and often strong winds—typical depression weather—alternate with brighter interludes brought by the intervening areas of higher pressure. The entire changeable spell may take five or six days to run its course.

The sequence does not continue indefinitely. The process of repeated waving tends to push the Polar Front further and further south of its usual position until eventually it trespasses into an unfavourable meteorological regime, that of the Sub-Tropical High Pressure, where it quietly frontolyses. We return to this significant development in Chapter 7.

Another common occurrence is the appearance of a so-called *secondary depression* within the circulation of the main (or *primary*) depression. Beginning as a minor low pressure area, the secondary may deepen until it is comparable with the primary, thus giving rise to a complex depression with two centres which tend to circle each other slowly. Systems with three centres are also sometimes seen. A secondary often develops as a result of renewed waving on the trailing cold front at the rear of a primary depression (Fig. 47, *a*): this sometimes betrays itself on the synoptic chart by the slowing down of the cold front and therefore of the occlusion process. On some occasions the secondary remains as a wave travelling up the cold front, on others it develops into a fully-fledged depression. This may mark the beginning of a depression family as described above.

Also a favoured area for the formation of a secondary is the tip of the residual warm sector of a partially occluded primary depression (Fig. 47, *b*). Sometimes such a secondary develops to become the major centre of the system, deepening at the expense of the primary which may degenerate into a mere *trough* of low pressure marked only by an outward tonguing of the isobars. Under these circumstances the trough and the northern end of the occlusion may be carried round by strong winds in this part of the system to form what is called a *back-bent occlusion*, (Fig. 47, *c*). This development adds a third front and its associated bad weather to the sequence experienced by an observer in the path of such a system. In still other cases, the secondary forming at the tip of the warm sector moves rapidly eastwards away from the primary and is then described as a *breakaway depression*.

It is common to find three fronts associated with a depression for quite

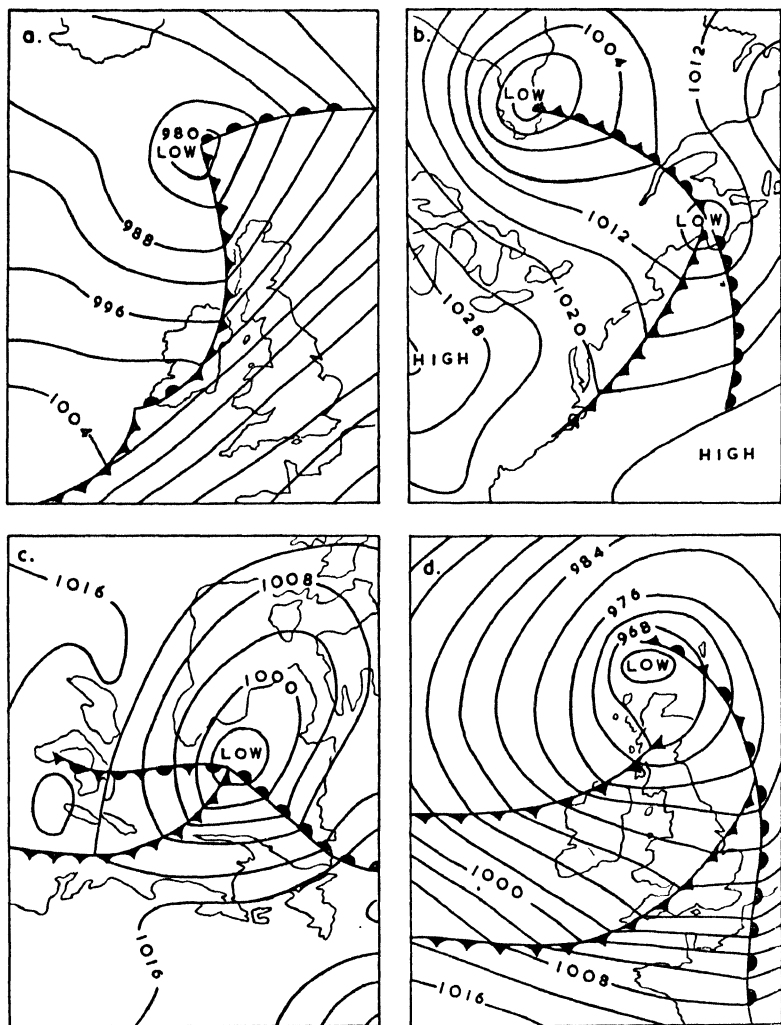


Fig. 47. Complex frontal patterns. (a) Wave on the cold front of a depression, 22nd December, 1946, 1200 G.M.T. (b) Secondary depression at the tip of the warm sector of a partially occluded primary, 29th November, 1955, 1200 G.M.T. (c) Back-bent occlusion, 21st June, 1951, 1200 G.M.T. (d) Secondary cold front, 16th December, 1952, 1800 G.M.T.

another reason. The main cold front is often followed anything from 50 to a few hundred miles behind by a *secondary cold front* which is often the more active of the two (Fig. 47, d). There are probably several reasons for the formation of such a secondary front. Sometimes it results from frontogenesis between fresher, colder Polar air coming in behind more modified air of similar origin in the rear of the depression. Often it clearly

represents a portion of the Arctic Front drawn into the circulation of the depression. In either case, the arrival of really cold air is not to be expected until the secondary cold front has passed through. Lows with complete double frontal patterns are not uncommon.

These few examples by no means exhaust the possibilities but may help close the gap between the relatively simple 'text book' patterns and the complexities that can be appreciated only through regular acquaintance with the synoptic chart.

Some Problems of the Frontal Depression

The genesis of most middle-latitude depressions as wave-like disturbances on the Polar (or another) Front and their subsequent development as outlined in this chapter are facts that have been established and confirmed time and time again on the synoptic chart. While theory and observation march together a considerable way, we are far from knowing the full explanation of the facts. Why the initial disturbance that sets in train such a sequence of events occurs in a given place and at a given time is a question to which we have as yet no simple and direct answer.

It is a matter of observation that some waves are set off by other disturbances nearby—for example, an existing Polar Front depression may induce a 'sympathetic' wave on an adjacent part of the Arctic Front—but this merely shifts the problem without solving it. A variety of geographical factors deserve consideration. It has been suggested, for example, that, if a slowly advancing cold front moves against a physical obstruction—such as an island interrupting a smooth expanse of ocean, or the tip of a land mass—the flow of cold air may be impeded, allowing the warm air to encroach within it and thus initiate the wave. At coastlines or mountain ranges sudden changes in the frictional resistance to the air flow may likewise be disturbing factors. Many meteorologists see the developments depicted on surface synoptic charts as mere reflection of events occurring at higher levels. These we have yet to consider.

Then there is the problem of the fall of pressure which is an integral part of cyclogenesis. So far we have concentrated on the interplay of air masses and fronts but it is now time to superimpose on these two patterns that of the associated pressure change. Nor must it be thought that all depressions are frontal. These problems and others concerned with atmosphere circulation will engage our attention in the next chapter.

ATMOSPHERIC CIRCULATION

Circulation is a fundamental feature of our atmosphere. On the smallest scale we see it in the eddies that swirl dust and leaves about on the ground. On a vast scale that swamps our comprehension, it dictates the travel of air masses and fronts and thus lies behind the day-to-day fluctuations of our weather. Atmospheric circulation is always a response to inequalities of pressure. The distribution of pressure as shown by the pattern of isobars (built up from the plotted pressure data) forms the framework of any weather map. Pressure patterns and horizontal air flow (winds) are intimately related: so are horizontal and vertical circulations. Pressure patterns are thus important not only in their rôle of steering air masses and fronts but also contribute to the weather in their own right.

Pressure and Winds

The weight of the atmosphere is by no means insignificant: at sea-level it amounts to nearly 15 lb. pressing down on every inch of us, but this is, of course, equalised by the outward pressure of air within our bodies. Except for occasional temporary deafness when pressure changes abruptly (in a London tube train or when descending a mountain by funicular) we are oblivious of this weather element. Sea-level pressures may rise or fall by 50 mb in a day. Such barometric changes are too small to be felt but their significance regarding the weather is considerable.

The wind, on the other hand, is a common part of our everyday weather experience. It influences our personal comfort out of doors and our sufferings at sea; it dries the washing, sails our yachts and is a perennial source of power or, in excess, of disaster (see Chapter 11). The relation between wind and pressure is our first concern in this chapter.

Balancing Forces in the Atmosphere

By analogy with the gradient of sloping ground we use the term *pressure gradient* to indicate the change of pressure from place to place (strictly, at right angles to the isobars). Gradients may be 'steep,' in which case the isobars are closely packed (see Fig. 44), or 'slack,' with widely

spaced isobars (Fig. 33). Where there is a pressure gradient, we should expect air to flow from high pressure (more air) to low pressure (less air) until the inequality disappears and that the flow should be stronger the steeper the gradient. The problem can be illustrated by trying to maintain the water level in one limb of a U-tube higher than in the other. But if we look at any weather map we find that while the speed of the surface winds indeed depends on the closeness of the isobars, their direction is not down the pressure gradient, *i.e.* directly across the isobars, but more or less along them. Pressure differences are the primary cause of wind but clearly other forces are also involved.

One of these is a consequence of the rotation of our planet from west to east about its axis once every twenty-four hours. A point on the Equator is constantly moving eastwards at something over 1,000 m.p.h. Nearer home, London, on a shorter ring of latitude, executes its daily rotation at only about 650 m.p.h. The North Pole, on the earth's axis, has no linear movement but simply rotates on the spot. The atmosphere above the earth at various latitudes is being whirled round with it at comparable but not necessarily identical speeds. If the movement of the air is exactly the same as that of the underlying surface, we speak of 'calm.' Only when the air moves with a speed or direction different from that of the surface are we conscious of 'wind.'

Let us now imagine calm conditions at the Equator, which is to say that both air and surface are moving eastwards at about 1,050 m.p.h. Suppose we suddenly shift some of this air (still preserving its eastward speed of 1,050 m.p.h.) northwards to, say, latitude 10° N., where the earth's surface is moving at only 1,035 m.p.h. An observer at 10° N. will in all honesty record a west wind of 15 m.p.h. (actually reduced somewhat by friction). Now air cannot be transported suddenly in this way but it is nevertheless a fact that air moving northwards (south wind) *seems* to acquire a westerly (eastwards) movement because of its advantage in speed over the surface below. Similarly, air travelling south from the Pole acquires a westward movement because it is entering faster-moving latitudes and is therefore left behind in relation to the lines of longitude on a rotating earth.

This argument cannot be applied when air moves along a line of latitude. But air travelling eastwards moves faster than the earth below it. This air develops an excess of *centrifugal force* (the force that tends to throw a rotating body outwards from the centre of rotation). It cannot be thrown out into space because of the pull of gravity but it is forced out to a latitude having a speed of movement equal to its own, *i.e.* equatorwards. Conversely, air travelling westwards—more slowly than the earth below—has a deficiency of centrifugal force and moves towards a latitude of similar speed, *i.e.* northwards.

In fact, whether air moves north or south, east or west, or in any other

direction, it is always urged to the *right* of its original direction in the Northern Hemisphere (to the left in the Southern). The force responsible, which is stronger the higher the latitude and the greater the wind velocity, is sometimes called the *Coriolis Force* (after the French physicist who first studied it) or simply the *deflecting force* due to earth rotation.

Air thus moves horizontally in obedience to the push of a pressure gradient but, once moving, is subject to this deflecting force. If an air stream pursues a steady straight course, it must be because these two forces exactly balance each other. We can suppose that air begins to move down the pressure gradient but is gradually deflected more and more to the right. The limit to this deflection is reached when the air flows at right angles to the gradient, for any further deflection would send it moving up the gradient, which is not possible. This balanced state is illustrated in

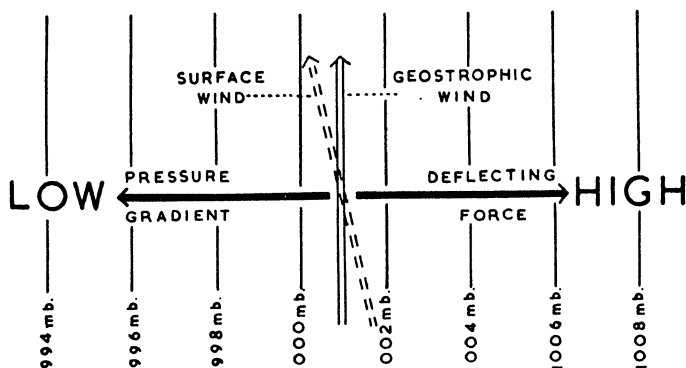


Fig. 48. Balancing forces in the atmosphere

Fig. 48: in order to counter the deflecting force, the push of the pressure gradient must be towards the left. This relationship is summed up in an old-established meteorological rule, first stated about a hundred years ago by the Dutch Professor Buys Ballot: if we stand with our back to the wind the lower pressure is always to our left (to our right in the Southern Hemisphere).

The wind adjusted to this balance of force is known as the *geostrophic wind* and flows along the isobars at a speed proportional to their spacing. Not all winds are geostrophic. When the pressure pattern changes, the balance is upset, but the deflecting force always strives to readjust itself to the new pressure gradient and so eventually to restore the balance.

When air follows a curved path it is subjected to another compulsion, centrifugal force, tending to throw it outward from the centre of curvature. Three forces are now in balance and the wind so maintained is called the *gradient wind*. Like the geostrophic wind, it blows along the isobars

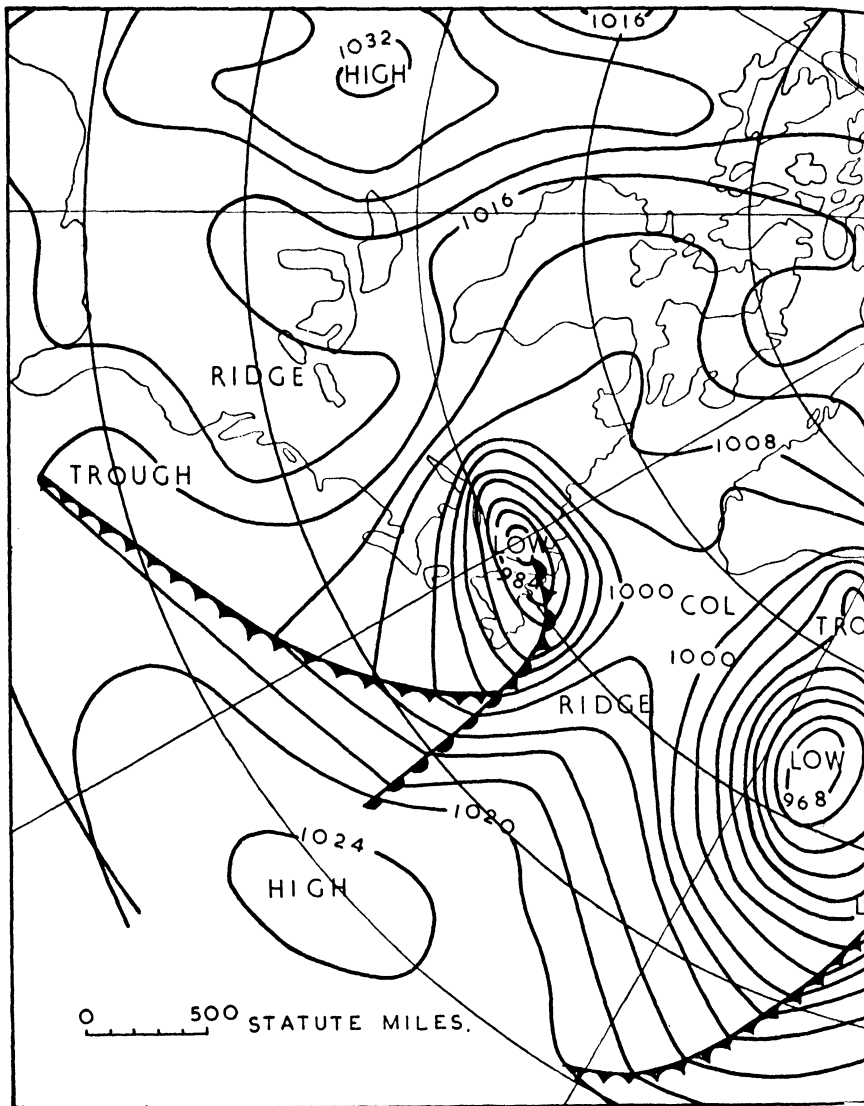


Fig. 49. A Northern Hemisphere synoptic chart

(curved in this case) and is equally subject to Buys Ballot's Law. Centrifugal force becomes important only with pronounced curvature of flow.

Surface Winds

Near the ground, moving air is slowed up by friction against more or

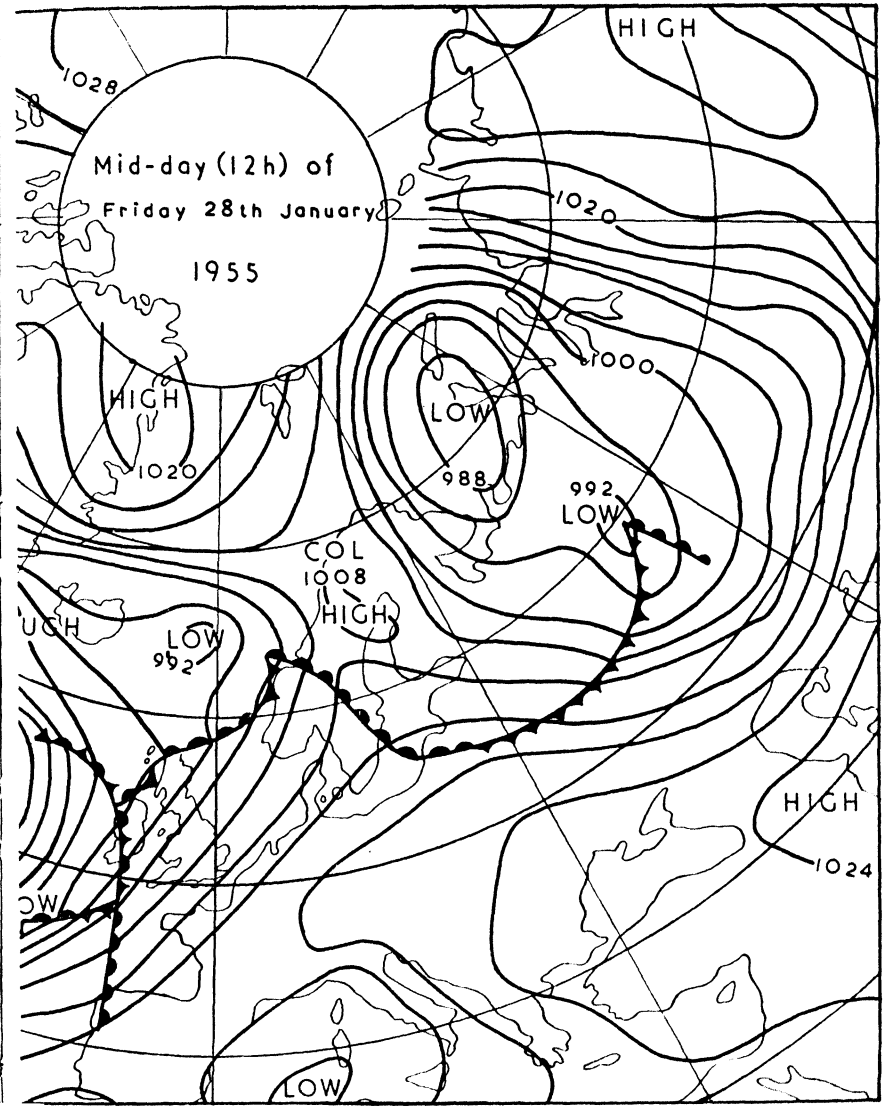


Fig. 49. A Northern Hemisphere synoptic chart

less irregular surfaces. This upsets the balance of forces, for the deflecting force (and the centrifugal force) depend on the wind velocity. If this is weakened the deflecting force is weakened and the pressure gradient asserts its greater strength by causing the air to flow somewhat towards the lower pressure (Fig. 48). All but very light surface winds in fact blow

at a small angle to the isobars from high to low pressure. Reference to any weather map will confirm this.

In general, the effect of surface friction is felt in the lowest 2,000 feet or so of atmosphere—the so-called *friction layer*. At about 2,000 feet the wind is adjusted to the isobars (geostrophic or gradient wind). The surface wind differs from the undisturbed wind in both speed and direction, to an extent depending on the roughness of the surface. This is least over the sea, where the surface wind may have three-quarters the speed of the gradient wind and deviate from it by only 10° . Inland, during the day, the surface wind is nearer half the gradient wind and the angle between is 20° to 30° . At night, the wind generally drops as turbulence dies down, no longer transporting the faster velocities of higher levels down near the ground: the surface wind may then be only a fifth of the gradient wind and the angle may increase to 40° or 50° .

Common Pressure Patterns

With these principles in mind, we may briefly review the common pressure patterns, which are all illustrated in Fig. 49. We find extensive areas of straight isobars. Elsewhere are regions of low pressure, some with well defined centres enclosed by isobars (depressions or 'lows'), some mere extensions or *troughs* of lower pressure. The high pressure centres (anti-cyclones or 'highs') bulge out in *ridges* or *wedges*. The least determinate pattern is the *col*: this is a kind of no man's land between two highs and two lows, a region of stagnant air. In all cases we may remember the direction of circulation by applying Buys Ballot's Law in reverse: this tells us that the circulation is anti-clockwise around lows and clockwise round highs (reverse these for the Southern Hemisphere), always with the surface winds blowing slightly towards the low pressure.

Upper Winds and the Jet Stream

The high level winds often differ markedly from the surface and even the 2,000 feet wind. We may appreciate why with the help of Fig. 50. This shows two air columns, A and B, each containing the same weight of air (so that the surface pressure at *a* equals that at *b*) but column A is warmer throughout than column B. The air in column A is consequently less dense than in column B (which must then be the shorter of the two). At some upper level, say *a'b'*, there must be a greater weight of air above *a'* than above *b'*. In other words, there is a pressure gradient at the level *a'b'* where none existed at the surface, because pressure decreases more slowly with height in warm air than in cold. At the upper level a wind flows in obedience to this pressure gradient, initially from *a'* to *b'* but eventually adjusted by earth rotation to a wind blowing into the

paper. We call such an upper wind a *thermal wind* because it results from temperature differences in the atmosphere.

The upper air flow is largely dominated by the distribution of warm and cold air. As we go upwards, there is a gradual change from the winds adjusted to the sea-level isobars to those obeying the thermal pattern. Hence its important rôle in modern forecasting. The thermal winds are represented on charts of *thickness lines* (lines of equal thickness) which are in effect isotherms. The vertical thickness of an air layer between two pressure levels (e.g. 1,000 and 500 mb) is a measure of its mean temperature, for the

warmer the air the more it expands and the greater its thickness. Actual high level pressures and winds (derived from radio-sonde data) are represented by isobaric charts for various levels or, in British practice, by *contour charts* which show the varying height of a fixed pressure surface (e.g. 500 mb) just as relief contours show the varying height of the land. Contour charts show varied patterns of upper level highs (regions of warm air, like column A in Fig. 50) and lows (cold air like column B). Contour lines provide a measure of the direction and speed of upper winds and these charts are essential tools for estimating and forecasting such winds for high-level aviation.

Before the last war there had come reports from time to time of high-flying aircraft being either slowed up or greatly speeded up or diverted far off course by unexpected and unpredicted winds of great velocity. Since the regular construction of high-level charts, we know that narrow belts of wind speed exceeding 100 and sometimes 200 m.p.h. are a normal feature of the upper troposphere. These have been christened *jet streams*. They are obviously very intense thermal winds, being always associated with sharp horizontal temperature changes. On their margins wind speed changes rapidly over short distances and turbulence may be severe. Jet streams are of the utmost significance to modern aviation—headwind jets

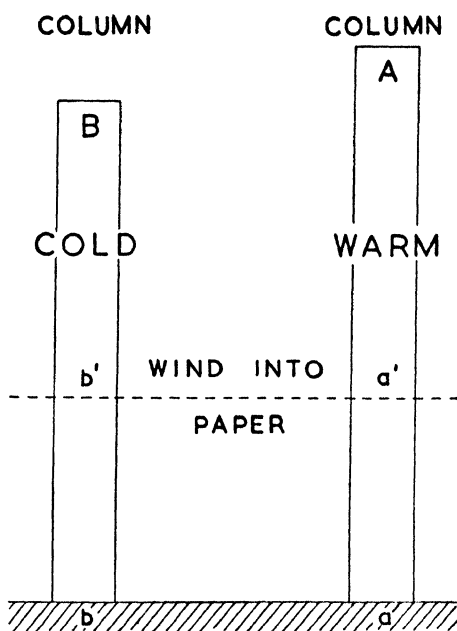


Fig. 50. Illustrating the principle of thermal winds

must be avoided but tailwind jets are a boon—and the successful forecasting of the position and movement of these invisible currents becomes a matter of importance.

Pressure and Winds at Fronts

We can now see that at a stationary front (Chapter 6) the cold air is prevented from completely under-running the warm because a condition of geostrophic balance is achieved. There is, in fact, a pressure gradient from cold (heavy) air to warm (light) air but the resulting wind is slewed round to the right by earth rotation until it blows parallel to the front rather than across it. The isobars appropriate to this balanced condition must also lie parallel to the front, which is how we recognise a stationary front on a synoptic chart.

A moving front must be crossed by isobars for these represent the push that carries it along. As moving fronts lie along the axes of more or less pronounced troughs of low pressure, the isobars must change direction from one side of the front to the other and are always drawn more or less sharply kinked across it (*e.g.* Figs. 38 and 39). As a warm front approaches, *i.e.* as warm air steadily replaces cold, pressure must fall: after the passage of a cold front, pressure rises as the cold dense air deepens. The wind-changes at a front (Chapter 6) follow in part from the changed direction of the isobars.

With well-marked fronts the temperature contrast across the transition zone is found at all levels from near the ground to the very top of the troposphere. Because of this a strong thermal wind must exist, blowing parallel to the frontal surface with (see Fig. 50) the cold air to the left. The *frontal thermal* reaches its maximum strength just below the tropopause and this explains the location of the jet streams of middle latitudes. The position of the typical frontal jet is shown diagrammatically in Fig. 51. They lie broadly parallel with the surface fronts, move generally with them and, like them, often wave and become distorted. The *Polar Front jet* is a common feature of middle-latitude upper-air charts. Not all jet streams however are frontal. The most persistent are found far from fronts in sub-Tropical latitudes and their rôle in the atmospheric machine is by no means clear.

The Mechanism of Lows and Highs

The mechanisms of the various types of low- and high-pressure systems are not always easily understood in detail but basically all are variations on relatively simple themes. Consider first the formation of a low in a region of previously uniform pressure. The key to the mechanism is the

removal of air by one means or another. This reduction of the total weight of atmosphere over part of the region sets up a surface pressure gradient in response to which air must move in from all sides. But air cannot pile up indefinitely and some must be pushed upwards. Such an accumulation of air which must be relieved by vertical currents is known as *convergence* (a meteorological term implying rather more than the everyday use of the word). The same process operates at an active front (Chapter 6).

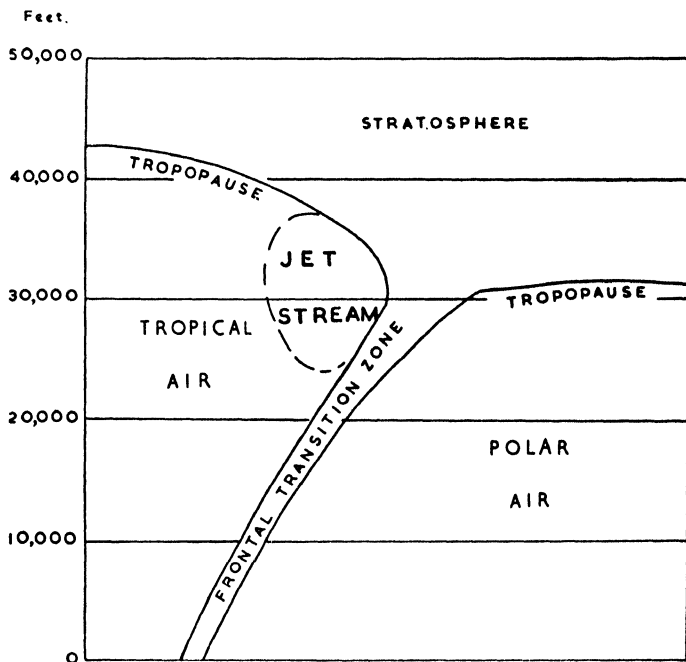


Fig. 51. Diagrammatic vertical section across a frontal jet stream. The direction of the jet is at right-angles to the paper

Of itself, this influx of air would quickly even out the pressure deficiency. If the low is to be maintained, air must be evacuated somewhere at higher levels and at a rate exceeding that of the surface inflow. Outflowing winds at height, fed by the rising air (*divergence*), are a necessary part of the system.

At first the flow of air is in direct obedience to the pressure gradient (that is, non-geostrophic) but eventually the deflecting force due to earth rotation asserts itself. The inflowing low-level winds are deflected to the right to give a typical cyclonic circulation, while the outflowing upper winds are similarly deflected to give an anticyclonic circulation. But while the low is active, there must be sufficient flow across the isobars—inwards

at low levels, outwards at height—to maintain the mechanism. At a frontal convergence zone, the tendency of the surface winds to flow towards the lowest pressure explains the backing wind ahead of the approaching front. In the dying stages of a depression inflow dominates and the low ‘fills.’

In the case of a high-pressure system it is easy to see that the same mechanism must operate but in reverse. Pressure rises over an area and there is a low-level outflow in deference to the newly-created pressure gradient. This leaves a deficit of air which is made up by descent or *subsidence* from above. To maintain the high, this low-level divergence must be more than compensated by air drawn in as a result of high-level convergence. Earth rotation again obtrudes to induce an anticyclonic circulation at low levels and a cyclonic one at height. Even when pressure is no longer rising, the ‘frictional leak’ of air near the surface keeps the

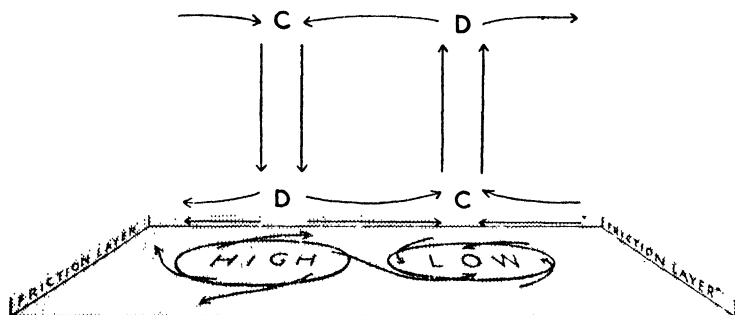


Fig. 52. Horizontal and vertical circulations in adjacent high and low systems.
C, convergence. D, divergence

mechanism going. When high-level air no longer comes in to feed the down-currents, the high dissipates.

The relations between these horizontal and vertical circulations in an adjacent high and low are illustrated in Fig. 52. It will be clear that, in these developing systems, enormous weights of air are transported to and fro at various levels but what our surface barometers record are only the *net* changes which are small in comparison. There remains the all-important question of what causes the pressure changes that set these developments in train. We are far from knowing all the answers but it seems that such changes may result from the heating or cooling of the atmosphere near the ground, or from the horizontal transport (advection) of warm or cold air, or from purely orographic causes.

Low-Pressure Systems

However they originate, all low-pressure systems share in common the fact of ascending air, whether this is localised in definite belts as with

frontal depressions or general as with other types. To this characteristic they owe their habit of yielding cloud and precipitation. In the following pages we deal briefly with the various types from the point of view of their modes of origin.

Thermal Depressions

Thermal depressions are caused by the over-heating of parts of the surface. This occurs, for example, over islands and peninsulas in summer or over lakes and inland seas in winter, owing to the different properties of land and water (Chapter 3). An island is warmer than the surrounding sea in summer. The air above the land becomes warmed and expands vertically: this produces high pressure aloft as air accumulates from below. In response to this upper-level pressure gradient, air flows outwards thus reducing the total weight of atmosphere above the island and increasing it above a belt of surrounding sea. By this removal of air the mechanism is set going and descending air above the sea and inflowing winds at low levels complete the circulation.

This illustrates the important meteorological principle of the *thermal circulation*. Essentially the same system is found in any cumulus cloud (though here the scale is too small for earth rotation to be effective). The same mechanism operates on a continental scale to give what is known as a *monsoon low*, which dominates the summer weather and climate over vast areas. It is also, as we shall see, the mainspring of the global circulation.

The commonest thermal depressions we experience are the shallow *Polar Air Lows* that arrive in winter embedded in a Pm airstream. In these it seems to be the latent heat released during condensation in frequent showers that sets the thermal mechanism in motion. (This latent heat must provide part of the energy supply in any depression.) The shallow thundery lows of summer which reach us from the south are of similar character. Most thermal lows give heavy showers or thunderstorms rather than continuous bad weather.

Lee Depressions

Of quite different origin is the *orographic* or *lee depression*. This forms when an airstream moves directly against a high mountain barrier. Some air skirts the mountains, some flows over the crests, but there is bound to be a considerable piling up of air on the windward side and a relative deficit on the leeward side. This 'partial vacuum' is often sufficiently marked to show itself on the chart as a shallow low or trough, while a ridge of high pressure appears on the other side. This type of low, while primarily non-frontal, may draw nearby fronts into its circulation. Oro-

graphic depressions are often found east of the Rocky Mountains and over the Italian plain in the lee of the Alps.

The Origin of Frontal Depressions

It seems a far cry from the simple mechanism of the thermal low to the complexities of the frontal depression (Chapter 6). The last word has not been said on this subject but the frontal low is really a convective cell of a kind, with warm air rising at the fronts (especially the warm front) and cold air descending in the rear of the system. The fall of pressure must be related to the encroachment of warm air into cold at the waving front which is, in a sense, equivalent to direct heating from below.

Part of the difficulty lies in the close mutual adjustment of pressure and winds, which makes cause and effect hard to disentangle. A fall of pressure eventually produces a cyclonic circulation: equally, from the law relating pressure and wind, a distortion of air flow giving a cyclonic

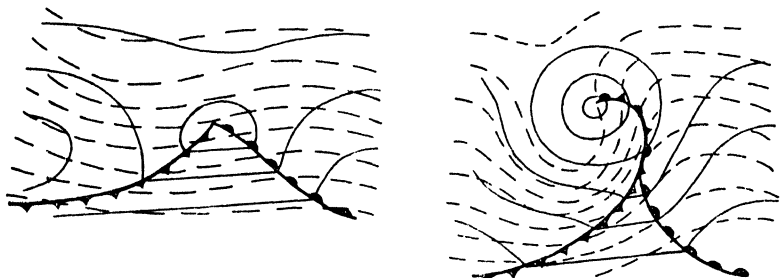


Fig. 53. Upper air flow in a frontal depression at two stages of development. Surface flow shown by continuous lines, upper flow by broken lines

circulation must result in a fall of pressure. It is the meteorological version of the problem of the chicken and the egg.

Another open question is whether the depression originates at low or high levels. A well-developed frontal low is always linked with a disturbance of the upper-air flow, as indicated in Fig. 53. This shows an upper ridge ahead of the surface warm front (large-scale ascent of warm air) and an upper trough following the surface cold front (cold subsiding air). Which is the fundamental pattern? The Norwegians originally placed the emphasis on surface origins but today it is thought that, in many cases, the surface pattern merely reflects the upper-air flow. The upper trough induces a fall of pressure at the surface, the low-level winds respond to this and the mechanism is set off. The problem is still far from solution.

Tropical Hurricanes

Of quite a different order are the intense tropical low-pressure systems, which are variously known as *hurricanes* (West Indies), *cyclones* (Indian

Ocean), *typhoons* (China Sea) and *Willy-Willies* (Northern Australia). These appear on synoptic charts (Fig. 54) as exceptionally deep lows with a concentric pattern of tightly packed isobars. The central pressure is generally around 950 mb but may be lower: the lowest pressure ever recorded—887 mb—was measured in the heart of a Pacific cyclone. The system may be 50 or 500 miles across.

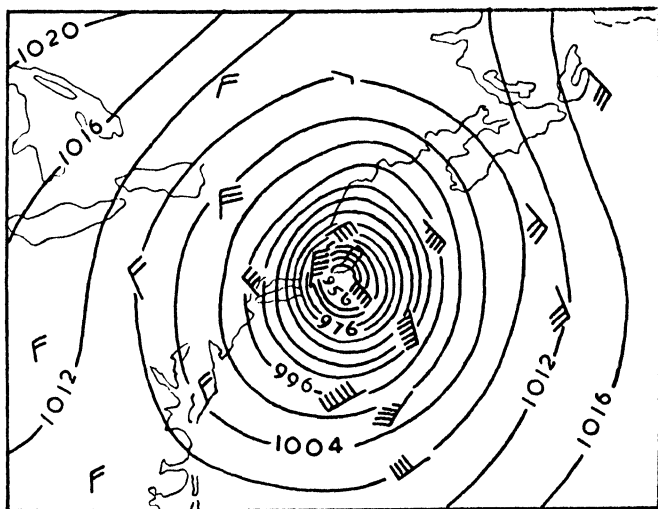


Fig. 54. Hurricane 'Edna,' 11th September, 1954, 1830 G.M.T.

It is not surprising that comparatively little is known of these violent storms in which one's interests are centred on survival rather than on careful weather observing. Probably Nature has conceived nothing as frightening as the hurricane. The fierce winds whip up tremendous seas in which ships are pummelled and tossed like corks. Sheeting rain merges with flying spray until sea and atmosphere seem to be one. Waves and wind and sometimes thunder merge in the pandemonic roar that has been the requiem of many a foundering ship. On shore the rattle of doors and storm shutters and the crash of debris from buildings that succumb to the battering are heard above the shriek of the wind and the pounding of the surf.

In more recent years, determined investigation—including aircraft and radio-sonde as well as surface reports—has enabled us to build up something of a picture of the hurricane. A simplified idea of the supposed structure is given in cross-section and plan in Fig. 55. There is a central core or eye, anything from 5 to 50 miles across, in which the wind abates, the air is warm and dry and there is little or no cloud (Plate XV). This

results from the great centrifugal force developed near the heart of the vortex which flings back the inflowing air to spiral upwards in a ring round the centre: high-level air is sucked down to replace that thrown outwards and is warmed adiabatically and dried in descent.

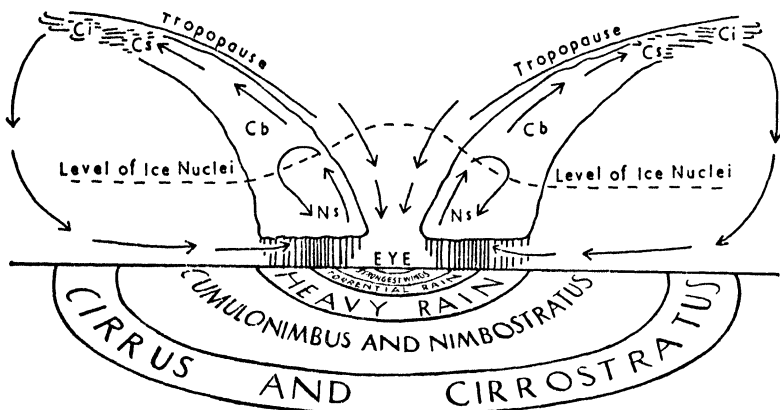


Fig. 55. The structure of a hurricane in vertical section and plan (after Bergeron)

Around the eye is a great cylinder of cumulonimbus and nimbostratus cloud reaching to 35,000 feet or more. Beneath this cloud ring, seething with fierce vertical currents, blow the strongest winds which may exceed 100 or 200 m.p.h. Here too the Bergeron mechanism operates vigorously to give torrential rain falling at a rate of an inch or more an hour. Thus to the observer the storm seems to blow from a particular direction, then follows the deceptive calm of the eye and then the storm 'returns' in full fury, with the wind now in the opposite quarter.

In origin, hurricanes are in part convectional, forming only over the warmest tropical water. Most of the energy that drives them comes from the latent heat freed during the very heavy condensation in air kept highly charged with moisture from the underlying sea. They form within the Tropics (Fig. 56) but not too near the Equator (where the deflecting force due to earth rotation is too small for a circulation to develop). Much evidence suggests that they form on or near the Inter-Tropical Front: at any rate low pressure, *i.e.* a general tendency to convergence, seems a necessary condition for their formation. Their occurrence is seasonal, being mainly in summer and autumn.

Hurricanes move—quite slowly, generally less than 15 m.p.h.—in compliance with the general circulation of Tropical regions, that is, westwards at first, then curving polewards towards middle latitudes where they turn eastwards (see Fig. 56) and soon die out over cooler waters. Some, however, draw fronts into their circulations and become revitalised as middle-

latitude depressions: more than one of our own homely frontal lows began life as a West Indies hurricane. Sometimes the storms move over land where, cut off from the inexhaustible supply of heat energy provided by the warm ocean and their winds countered by greater friction, they usually degenerate rapidly.

It is in coastal areas that the great hurricane catastrophes usually occur (Chapter 11). But sometimes hurricanes break their habitual rules of behaviour and move vigorously inland, like the New England hurricane of 1938 which whipped 150 miles inland at 65 m.p.h. Autumn 1954 will long be remembered in eastern North America for its unusually severe hurricane activity in which 'Carol,' 'Edna' and 'Hazel' each took their toll in deaths and damage: 'Hazel' in particular left devastation in her wake from the Caribbean to the Arctic (Fig. 57).

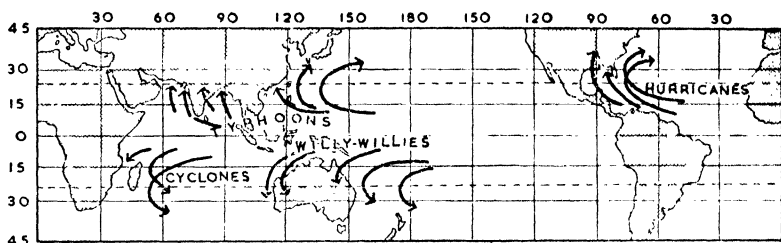


Fig. 56. Frequent tracks of tropical hurricanes

Tornadoes

Tornadoes are even more violent than hurricanes but being much smaller (averaging 250 yards in diameter) their effects are much more localised and therefore less generally destructive. They form over land during afternoon or early evening on spring or summer days. They are again vortical systems with an intense cyclonic whirl around a vertical or inclined axis. The wind speeds are quite fantastic, reaching perhaps 300 m.p.h. (these are estimates based on the damage caused). There is always a large cumulonimbus cloud with a characteristic *pendant* or *funnel cloud* from its base to near the ground (Plate XVI). This is due to the great pressure fall in the centre of the whirl which causes adiabatic cooling and, hence, condensation. Within the funnel a narrow eye (a few yards across) is sometimes observed.

The tornado is accompanied by violent thunderstorms, heavy rain, sometimes giant hailstones, and a general roaring sound compounded of wind, thunder and the noise of destruction. The damage is caused by the combination of pressure and suction effects exerted by the wind: dust and debris (often including trees, roofs or even vehicles) are sucked up before

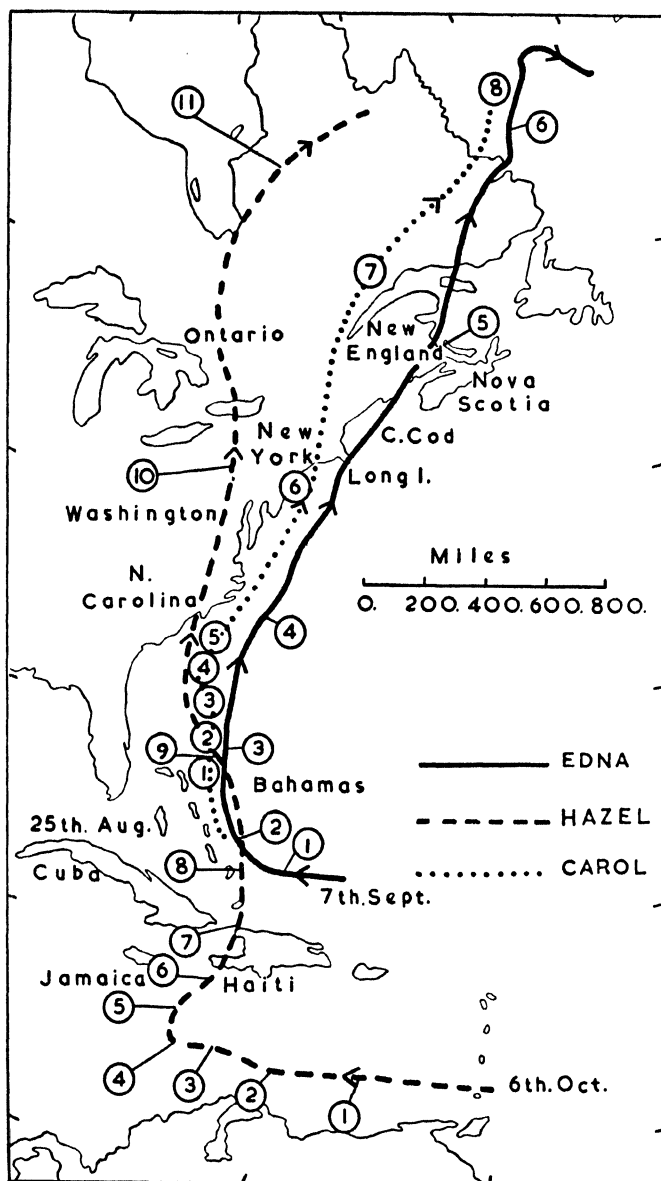


Fig. 57. The tracks of Hurricanes 'Carol,' 'Edna' and 'Hazel.' The figures indicate the positions reached at so many days since inception

being flung outwards by centrifugal force. These then become powerful tools of destruction. Tornadoes travel usually at 30 to 40 m.p.h., but occasionally much faster. Their path may be long or short, straight or tortuous. Happily, their duration at any one spot is unlikely to be long.

Like hurricanes, tornadoes are convectional phenomena, occurring in very unstable damp air and largely maintained by the release of latent heat. They nearly always form on or near an active cold front. Local surface conditions conducive to strong convection often play a contributory part.

North America suffers most from tornadoes and nearly all the 150 or so reported there annually occur in the Mississippi lowlands. Temperate regions have occasional visitations, as we do here in Britain. Recent examples include the Buckinghamshire tornado of 21st May, 1950, the Tibshelf (Derbyshire) tornado of 19th May, 1952, and the unusual winter tornado (8th December, 1954) which moved from near the Isle of Wight through Sussex and Surrey into north-west London.

The equivalent of the tornado over the sea is the *waterspout*, frequently seen over warm waters. Sometimes, these are land tornadoes that have moved out to sea. Some, however—*fair weather waterspouts*—are more akin to dust-devils (Chapter 4).

High-Pressure Systems

The anticyclone is the meteorological counterpart of the low in most respects. It is, however, often a much larger system, with more widely spaced and less streamlined isobars. These indicate a slack pressure gradient and indeed, while lows are areas of vigorous circulation, highs are relatively stagnant portions of the atmosphere. Except on the margins, calms or only light and variable surface winds prevail.

The key to the understanding of anticyclonic weather is subsidence, the slow down-spiralling of air which is warmed adiabatically in descent and therefore dried. It is to this fact that anticyclones owe their reputation (not always deserved) for fine weather. Subsidence imposes great stability on the air, clearly visible on a tephigram or temperature-height diagram as a *subsidence* or *anticyclonic inversion*. This first appears at high levels and slowly descends to within 1,500 to 5,000 feet of the surface where it remains (Fig. 58). Very dry air distinguishes the anticyclonic inversion from other types.

Anticyclones represent an excess of air and the problem is to account for it. Sometimes there is cold heavy air in the lower troposphere (*cold highs*). In other highs, the troposphere is unusually warm throughout and it seems that abnormally cold air in the overlying stratosphere must be responsible for the high pressure at the surface. These *warm highs* are

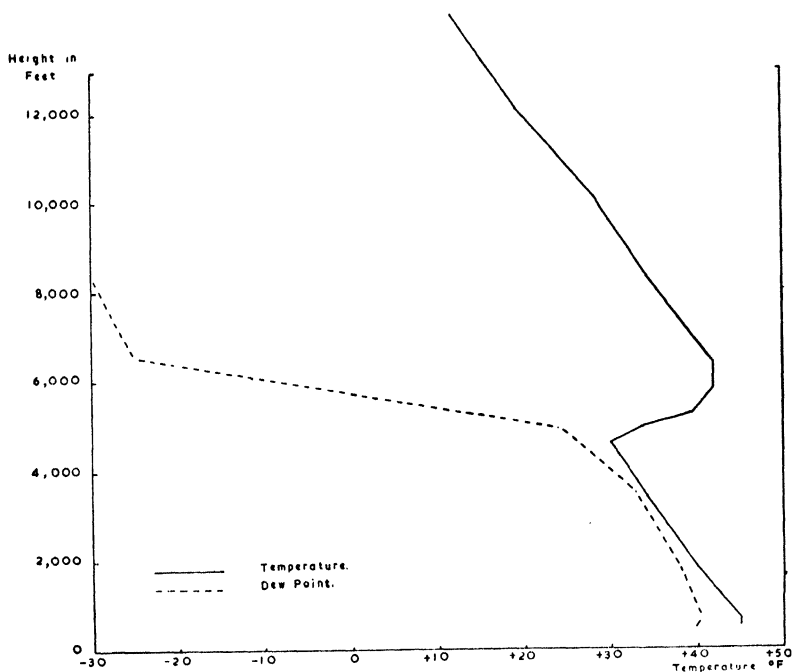


Fig. 58. A subsidence or anticyclonic inversion on the temperature-height diagram. Crawley, 30th December, 1954, 0200 G.M.T.

something of a mystery. Within each category there are highs that are more or less *permanent* features of the global pressure pattern, some that are *seasonal* and others that are *temporary*, making only a brief appearance on the weather map.

Cold Anticyclones

The cooling of part of the earth's surface may give rise to a *thermal high* by a process that is the exact reverse of that producing a thermal low. The column of air above the cold surface is itself cooled and contracts or shrinks vertically, leaving less air at high levels (upper low) and inducing an inflow. This increases the total weight of air in the column and produces the surface high. The process now continues as described earlier (page (154)).

This mechanism is best seen in the great *seasonal anticyclones* that develop over the continental interiors in winter. Here under the vast cushion of cold dense air are registered the highest-known surface pressures (1,075 mb has been recorded in the heart of the Siberian High). But pressure decreases very rapidly with height and even at 5,000 feet the

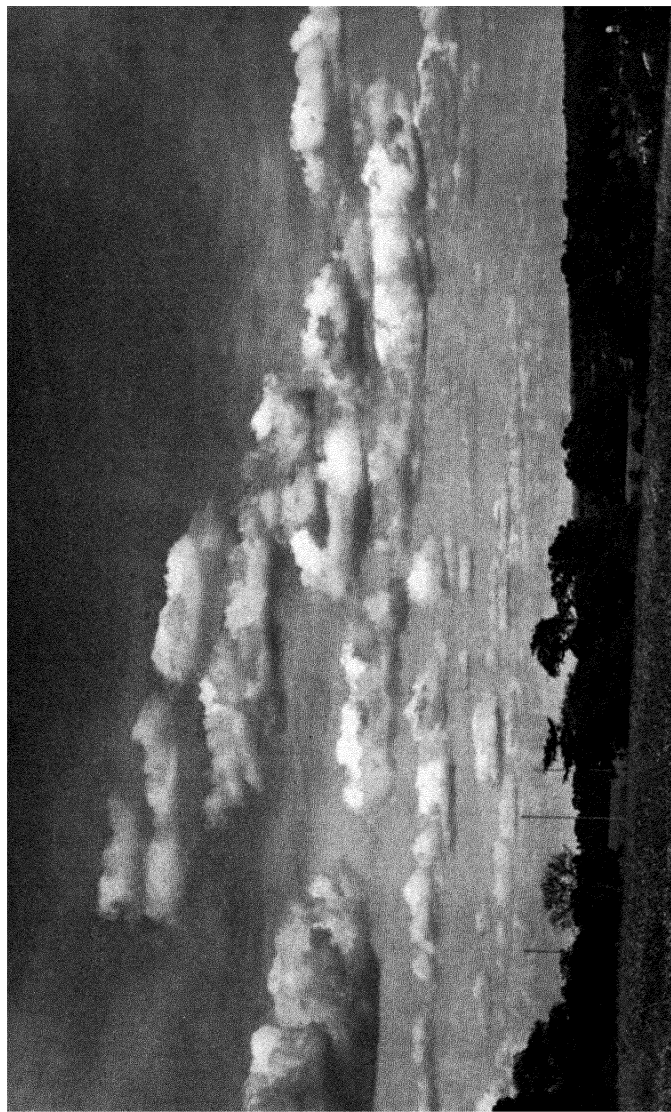


PLATE XIII. Anticyclone weather, summer: fair weather cumulus

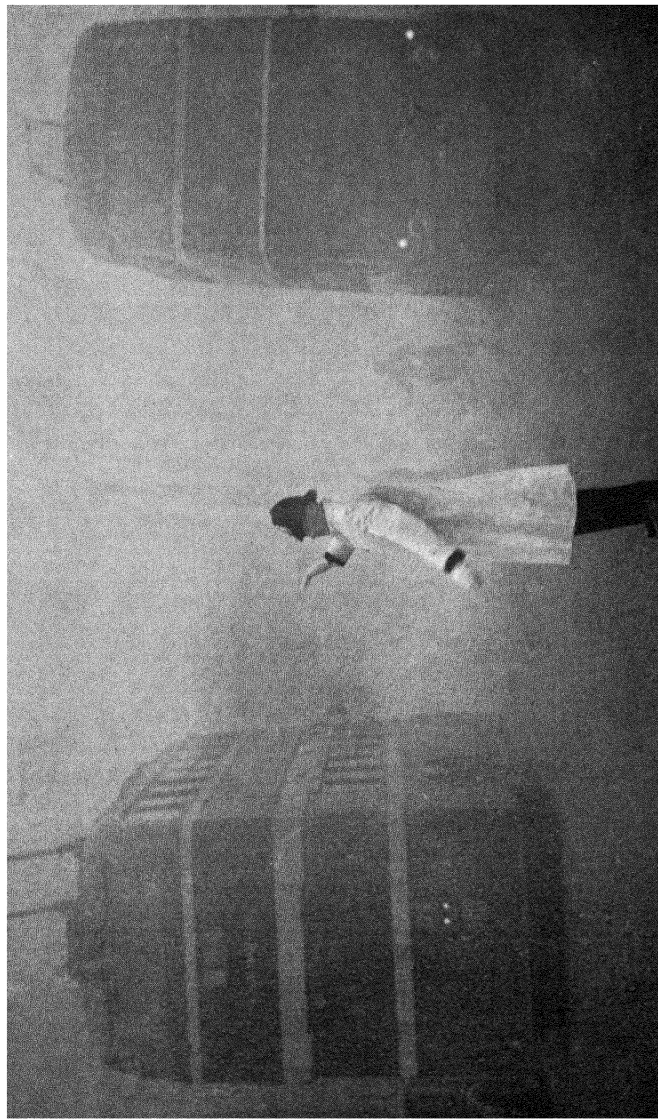


PLATE XIV. Anticyclone weather, winter: fog slows up traffic

anticyclone has largely vanished while above it is replaced by a cyclonic circulation (which may be an upper trough rather than a closed low).

Transitory cold anticyclones developed in Pm air separate the successive members of a depression family and provide the brighter interludes (lasting perhaps a day) of these unsettled spells. Often they are no more than ridges but sometimes, especially in the southward surge of cold air in the rear of the last depression of the family, they build up into substantial centres (Fig. 30). They are again relatively shallow systems.

The weather brought by these temporary cold types—the most common we experience—depends partly on the season but always stability is the key factor. In summer, the convection cloud typical of Pm air builds up but is limited by the inversion to flattened fair-weather cumulus (Plate XIII). In winter, especially with very moist air, extensive and persistent stratocumulus sheets may spread out beneath the inversion. Urban and industrial smoke is trapped under this thermal 'lid' and the combination of cloud and smoke sometimes produces a veritable darkness at noon which Londoners know only too well and which has been dubbed *anticyclonic gloom*. At night these cloud sheets disperse and there is the danger of frost and radiation fog inland (see Chapter 5 and Plate XIV).

Warm Anticyclones

As pressure decreases only slowly with height in warm air, warm anticyclones are still evident even on high-level charts. They are the most stable, slow-moving and persistent of pressure systems. Permanent warm anticyclones are found over the sub-Tropical oceans. Our nearest example, the Azores High, is a near-constant feature of the weather map, though its actual position and size vary from time to time. The *Sub-Tropical Highs*, we shall see, are integral parts of the general circulation.

From the Azores High *temporary warm anticyclones* sometimes extend towards Britain, either as ridges or as separate centres which bud off the main high. When this happens we see the best of Tm weather (Chapter 5). Another kind of temporary warm high forms by gradual change from a temporary cold type through the warming agency of the subsidence process. This transformation commonly occurs when a cold high remains stationary for a long period and the warm high thus produced often gives very fine bright weather.

Anticyclonic Spells in Britain

It will be clear that anticyclones always provide dry, but not necessarily good, weather. Broadly speaking, our summer highs, formed in Tm or sometimes Pm air, can be relied upon for fine or fair weather (windward

coasts excepted) but winter highs, in Pm or sometimes Pc air, may give bright weather or, just as often, dull or foggy conditions.

Studies by Belasco of the frequency of occurrence of anticyclonic weather in Britain over a 50-year period show that September is the month with most anticyclonic days on average ($10\frac{1}{2}$) while July and August have least (only 5 each). These are also the two months most likely to have no anticyclonic days at all. The implication of this for summer holidaymakers is obvious but, of course, no particular year can be relied upon to conform to average. 1947 and 1955 in recent years provided notable exceptions. February, May and June are months with more than the average number of anticyclonic days, which is just over 7 per month.

The General Circulation

And so to the major problem in meteorology, the grand scheme in which all the elements we have considered—the air masses, the fronts, the pressure systems—find their due place. It must be said at once that no one has yet put forward a satisfactory theory of the general circulation. Not enough is known of the facts let alone their interpretation. Much of what follows in this section is therefore tentative and open to challenge. In this branch of meteorological thought fashion is apt to change rapidly and often.

In Chapter 3 we drew attention to the unequal heating of the earth's surface and suggested that this must be ironed out by the transport of excess heat from low to high latitudes. In this lies the need for the general circulation, which we therefore recognise as primarily thermal in character. But the problem is greatly complicated by the rotation of the earth and the varied nature of its surface.

The Mean Upper Circulation

Meteorologists today increasingly take the view that the upper-flow pattern gives a truer picture of the basic circulation than the surface pattern which is so much more vulnerable to chance irregularities. Average pressure and wind charts for levels in the middle and upper troposphere show that the dominant feature of the global circulation is a vast *circum-polar vortex*, a westerly whirl of air above all latitudes except the Equator (where easterlies seem to prevail). These upper westerly winds reach maximum strength on average at around latitude 30° : this is the sub-Tropical Jet Stream (the Polar Front Jet Stream is too irregular and variable to show up separately on average charts).

The upper westerlies blow between high pressure aloft over lower latitudes and marked upper lows over the Polar regions. This arrangement

follows from the distribution of mean temperature in the troposphere, *i.e.* warmer towards the Equator and colder towards the Poles (see Fig. 50). The upper winds are predominantly thermal in character.

The 'Idealised' Surface Circulation

Near the surface the distribution of land and sea and the irregularity of the land so disturb the flow that it is difficult to say what the circulation would be like on a uniform globe. Perhaps only over the wide Pacific Ocean do we approach this ideal pattern. Even here, as average wind charts show, the picture is complex enough. The westerlies that dominate the upper flow extend to the surface only in middle latitudes. They are flanked to poleward by the *Polar Easterlies* and equatorwards by low latitude easterlies better known as the *Trade Winds*.

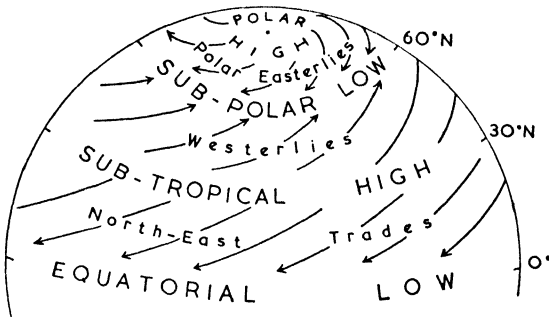


Fig. 59. The idealised 'planetary' circulation

The pressure distribution supports these winds in accordance with Buys Ballot's Law. Considering the Northern Hemisphere, the Trades blow between a sub-tropical belt of high pressure at about latitude 30° ('Horse Latitudes') and an Equatorial low-pressure region ('Doldrums'). Because of surface friction, they blow somewhat across the mean isobars towards low pressure as North-east Trades. The Westerlies become South-westerlies, blowing from the sub-Tropical High towards a low-pressure belt (sub-Polar Low) at 55°–65° N. while the Polar Easterlies (or North-easterlies) blow out from the Polar High. Fig. 59, which shows these relationships, is of course the *idealised planetary circulation* we were taught at school. A hypothetical circulation referring to a uniform globe, this is far from the realities of the situation. It serves, however, as a stepping stone to fuller understanding.

How is this pattern to be interpreted? In the very simplest terms, we can regard the Equatorial low-pressure zone as a great elongated thermal low and the sub-Tropical Highs of each hemisphere as the complementary

flanking highs. Air ascends in the low, moves outwards (upper divergence) as high-level westerlies with a slight poleward inclination, descends in the sub-Tropical Highs and returns as the Trades to feed the Equatorial low-level convergence. It may be that the sub-Tropical Jet Stream is the simple consequence of the inevitable apparent increase in westerly momentum undergone by these high-level winds as they drift slowly poleward over smaller latitude circles (*see* page 146). The high-latitude circulation is also essentially thermal: air descends in the Polar (thermal) Highs, flows outwards (Polar Easterlies), ascends in the sub-Polar Lows and returns as upper westerlies. The sub-Tropical Highs are zones of low-level divergence, feeding both the Trades and the Westerlies. The two thermal circulations are linked, as we shall show, by the middle-latitude zone of travelling depressions and anticyclones.

Pressure Cells in the Surface Circulation

As a step nearer reality, we must acknowledge the existence of the continents and oceans. Their effect broadly is to break up the pressure belts into separate *cells*, each with their independent circulations. In the Northern winter the intense cold of the Eurasian and North American land masses gives rise to two great thermal highs which dominate the circulation of the hemisphere. They extend northwards to split the sub-Polar Low into two cells, the Icelandic Low (North Atlantic) and the Aleutian Low (North Pacific). The sub-Tropical Highs over the oceans—Azores and North Pacific Highs—are overshadowed in strength by the great Continental Highs.

In the Northern summer the thermal systems over the continents are transformed. A great monsoon low develops over southern Asia and a smaller one over North America. These completely interrupt the sub-Tropical high-pressure belt leaving only the Azores and North Pacific cells which, however, stand out more strongly than in winter. The Asiatic monsoon low is so intense that it swallows up the Equatorial Low in the Indian Ocean, thus creating a continuous pressure gradient from the Southern Hemisphere sub-Tropical High to the Asiatic low. The sub-Polar Lows are weak in summer, overshadowed by the continental monsoon lows.

In the Southern Hemisphere, with so much less land, the mean circulation is less cellular, and is more like the theoretical circulation on a uniform globe. Nevertheless, the sub-Tropical High still consists of three centres, one in each ocean, especially in the Southern summer. But because of the virtual absence of land in this latitude, the sub-Polar Low extends as an almost unbroken belt.

We can now demonstrate in a general way how the patterns of air masses, fronts and pressure are inter-related. Fig. 60 shows, for mid-

winter and mid-summer months respectively, the average sea-level pressures, the mean surface air flow, the location of the air masses and the average positions of the major fronts. Each high pressure cell is a source region from which the air mass moves out in obedience to the anticyclonic circulation. From the Azores High, for example, air flows equatorwards as part of the North-east Trades (in air mass parlance TmK air and also polewards as part of the Westerlies (TmW)).

It is the location of the pressure cells that determines where contrasted air masses shall converge (frontogenesis) or diverge (frontolysis). Thus Pm air sweeping south-eastwards in the circulation of the Icelandic Low meets Tm air at the Atlantic Polar Front. Or cold Pc air, bred in winter over Siberia, flows out, becomes modified to Pm over the Pacific and eventually forms a front with Tm. On the other hand, at the eastern edges of the sub-Tropical High and sub-Polar Low the air flow is divergent and frontogenesis does not occur. Air of various origins makes up the Trade Winds and the two great streams from north and south converge at the Inter-Tropical Front.

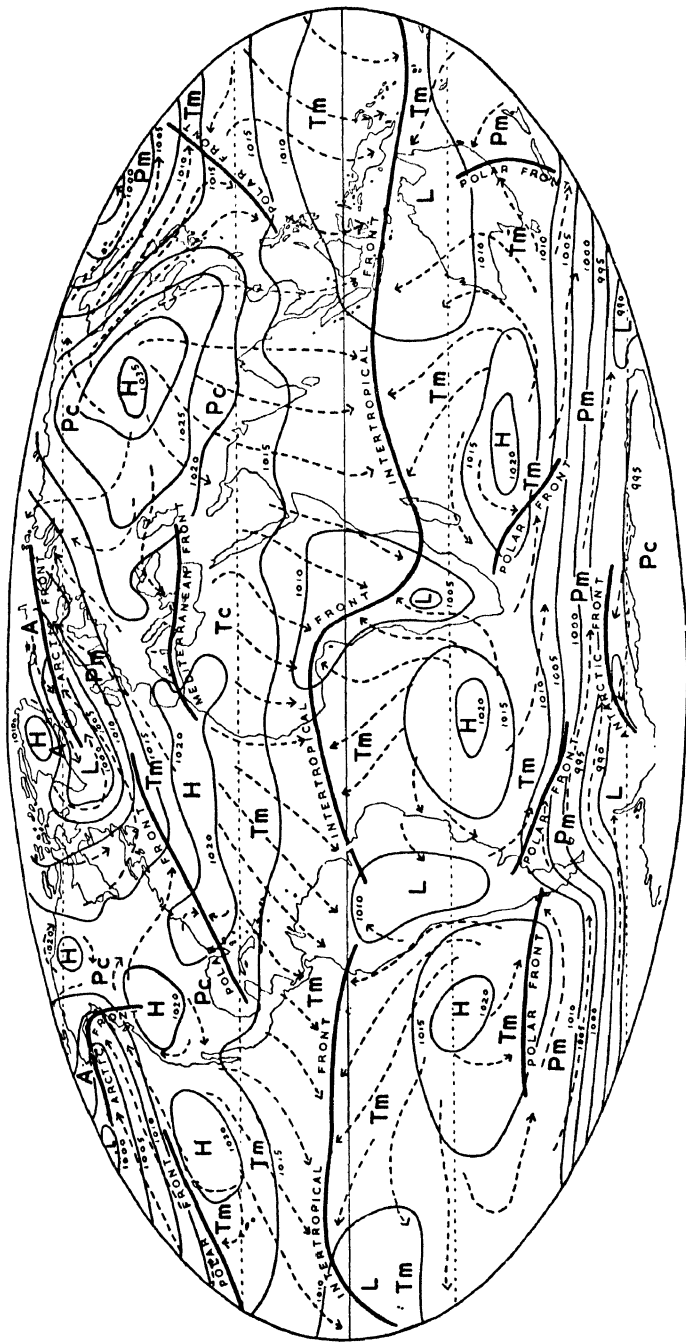
Realities in the General Circulation

Although we may learn much from maps of mean conditions, averages in meteorology often conceal as much as they reveal. The pressure cells are not fixed features, even the 'permanent' sub-Tropical Highs fluctuate in strength and location. The so-called 'Icelandic Low' is an abstraction representing the fact that travelling depressions are particularly frequent in the Iceland region. The major fronts, lines on the mean circulation maps, are pulsating zones of cyclogenesis.

These maps provide a static view (an abstract one, at that) when we really require a moving picture, a cinematic view. Moreover, the cellular patterns of Fig. 60 comprise a number of closed systems with independent circulations, whereas the prime need of the general circulation is the engineering of mass exchange of air between high-latitude and low-latitude circulations. This cannot be understood from mean circulation patterns. We must seek the moving picture.

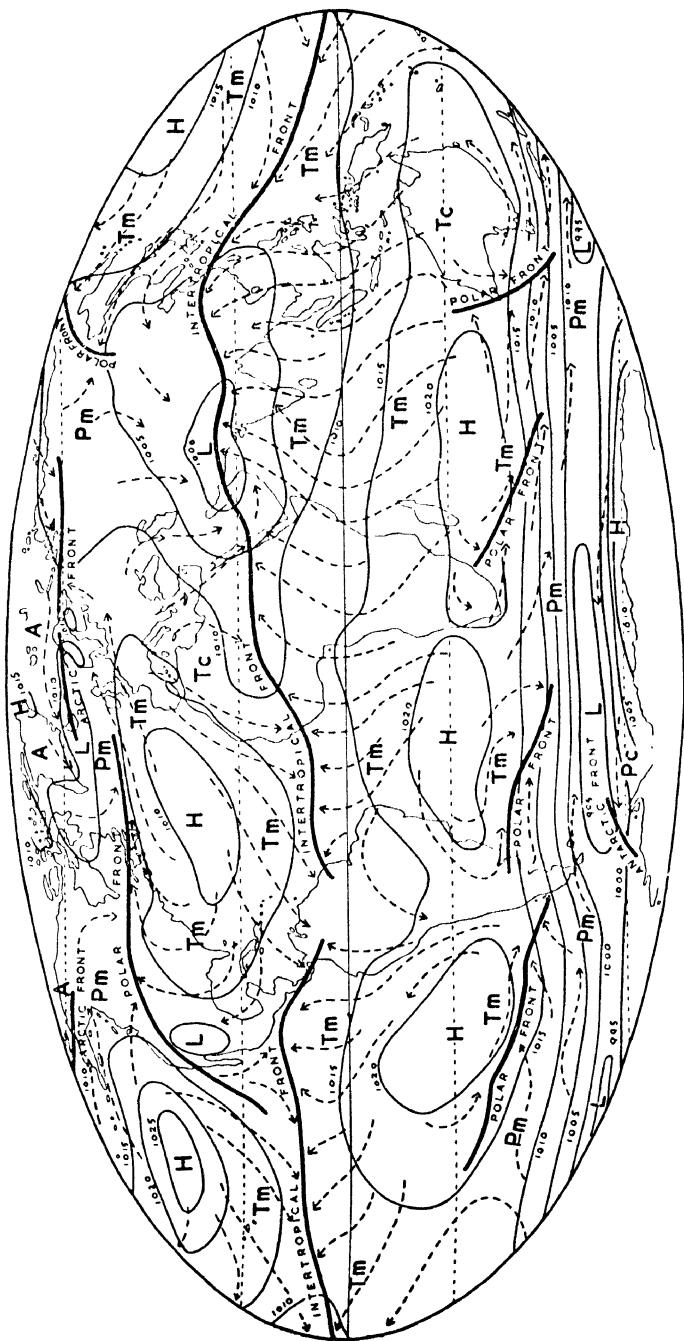
Long Waves in the Westerlies

The realities of behaviour of the upper westerlies are illustrated in Fig. 61. This shows, on a polar projection of the Northern Hemisphere, contour lines of the 500 mb surface, which we may regard as depicting the air flow at about 18,000 feet. The circumpolar whirl is not steady, as mean charts suggest, but oscillates in ridges and troughs—*upper long waves*—around an ever-changing, amoeboid-shaped high-level Polar low. Fig. 61



(a)

Fig. 60 (a). Average surface circulation in January, showing mean sea-level isobars (continuous lines), mean winds (broken arrows), mean positions of fronts (heavy lines) and air masses



(b)

Fig. 60 (b). Average surface circulation in July, showing mean sea-level isobars (continuous lines), mean winds (broken arrows), mean positions of fronts (heavy lines) and air masses

is not an actual chart but shows a common pattern and gives an idea of the number and size of the waves. Embedded within the meandering westerlies is a belt of strong winds (*i.e.* close-packed contours)—the Polar Front Jet Stream.

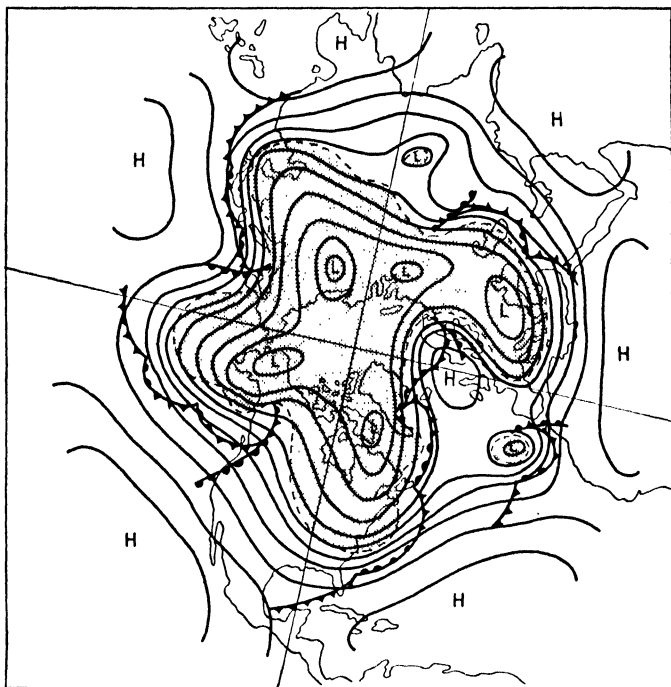


Fig. 61. Schematic circumpolar chart showing long waves in the upper westerlies (after Palmén). Thin lines are 500 mb contours; the dotted line represents the Polar Front at that level. The extent of polar air at 500 mb is stippled

The surface fronts have been inserted to show broadly how high-level and surface patterns are related. The upper troughs are regions where polar air is displaced equatorwards and the upper ridges regions where tropical air is displaced polewards. In the waving upper westerlies we may perhaps see the reason for the tussle between air masses at the Polar Front. An upper wave is linked sometimes to one large surface depression but more often to a depression family: each member is represented by a minor ripple superimposed on the long wave (these are omitted from the diagram).

The waves move round from west to east in an uneven, pulsating and constantly distorting pattern. Sometimes a particular wave becomes so

distorted that the southern portion is pinched off, remaining as a *cut-off low*, a pool of cold air stranded in lower latitudes. Part of an upper ridge may similarly become detached. Examples of both are shown in the diagram.

Why should the upper flow be disturbed in this way? It may be that we must accept this mode of behaviour as a property of our atmosphere, the dimensions of the earth and the nature of its surface being as they are. The great mountain barriers of America and Asia, extending to such heights that half or more of the weight of the atmosphere lies below their peaks, are bound to interfere with the general flow. Air piles up to windward and leaves a deficit to leeward. The continents develop their own circulations which are superimposed on the global one. Combine these effects, add any number of smaller-scale geographical influences and the resulting picture can be complicated indeed.

The Role of Middle-Latitude Travelling Disturbances

In all this can be seen the role of the travelling depressions and anticyclones or of their upper-level counterparts in the long waves of the westerlies. Undisturbed, the Polar Front is a barrier separating warm air from cold. Here is a mechanism that breaks it down. In the warm sector of a depression tropical air is drawn far to poleward while in the rear and in the following cold anticyclone, polar air surges equatorwards.

Sometimes the temporary cold high builds up strongly (especially in the rear of a depression family) and coalesces with the sub-Tropical High. The intervening front is eventually wiped out by frontolysis. This seems to be an important way in which polar air gains entry into the low-latitude circulation. Thus depressions and anticyclones act as travelling exchange mechanisms throwing warm air polewards and cold air equatorwards.

The Ever-Changing Circulation

The general circulation presents a constantly changing picture. Recent studies suggest that the fluctuations are not haphazard but show a swing between two extreme patterns. In one the circumpolar vortex is at its strongest, the upper flow is markedly *zonal* (along a line of latitude) and the exchange of air between high and low latitudes is at a minimum. For us this means unsettled 'westerly' weather, the rapid succession of lows and highs from the Atlantic, much wind and rain and temperatures above normal in winter and below normal in summer. The summer of 1954 will be remembered as a period in which this type of circulation dominated.

At the other extreme, the flow of air is mainly *meridional* (along a line of longitude). Great upper highs lie athwart the usual path of the

westerlies. Under these *blocking* conditions, warm air encroaches northwards even to Arctic regions and cold air thrusts unusually far south into sub-Tropical latitudes. This situation allows the maximum exchange of air across the lines of latitude. It gives us persistent weather, sometimes unseasonal, depending on which stream of air holds sway. The early months of 1947 and 1956 provide recent examples of winters dominated by Pc air under these conditions.

In such complex ways as this the general circulation of the atmosphere is achieved. With large gaps in our knowledge, we are still a long way from understanding exactly how it is done.

WORLD WEATHER TYPES

In the course of a lifetime spent with the British weather we cease to wonder at its extraordinary behaviour and come to accept with equanimity its most extravagant excesses. By long exposure to its influence we have, as it were, acquired a degree of immunity from surprise at anything it can throw at us. Other countries, too, have their weather—not many of them as interesting and varied as ours, it is true, but, like us, their inhabitants become accustomed to the weather they know.

However, if we stay some time in foreign parts, in Bombay or Winnipeg or Manila or Cuzco, even the least observant of us will notice differences from the weather we get at Edinburgh or Southend, and we may begin to ponder wherein the difference lies. One thing we shall notice is that British weather, though very wide-ranging, has no outstanding qualities other than its changeability. We get some very cold spells and complain about them bitterly enough, but our worst are not nearly as cold as the winter at Winnipeg; our hottest spells are mild compared with Bombay in May; our floods and cloudbursts, frightening though they may be to anyone involved in them, are second class compared with the record rainfall of 45 inches in 24 hours in the Philippines; our strongest gales pale into insignificance beside a Chinese typhoon. Our weather, in fact, is a versatile all-rounder but lacks the high degree of specialisation needed for record-breaking.

With this contrast between versatility and specialisation in mind we may recall the comment that other countries have climates, but Britain has only weather. This is an exaggeration, but not without a germ of truth; it is our exposure to so many influences, polar and tropical, continental and oceanic that explains our variety. Those territories that lie near the centres of air masses or in steady streams of air-flow like the Trade Winds have weather so uniform and reliable by comparison that it might well be referred to as climate. The forecaster's task in such places is relatively easy and it is rarely necessary to enter the automatic British caveat 'Weather permitting' when fixing some outdoor entertainment.

But this implied contrast between weather and climate raises the questions 'What is climate?' 'How does it differ from meteorology?' 'What is a climatologist?' The short but imperfect answer is that climatology is

concerned with the average condition of a place with respect to temperature, rainfall, humidity, sunshine, etc. That is to say, the climatologist describes each element of the weather separately, using the quantitative instrumental and observational records as his data. Such is not our object in this book, as the title of this chapter shows. We wish to keep together in their proper perspective and interrelationships all the varied aspects of the weather—which, dismembered and separately described, constitute the climate in the traditional treatment. An alternative definition of climate is sometimes given as ‘the average weather.’ But attempts to implement this definition have hitherto broken down because weather is such a complex thing, involving so many different elements (*e.g.* wind, storm, rain, warmth or cold, sunshine, cloud, mist or fog) that it has eluded orderly classification.

Can we not find some all-embracing concept which will convey the whole complex of weather associations in much the same way as ‘Parkinson’s disease’ or ‘General Paralysis of the Insane’ conveys to the trained mind of the doctor a fairly comprehensive idea of the patient’s appearance, state of health, suffering and future prospects, without the necessity of enumerating and describing the symptoms one by one. We believe that the best unit for this purpose in meteorological description might well be the air mass. Now that we have learnt to associate certain types of weather and successions of weather events with each air mass we can conjure up a mental picture of the general features of what we might call the ‘weatherscape’ simply by quoting the air mass. This is similar to the device used by the geomorphologist who can conjure up a picture of the physical ‘landscape’ by referring to three variables: (1) geological structure, (2) the operative process of weathering and erosion (humid, arid, glacial or marine) and (3) the stage reached in the evolution of the erosional cycle (youth, maturity or old age).

If the method is valid it gives us a genetic basis for a truly dynamic system of classification of weather types. But there are two further problems to be solved before the system can be extended to the classification of geographical climates, for the weather types are on the move, travelling across the face of the earth whose geographical features are fixed. It is therefore necessary in the first place to know where each air mass is likely to be found at each season of the year, *i.e.* to know the normal seasonal fluctuations of the atmospheric circulation as shown in Fig. 60 and on page 166. In the second place we have to remember that weather is in some measure a product of the geographical environment as well as of the nature and life history of the air mass in residence or in passage. It is therefore necessary to know the reaction of each type of air mass to each type of relief and surface with which it may come in contact.

All these things have already been studied in the preceding chapters

and we are now in a position to find an explanation for the weather at any place on the earth's surface in an appreciation of the interaction between each air mass of known quality and behaviour and the known geography of its position and path.

In our description it will be simplest to start at the equator and to work towards the increasing complexity of the so-called temperate regions to which we have devoted our attention in the preceding chapters. We can then pass on to the relatively simple weather of the polar regions. It is as well, however, to appreciate that the apparent simplicity of the equatorial and polar weather may partly be attributed to the general state of our ignorance about them, the scanty attention they receive and the scattered and inadequate network of recording stations.

Equatorial Weather (The Amazon and Congo Basins, Ceylon, Malaya)

'Uniformly hot and moist' is the usual description of the equatorial climate and it could hardly be otherwise. No cooling breath from the temperate regions can reach these latitudes and the dry air of the deserts cannot penetrate so far and still retain its character, for nearly four-fifths of the length of the equator is ocean, and only in crossing South America and Africa does it go over the land. Nor is there ever in these low latitudes an anticyclone in which a true air mass can develop. On the contrary, it is a region of permanent low pressure into which air is being drawn (Trade Winds) and perpetually escaping, usually upwards. The use of terms like 'air mass' and 'front' therefore have least meaning here. What is often referred to as an Equatorial air mass (E) is more truly a sluggish air stream of Tm air, and the ill-defined boundary between convergent air streams, usually referred to as the Inter-Tropical Front (I.T.F.) rarely has any true frontal character at all and is more accurately described as an Inter-Tropical Convergence Zone (I.T.C.Z. for short).

In spite of the seasonal north-and-south sway of this inter-tropical zone of calms (the Doldrums) and convergent winds, some parts of the world, such as the Amazon and the Congo Basins, remain under its influence throughout the year. In such places there are no seasonal changes, no summer or winter, no dry season. So long as the typical equatorial conditions persist at any one place, one day is very like another, it begins in the same way and runs much the same course. The daily sequence of weather is more important than the seasonal change. Sweaty heat is the order of the day and some measure of relief is found only at night; night, it is said, is the winter of the tropics, and though the drop in temperature is only a few degrees it seems chilly enough to call for a blanket. The nocturnal cooling may produce an inversion of temperature and even morning fog in the damper forest belts near the coast. Fog quickly clears

and the inversion breaks down when the sun abruptly rises about 6 a.m. As the day wears on convection sets in and cumulus clouds appear, building up into cumulonimbus from which heavy thundery rain may fall. The time of onset of the daily rain is in many places characteristically regular and highly reliable. The sun goes down about 6 p.m. and after the briefest of twilights the cycle of radiant cooling and clear night skies begins again.

In the equatorial latitudes south of the great land mass of the Old World things are rather different because the monsoons flow back and forth in summer and winter, sweeping away the Doldrum calms. Such regions, e.g. Ceylon, the East Indian archipelago and, less markedly, the coast of West Africa and the north coast of Australia have complex climates which are monsoonal for two quarters of the year and only typically equatorial during the 'spring' and 'autumn' seasons of calms at the change-over of the monsoon. The Tm air of the summer monsoon brings rain, to windward coasts and mountain slopes, and this is probably the wettest season of the year in such exposures, but on the lee-sides of islands there may well be a dry season as in the tropical monsoon climates described below.

The response of nature to the constant heat and humidity is vigorous. Vegetation grows apace and the natural plant cover is a dense forest of evergreen trees, many of them giant species. As there is no seasonal rhythm growth goes on all the year round and there is no special season of flowers or fruit. Moulds, fungi and bacteria thrive and grow quickly causing wood, leather and fabrics to rot. Insect life is exuberant, invading the house and fluttering round the lamps at night. Many flies and mosquitoes carry diseases such as sleeping sickness and yellow fever. The evil reputation of such climates (e.g. 'the White Man's Grave') is largely attributable to the bacterial decay of organic substances and the proliferation of parasites and disease-bearing organisms. It is being overcome by proper attention to sanitation and hygiene. At the same time there is no doubt that the humid heat is enervating, makes a man lethargic and lowers the resistance of his body to attacks of disease.

Tropical Trade Wind Coasts and Islands (e.g. Durban (Natal), Queensland, East Brazil, the Caribbean Islands and Coasts)

Visited in mid-summer by the equatorial calms, these regions lie for most of the year in the track of TmK air of the Trade Winds, whose cooling effect offers a pleasant contrast to the sultry heat of the Equatorial (E) air just described. Over the open sea the Trade Winds generally give fine weather, with blue sky flecked with cumulus cloud, covering perhaps a quarter to a half of the sky. Their base is generally about 2,000 ft. and the tops rise to 6 or 8,000 ft. where there is generally an inversion of tempera-



PLATE XV. Tropical weather: inside the 'eye' of a typhoon. This remarkable photograph of the Pacific typhoon, 'Marge', was taken by Dr. R. H. Simpson (Director of the National Hurricane Research Project, U.S.A.) from an aircraft at a height of 17,000 feet



PLATE XVI. A tornado in Britain. The Bedfordshire tornado of 8th June, 1955

ture due to the descent of air from above, warmed in the process. Only rarely do they rise to 10 or 12,000 ft. and give scattered showers.

Where islands stand in the way, the orographic lift given to latently unstable air, combined with the effect of the daily convection, causes a cloud-banner to form by day and rain may be produced. On large or mountainous islands rain usually falls from the large clouds formed, and the windward sides of such islands are among the wettest places on the earth. The north-east facing slopes of Hawaii behind Hilo get more than 150 inches a year. But descent on the lee side warms and dries the air, dissipating the cloud. Furthermore, the daily sea-breeze, fighting against the prevailing winds, makes the lee side relatively calm and generally dry by day; there is almost a desert on the south-west of Hawaii with only about 30 inches of rain per year.

The same circumstances that are seen in the islands are repeated on a larger scale on the coasts of continents, and since the coasts in this region are, for the most part, hilly or mountainous, the orographic lift that they give to the air sets off the rain in heavy showers; the coasts are well-watered, but droughts are often serious in the 'rain shadow' inland. There is no dry season on these Trade Wind coasts, for both E and TmK air are unstable, especially by day, but there are seasonal differences of temperature and of wind direction and force.

Into this generally serene climate occasional stormy passages are introduced by tropical hurricanes. They occur at the end of the summer during occupation by the I.T.C.Z. Troughs develop along the edge of the convergence zone and travel westwards giving storms and heavy rain. Sometimes such storms, aided by active convection, grow into hurricanes (see page 156) which may come ashore doing considerable havoc. The favourite place for the birth of such storms lies on the western sides of oceans in about latitude 20°, far enough from the equator for differential rotation rates to produce deflection and give rise to strong vortices.

Air from polar sources rarely invades these climates, but occasional waves of relatively cold Pm air, creeping round the sub-tropical high-pressure cells, may lower the temperature into the forties in winter, and night frost is not unknown inland, especially on the plateaux.

The natural vegetation is forest, differing little from equatorial forest.

The Savanna Lands (Dakar to Lake Chad, East Africa, Queensland and Northern Territory, the Parana-Paraguay Basin)

These regions have a climate of strongly-marked seasonal contrast. In the mid-summer months of the appropriate hemisphere they come under the influence of the Inter-Tropical Convergence Zone, but for the rest of the year the air is mainly Tc (e.g. the Harmattan of West Africa) or dried-

out TmK. Summer therefore brings the warm rains, winter the hot drought. The length of the rainy season increases equatorwards until it merges into the continuous wet season of the equatorial type. Polewards the dry season lengthens into the constant drought of the hot deserts under the tropical high pressures where Tc air prevails throughout the year (see page 180).

During the dry season in, say, Northern Nigeria the sky is cloudless in the morning and the air is clear or hazy with dust. The wind freshens during the morning from the north or north-east and dust increases. Sometimes the wind blows with considerable force (the Harmattan). It is a hot and desiccating flow of Tc from the Sahara. Perspiration evaporates as quickly as it forms, keeping the body relatively cool. If, as often happens, this dry Harmattan reaches the humid regions of the coast, it is welcomed for its cooling effect (by evaporation) and is known as the 'Doctor.' But in the interior, where the effects are far from pleasant, the body begins to dry up, thirst is a pain, the skin, hair and finger-nails become brittle and crack; vegetation wilts and crops are damaged by drought.

By the end of the dry season the stunted trees are bare of leaf, the grass is dead and yellow, the soil is parched and the river-courses are dry beds of stone or sand. When the rains come they announce themselves with noisy thunderstorms known as Tornadoes (not to be confused with the prairie tornadoes mentioned on page 159). Dark clouds come up from the south-east, lightning flickers and, as the cloud comes overhead, the rain falls in torrents. This is the invasion of the Inter-Tropical Front, the line of separation between two air masses, here strongly contrasted. Several such tornadoes may occur before the Equatorial air (really Tm for it has come as a sort of monsoon from south of the Equator) is fully established. Thereafter the rain is of the equatorial type but the surface on which it falls is not so soft and spongy as the equatorial forest. If the surface is sandy the water soaks in quickly, but if it is bare rock or impervious clay there is a great loss of valuable water by run-off. Eventually the South-West Monsoon retreats before the advancing Tc from the Sahara and the dry season is restored.

The Tropical Monsoon Lands (India, Burma, etc.)

The pressure systems, the air masses and their associated weather complexes swing slowly north and south following the sun in its yearly movement like the steady rise and fall of the tide upon an open coast. But the monsoon is quite different. It bursts on India like the tidal bore with which the surging flood-tide over-rides the dying ebb in a great estuary like the Severn or the Yang-tse-kiang. It is the sudden over-riding of Tc by Tm air that provides the most characteristic and spectacular event in

the Indian climatic year. Throughout the hot season, from March to early June, the Indian plains have been virtually a self-contained climatic region in which a generous daily measure of scorching sun from a cloudless sky has built up a great body of Tc air in an area of steadily falling pressure.

The weather in the plains can be imagined, it is like the hot deserts. The shimmering heat increases daily until by the end of May it is well-nigh intolerable for white people and most of the Europeans have taken refuge in hill-stations.

Meanwhile in the high pressure cell of the south Indian Ocean prodigious reserves of hot, humid 'E' and TmK air are building up, which eventually spill over the feeble barrier of the dying high of the north Indian Ocean and rush into the Indian Low. The latent instability of this air is set off by the heat of the plains and by the latent heat released during the orographic ascent as the monsoon current surmounts the Western Ghats. Torrential rain arrives with vivid thunderstorms and the blistering dry heat is replaced by enervating steamy warmth in which crops, weeds, moulds, mildews, insects and neuroses flourish inordinately. The strongly prevailing south-west winds of the monsoon season appreciably raise the level of the sea along the west coast, flooding the Ranns of Cutch, which become mud flats or saline marsh when the north-east wind blows in winter.

The burst of the monsoon is a truly frontal phenomenon—the arrival of the Inter-Tropical Front—for India provides the perfect geographical circumstances for the clash of two strongly contrasted air masses. Eastern Asia, as far north as Korea and Japan, also provides good conditions, lying as it does between the largest landmass of the world (Asia) and the largest ocean (the Pacific) each of which produces its characteristic air mass. These are true Monsoon Lands, of which the keynote is the suddenness of the seasonal changes of weather. Elsewhere the change is more gradual and the seasons slide into each other.

But it would be an error to assume that the rains, once established, necessarily remain. To the great sorrow of the Monsoon Lands there is too often a reversal to dry season conditions; such breaks in the monsoon may result in famine, the failure of crops and the starvation of millions. Also despite its reputation for regularity the monsoon is not always punctual; its onset may be delayed for weeks beyond the expected date (5th June, at Bombay, for example). The date of its departure, too, may be several weeks early or late. In general the retreat of the monsoon, by contrast with its arrival, is gradual; and the northerly air (Pc from Asia) slowly drives back the southerly stream (Tm) and the I.T.F. retreats southwards to its normal position near the equator, losing its frontal character as it goes and resuming its normal character of a convergence zone.

As the low-pressure convergence zone slowly retreats southwards down

the Bay of Bengal and the Arabian Sea, violent storms are often born, cradled in the trough, but moving, generally, northwards towards the Indian Coast. Having their origin at some distance from the Equator they feel the deflecting force of the earth's rotation powerfully enough for the winds to acquire a rotational swirl; they are true hurricanes or typhoons of the type described on page 156. Such storms also happen during the advance of the low-pressure trough preceding the onset of the South-West Monsoon. About a hundred of these storms occur each year, mainly in the months from April to June and again in October and November. As long as the monsoons are blowing steadily the conditions are unsuitable for their generation; the risk is very slight in February to March and again in July and August.

By January the low-pressure trough has reached its southerly limit and is bringing welcome rain to the dusty plains of Northern Australia. Thus the retreat of the monsoon gradually allows the establishment in India of the Cool Season (November to February) when the sun is in the other hemisphere and the mean temperature in the plains is in the upper sixties; it is the pleasantest time of the year. The prevailing wind at this season is north-west and may be regarded as Pc, though little of the true Pc air, bottled up in the basins of inland drainage in the heart of Asia, escapes over the barrier of the Himalayas. Weak frontal depressions, not unlike those of the Mediterranean winter, pass at this season south-eastwards along the foot of the Himalaya bringing some useful rain to the agricultural lands of the alluvial plains and some snow to the slopes of the Himalayas. During the winter, India is almost a self-contained climatic region walled in on three sides by lofty mountain ranges which act as gigantic wind-breaks, shielding her from the bitterly cold air generated in the huge winter anticyclone over the plateaux and basins of the heart of Asia, but in the summer she lies open to the monsoon from the Indian Ocean and the Bay of Bengal.

The Hot Deserts (Sahara, Kalahari, Australian, Peruvian)

The Saharan, Arabian, Australian and Kalahari deserts are simple examples of regions of permanent Tc air lying under the mid-latitude high-pressure cells. High-level air is steadily descending and being warmed adiabatically, becoming very hot and dry. It lies on top of air relatively cooled from below, especially over the oceans, and a well-marked inversion is found over the High-Pressure cells and the Trade Winds. Rain is virtually impossible under such conditions and only occurs with some invasion by moist airs from outside the true desert areas. On the rare occasions when it does fall the rain is generally torrential but short-lived; it causes considerable erosion of loose soils, bare of vegeta-

tion, and chokes the wadis with thick muddy torrents. Weather has a strongly-marked diurnal rhythm: scorching by day—cool, even chilly, at night. Convection effects by day are strong, but cause only whirling dust clouds and no rain.

More interesting and varied weather occurs round their margins where the sea, generally with a cold current, washes the desert shore. Powerful sea-breezes are set up along the coastal belts of the Kalahari, Atacama, Westralian, Mauritanian and Californian deserts. These bring in moist oceanic air, keeping the temperature down and producing fog, which, however, cannot penetrate far into the dry desert air inland.

Polewards and equatorwards the deserts pass into Mediterranean and Savanna climates respectively and partake a little of their characters, receiving winter rain (polar frontal) or summer rain (I.T.F.) respectively.

The Interior Deserts (Mojave, Turkestan, Gobi)

The cause of drought in these regions is rather different; it is not a direct result of a locally-generated dry air mass, but rather of inaccessibility to rain-bearing winds. In the American deserts, Patagonia and Mojave, the actual distance from the sea to windward is not great, but mountain barriers athwart the flow of rain-bearing Tm and Pm air rob them of their moisture and hold up the passage of fronts; the deserts here are rain-shadow deserts. In the great deserts of the heart of Asia it is a matter of sheer distance in summer when convection is vigorous and rain would fall if the air were moist enough. In winter the effect of the continental high pressure, generating a cold cushion of Pc air, effectively limits precipitation to a little fine snow. As between the two great land masses of the northern hemisphere, the extent of Pc air in the heart of North America is much smaller than in Eurasia and the high pressure is less intense. In the face of the feebleness of frontal depressions often find their way across the American continent by a route known as the Northern Circuit, which follows the line of the Great Lakes. The aridity, therefore, is less complete in North America than in Asia and winter weather is more varied, with warm waves (often of a Föhn or Chinook type) from one direction, and cold waves, often armed with fine powdery snow (blizzards), from the other.

Though the summers may be as hot as in the tropical deserts, which are in much lower latitudes, the winters are extremely cold because of their high latitude and the great loss of heat during the long winter night. Warmth is not imported from the oceans to compensate for this nearly continuous loss, as it is in maritime climates. It can be seen from Fig. 60, *a*, that Pc air extends as far south as the Himalaya in Asia and, though warmed as it goes, continues on to the I.T.F. as the North-East Monsoon.

In North America it often reaches nearly to the Gulf of Mexico and when it flows powerfully it is mitigated only a little in its rapid passage. In fact, the heavy cold air of the interior plains, hemmed in between the Appalachians on the east and the Rockies on the west, often oozes down to the shores of the Gulf and to Florida where it is an unwelcome intrusion into the sub-tropical horticulture of that region.

Steppe Climates

It will not have escaped notice that desert climates are arid for a variety of reasons and that each represents a degenerate variety of some wetter climate, such as Savanna, Mediterranean, monsoon or cool temperate—degenerating by failure of the rains from whatever source, to reach so far. Outside the tropics the semi-arid climates bordering the deserts and forming the transition zones of climate and vegetation are known as steppe. The scanty rainfall is enough for grass, but not enough for trees, and the term steppe really refers to the native grassy vegetation of these plains.

The greatest expanses are in Central Asia and the western part of the interior plains of North America (prairies). In winter they are cold and dry, the air is mainly Pc; in mid-summer they are hot and dry, the air, heated on the spot, has the qualities of Tc, and though convectional showers occur the rapid evaporation diminishes their usefulness. At the change-over of the seasons, especially in spring when the continental highs are weakening, frontal and convectional rain provide the wettest period. This is a suitable climate for cereals and the prairies and steppes have raised great quantities of wheat and barley. But the wisdom of ploughing up the natural sod and reaping a rich harvest from the stored fertility of centuries is now recognised as short-sighted folly. For wind erosion has raped the top soil and many a mid-western farmer has seen his neighbours' fields go flying eastwards in a dust-storm.

The Mediterranean Climate

On the western side of the continent between the latitudes of about 30° and 40° north and south, according to the season, lies the poleward limit of the high-pressure cells of Tm air near its source. In summer these regions therefore share the drought of the deserts that lie towards lower latitudes. But in winter as the high-pressure cells retreat equatorwards and as polar influences strengthen, colder air from higher latitudes (Pm or Pc) sends tongues to meet the TmW spreading out from the sub-tropical highs, and conditions are ripe either for frontogenesis or for orographic rain on west coasts.

Here, then, within narrow latitudinal limits and on western margins only, is a peculiar and characteristic type of climate with the unique

quality of having its rainy season in winter and its dry season in the height of summer. Summer weather is almost desert-like in its sun, heat, glare and drought; winter weather is almost British in its alternation of frontal and orographic rain with bright intervals coinciding with the passage of high pressures between depressions. But the rain is less frequent than in Britain and heavier when it does come. Another respect in which this winter differs from cyclonic climates further north is the general absence of winter anticyclones in Pc air. This type of climate is known as 'Mediterranean' because although it is typically 'Western Margin' it has a wide distribution round the shores of that great inland sea, whose west-to-east expanse extends the effective penetration of western influence a distance of some 2,000 miles into the land mass of the Old World where it has had a profound effect on the development of the early civilisations along its shores.

Outside the Mediterranean this type of climate recurs in the proper situation in small areas round Cape Town, Perth, Valparaiso and San Francisco. If hot dry summers and mild wet winters are to be regarded as definitive of the Mediterranean climate then these regions are paradoxically much more typical than the type area of the Mediterranean coastlands themselves. The reason is that the relationship of land and sea is relatively simple along a straight coastline, but the Mediterranean Sea has a deeply-indented coastline with three mountainous peninsulas, and many islands, interpenetrated by long gulfs and island-studded seas. Consequently the weather of the Mediterranean is subjected to more variable and complicated geographical influences. Furthermore there lies to the north the great reservoir of winter Pc air in central Europe and though it is hemmed in by the line of mountains from the Cantabrians and Pyrenees to the Caucasus it often escapes through gaps in the mountain ring, giving cold winds such as the Mistral down the Rhône corridor and the Bora from the cols at the head of the Adriatic sea. These outbursts of Pc air occur in the rear of travelling depressions; the southerly winds experienced in front of the depressions are warm, enervating and dry on the African side, where they come from the Sahara (the Khamsin and Sirocco), and humid when they have crossed the Mediterranean (Solano).

Summer winds, described in most textbooks as 'Trades,' actually blow from a westerly quarter and are really integral parts of the circulation of stable Tm round the sub-tropical cells. In the Eastern Mediterranean, influenced by the monsoon 'low' of south-west Asia, they are northerly (the Etesian winds), fresh and constant, and bring fine refreshing weather.

The peculiar regime of the Mediterranean climates is one to which not many plants are adapted. Grass does very badly unless it can be irrigated in summer; bushes and trees with their deeper roots systems are better fitted. The typical vegetation is evergreen forest of a rather scrubby type,

stone pines, cypress and laurel. But the Mediterranean lands have been occupied and cultivated so long that little remains of the natural vegetation, which has been replaced by a sort of horticulture dependant on tree crops (olives, citrus fruit and vines). Cereals, autumn sown and harvested early before the drought is really serious, are grown for home consumption rather than export. Long a land of peasant farmers, the Mediterranean basin still carries the stamp of small farming in which a high yield can be obtained only by constant attention; it is a land of difficulty, but a land of promise. It is also a favourite holiday climate, both for its mild winters and for its dazzling sunny summers whose heat can be mitigated by sea-bathing. The beach (plage, piaggio, praia, praha) is an essential feature of every seaside holiday centre.

The Temperate Climate

It is a pity that the word 'temperate' is so well established in this connection as to be fixed for good and all in the vocabulary of climatology. Actually the temperate climates experience the greatest extremes of heat and cold and the greatest variety of weather, for there meet in these latitudes winds from the sub-tropics and the poles along the Polar Front. It is true that the original heat of Tropical air is tempered during its passage to our latitudes and that the cold of Polar winds is moderated by the time it arrives in the temperate zone, but enough of the original quality of each of these strongly-contrasted air masses survives not only to subject us to sudden and considerable changes of temperature but also to create strong frontal disturbances. Weather assumes an importance that it rarely acquires in tropical climates; in fact the only reliable quality of our climate is its unreliability. Yet certain general types of weather can be recognised as recurrent and, on analysis, it is clear that they belong either to air masses or to fronts. The expectable frequency of each type and the frequency of change from one type to another would seem to be the most realistic manner of describing the climate of temperate latitudes and this has been done in the earlier chapters of this book.

In general, frequent change is characteristic of maritime climates where marine and continental influences meet; changes are less frequent in the interior of continents, though the contrasts are greater when they do occur. The difference in temperature between the invading masses of winter Pc that gives us our hardest black frosts and the Tm that displaces them causing the thaw is probably only about 20° F., but the difference between mild Gulf air (Tm) at Cincinnati and the bitter cold wave of Prairie air (Pc) that follows may be more like 50°. And when a long spell of stagnant Pc air at Calgary is broken by the adiabatically warmed 'Chinook' the temperature may rise 50° in a few hours.

But the lowest and the highest extremes of temperatures occur during long spells of settled weather when an anticyclone of Pc or Tc air has been long in residence and this can only happen in continental interiors. In a winter anticyclone the night temperature has fallen to -90° F. at Irkutsk and summer day maxima of 90° F. have occurred at the same place.

Thus there are two distinct types of 'temperate' climate. (1) Maritime on the westward (*i.e.* windward) side of continents in the belt of prevailing westerly winds. This we find in Western Europe, British Columbia and the Pacific north-west, southern Chile, Tasmania and New Zealand. Its characteristic weather is marked by frequent and rapid change without very great extremes and by the importance of fronts with their typical weather sequences. (2) Continental, met only in the hearts of the northern continents of Eurasia and North America, for the land masses of the southern hemisphere are too small in the appropriate latitudes to nourish Pc air masses. Its characteristic weather is marked by the persistence of air masses, each with its diurnal sequence of weather events, throughout a whole season (Pc in winter, air akin to Tc in summer). Non-periodic weather changes are relatively rare, but the contrast between the seasons is very great indeed.

But what about the east coasts of the two land masses of the Northern Hemisphere? To some extent they are intermediate between continental and maritime types, but the two land masses of North America and the Old World also differ among themselves. From Chesapeake Bay to the mouth of the St. Lawrence the Polar Front is often to be found on the weather map not far out to sea. The cold weather is often due to Pc air from the interior and even if this air has been partially modified to Pm by passage over the sea it is still cold for its latitude. At the other extreme warm humid air comes up from the south over the Gulf streams (TmW) and the two meet along the Atlantic Polar Front.

The temperature contrasts are therefore greater than on the west coast, though far less great than in the interior. On the other hand frontogenesis is favoured all the way up the coast from Florida to the St. Lawrence. So the eastern seaboard has much more variable weather than the interior.

In eastern Asia there are the monsoons. The winter weather is therefore strongly continental in character and cold right down to the coast; the Pacific Polar Front lies generally far out in the ocean; so weather changes are rare. In summer the South-East Monsoon (Tm) brings the influence of the ocean far into the land mass where convectional and orographic lift cause heavy rains, especially during the heat of the day. Thus the seasons are well-defined and clearly contrasted. Altogether the climate is much nearer the continental than the maritime model even in coastal situations. But there is this difference—wind. The continental interiors are regions of calm, being the centres of air masses, the monsoon coasts are windy since

the air masses cross them in passage from strong highs to deep lows, whose pressure differences are the very cause of the monsoons themselves.

Arctic and Antarctic

The regions of permanent snow and ice that surround the poles and, indeed, embrace the whole of the Antarctic continent, are incapable of supporting settlement by human beings, who visit them only as explorers or scientists. For them the weather is of great interest and often of vital import. Thanks to the increasing interest in polar exploration and the great strides recently made in equipment for travel and communication and even living through the terrible winter, our knowledge of the weather of those hostile regions is growing yearly.

The dominant air mass throughout the year is A or Pc and inversions of temperature with extreme surface cold are characteristic. But the salient feature of the regions is, of course, the contrast between the midnight sun of summer and the long dark winter night, six months long at the poles themselves.

Within the polar circles the summer sun is never high enough in the sky to be really powerful (47° above the horizon at its midday highest at the solstice) and its oblique rays are largely reflected from a surface mostly snow or ice-covered. So even in summer it is inhospitable.

Throughout the winter the surface, mainly snow or ice which must be considered as part of the solid lithosphere even if it is frozen ocean, continues to lose heat by radiation continually all through the twenty-four hours, day after day, right up to the sunrise (the equinox at the pole). The end of winter, not the middle as in most climates, is the coldest time of the year. The inversion of temperature characteristic of Pc air is deepest and the temperature is at its lowest during calm weather, but the numbing, paralysing cold of such times can be more tolerable than the wind, which stirs up the air and breaks down the inversion, actually causing the temperature to rise. These blizzards whip up clouds of fine dry penetrating snow and blow away the warmth from the body, so that it is impossible, without shelter, to maintain body temperature and death cannot be far away. Exposed skin is quickly frost-bitten, damp clothes freeze into armour plating and breath freezes into icicles on the face.

Blizzards are the result of two main causes. (1) The indraught into frontal depressions passing nearby in the 'Roaring Forties' of the Antarctic Ocean and along the Arctic and Polar fronts of the North Atlantic and North Pacific. (2) Katabatic effects, especially around the edges of the lofty ice-clad plateaux of Greenland and Antarctica. Being essentially the products of permanent relief they are strongly localised. Certain areas such as Adelie Land, where downcast air is funnelled between steep ice-

clad slopes as it rushes down from the high plateau of Antarctica, are notoriously the 'Home of the Blizzard.'

Around the permanent snow and ice in the northern hemisphere, in the barren lands of Canada and Siberia, lies the zone of the Tundra in which the snow melts for a time in summer exposing a vegetation of mosses, sedges and stunted shrubs in swampy muskeg soils saturated, except on steep slopes, with ice-cold water. The weather is mostly that of the ice-caps just described, being mainly under Pc air, though visited occasionally by depressions in the summer months. The long hours of summer daylight permit the growth of vegetation and on fine days, despite its low angle, the sun can raise the temperature of the ground and the air in contact with it into the sixties.

Mountain weather (e.g. the Alps)

It all depends, of course, on where the mountains are and how high they are; Chimborazo has a different climate from the Matterhorn and Kilimanjaro differs from Mt. St. Elias because they are in different parts of the atmospheric circulation. Nevertheless, there are some features of weather characteristic of mountains irrespective of their latitude or geographical position; they are the result of the distinguishing features of mountains, their height, steepness of slope and exposure.

The Alps are visited by thousands of British and other tourists every year, for ski-ing in winter and spring and for climbing or walking holidays in summer; they provide good examples of mountain climates and what is said here can be illustrated from them.

Because of the height the air is thin (atmospheric pressure at about 18,000 feet is only half that at sea-level) and very dry. Furthermore, in mountains like the Alps that lie under an anticyclone in winter, there is often a strongly-marked inversion of temperature. Under these conditions, cloud often fills the valleys up to the inversion level, above which the peaks and the ski-fields rise into clear air where the sun shines whitely in a dark-blue sky.

But the air is cold because, being thin, clear and dry, it is penetrated by the sun's rays without getting much warmed in the process. The general rule that air temperature decreases with height at a rate of about 3° per 1,000 feet is approximately true of mountains if it is applied to mean air temperatures, but there are considerable variations in the lapse rate according to the weather conditions. In any case the sensation of bodily warmth or cold is not always closely related to what the thermometer says. In fine weather the direct radiant heat from the sun is a welcome source of warmth by day and the effect of sitting in the sun on the café verandah at the top of the funicular or ski-lift is not unlike that of warming oneself

at a gas fire in an icy room—scorched shins and frozen back. Furthermore, the sun's rays at high altitude contain a high proportion of ultra-violet which is absorbed in the lower reaches of the atmosphere. These rays are very active in producing sunburn and browning of the skin. The unwary can get painfully sunburnt on exposed parts, though the air temperature is below freezing, and as one is also receiving an extra dose of solar radiation reflected from the white surface of the snow unexpected parts of the anatomy may later on be found to have suffered. Since the dryness of the air is a contributory factor liberal doses of cold cream or ointment are advisable to prevent drying and chapping of the skin.

Glare from the snow surface is trying and may cause painful snow-blindness, though everybody now wears tinted glasses, preferably with side-flaps (otherwise the glare from the sides prevents you seeing forward clearly through the glasses). Sun goggles are very good since they cut out the glare from all sides (but don't wear them all the day, unless you want to look like a circus clown with white circles round the eyes in a brick-red face).

Not all winter days are of this ideal type, a favourite with the advertisers of winter sports. Such conditions depend on the establishment of an anti-cyclone in Pc air and are commoner in the late winter and spring months than at Christmastime, when many a would-be skier has spent a whole week watching, through the hotel window, driving snow or, worse, rain from passing depressions. Warm Tm air may find its way into the Alps in winter and is likely to provoke a Föhn wind (see page 199) with the unpleasant consequences of thaw and danger of avalanches.

Summer weather in the Alps is very variable and is related to frequent changes of air mass and the passage of fronts nearby. Yet certain features may again be said to be characteristic of and peculiar to mountains. The chief of these is the tendency to a diurnal regularity of weather events whenever the weakening of extraneous weather influences permits local factors to exert full control. The processes may be said to be set in train by the diurnal variation of insolation. This influences temperature both directly and indirectly and temperature, in turn, generates a daily change of wind direction, down-valley at night, up-valley by day (see page 200). The ascent and descent of air, in their turn, affect the distribution of cloud which tends to rise during the morning and to sink again at night. This is one reason why mountain climbers generally make their assault of the high peaks very early in the morning and are down again before the cloud obscures the view or makes their return dangerous. Another reason is that the powerful sunshine after sunrise spoils the surface of the snow by thawing. This happens especially round dark-coloured objects such as stones, releasing the grip of frost on boulders and rock fragments which fall, constituting a real menace to climbers on steep and sunny slopes.

The storms that rage round the mountain tops are often electric and can be alarming. The ice-axe begins to fizzle and give off sparks and the crash of thunder, re-echoing among the peaks, is awe-inspiring. The influence of aspect reaches supreme importance in Alpine climates (see page 197) and there may be a month's difference between harvest season on north and south-facing sides of a valley. Glaciers descend further on north faces before the sun can effectively melt the ice, and, at comparable altitudes, cirques are ice-filled on the north side of a peak but ice-free on the south face.

Within the tropics, of course, this does not apply—the sun shines more or less vertically down on Kilimanjaro and Cotopaxi. In the Himalayas in summer, when most of the big peaks are attempted, the sun is approaching the vertical. The contrast between north and south faces is rather one of exposure to air streams such as the monsoons.

At really high altitudes the effects of diminished atmospheric pressure and oxygen lack become very noticeable. Oxygen supplies are turned on at 10,000 ft. in service aircraft and mountain sickness affects many passengers on the railways of the Andes and even some tourists on their way up to the Jungfrauoch in Switzerland. The exertion of climbing adds to the distressing feeling of breathlessness the further suffering of internal chilling from the rapid, panting intake of thin cold air into the lungs.

LOCAL INFLUENCES AND LOCAL WEATHER

The weather of a region may be described in terms of the prevailing air mass, the presence or absence of fronts and the controlling pressure distribution, but more facts than these may be needed to explain the weather of a particular locality within the region. A variety of local influences—such as the nature of the ground surface, the presence of water, the shape of the land, the composition of the atmosphere and sometimes several of these effects combined—make their contribution towards the realities of weather. Forecasts issued for the public cover large regions and cannot make allowances for local peculiarities except in the most general way. The local factor thus explains the difference between the weather forecast we hear on the radio and the weather we may later experience. Familiarity with local weather, therefore, enables us to make our own adjustment to the regional forecast. Beyond this, local influences are of interest, for more specialist purposes, to people like farmers, gardeners, builders, planners and doctors. Local factors assert themselves most often (though not exclusively) when the general weather influences are weak, which means, in effect, in relatively calm and clear (anticyclonic or col) conditions.

The Influence of the Ground Surface

Any townsman who has observed how snow or hoarfrost vanishes rapidly from dark asphalt surfaces but lingers on garden lawns will appreciate that different kinds of surface respond differently to the same amount of solar heating; likewise the countryman, who knows from experience that some soils are warm and others cold. From the meteorological point of view, soil consists of solid particles (which conduct heat relatively well), air (which is a poor conductor) and water (which conducts heat well but, having a high specific heat, responds only slowly to heat changes). Temperature conditions at and immediately above bare ground surfaces depend partly on the relative proportions of these constituents.

Soil Types

At one extreme, bare rock surfaces may warm up considerably in sunshine (especially if they are dark) but much heat is conducted downwards

and stored up; during the night this heat moves upwards and is released, maintaining the surface relatively warm. The brick-work, masonry, concrete and asphalt, etc., that make up the surfaces of built-up areas behave similarly and partly for this reason towns tend to be warmer than the countryside (walking in city streets in summer can be so trying to the feet).

A dry soil containing a good deal of air (a sandy soil) heats up much more strongly at the surface during the day precisely because the heat does not penetrate. A sandy beach, of course, shows this effect to perfection. On the other hand, this heat is quickly lost by outward radiation at night. A sandy soil therefore displays great diurnal extremes of temperature and is prone to low minima and to frost. This makes itself felt even at Stevenson screen level where minima over sandy soils may, on suitable nights, be 10° or 12° lower than at comparable stations on wetter or more compacted soil.

A soil containing much water (e.g. a clay soil in winter) shows the least diurnal variation of temperature due both to the high conductivity of the water and its slow reaction to heat change. A wet soil also has a constantly evaporating surface and, since evaporation lowers both surface and adjacent air temperature, tends consequently to be a cold soil. The wetting of any soil by rain or artificial means adds some of these characteristics.

Somewhere between these extreme types, a dry, well consolidated soil such as a medium loam has moderate thermal properties, being generally warm and free from sharp temperature fluctuations.

Snow Cover

The coldest surface of all is a snow cover. During the day something like 80 per cent (less for old dirty snow) of the incoming solar radiation is reflected away and there is little gain of heat. At night heat loss by long-wave radiation proceeds rapidly and very low minimum temperatures are recorded both on and above the surface. At the same time, since snow is about 90 per cent air it forms a most effective insulating blanket and temperatures at the ground surface beneath are kept at or only slightly below freezing. Because of the high reflectivity, a thick snow cover is slow to thaw but with a thin cover the radiation that does penetrate can more quickly warm the underlying surface (or buried objects like stones and twigs) which in turn warms and melts the snow in contact. Some of it, of course, disappears by sublimation, especially in windy conditions.

Vegetation Cover

Vegetation-clad ground is no longer directly exposed to the elements. To a greater or lesser extent, depending on its structure and density, a



PLATE XVII. Local weather: a solitary hill cumulus sits above the highest hill. Lleyn Peninsula from near Harlech

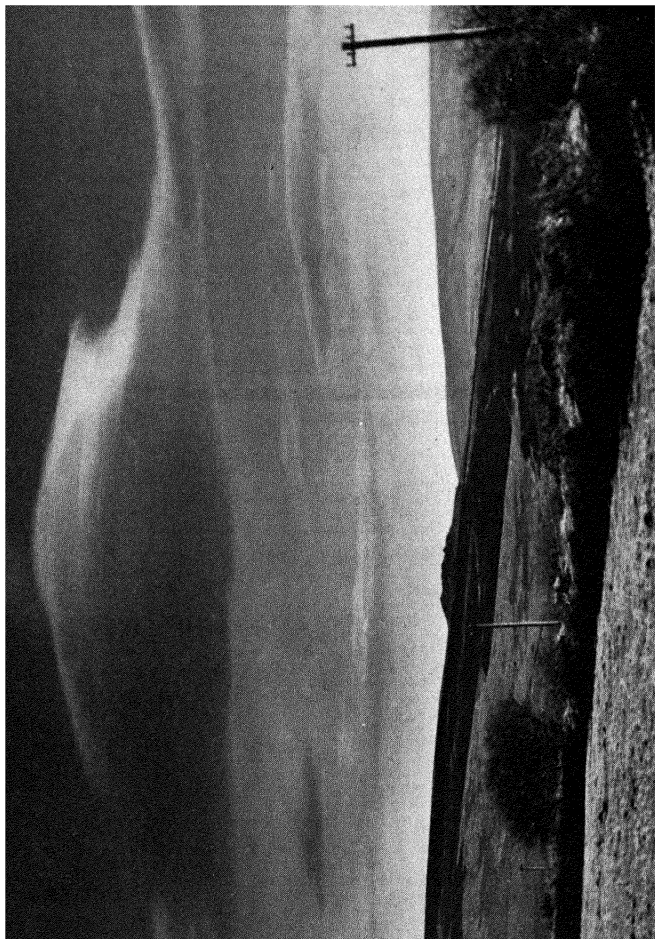


PLATE XVIII. Local weather: wave clouds over Dartmoor suggest the shape of the land-forms producing them

vegetation cover protects the underlying surface from the temperature extremes that may be observed over bare ground. It is now the upper surface of the vegetation that directly receives solar radiation and emits terrestrial radiation and it is here that the extremes occur. In woods and forests, the tree crowns high above our heads warm up most during the day leaving the ground below shaded and cool and we make grateful use of this temperature inversion for our afternoon stroll. At night, temperature differences within the forest are slight. The tree crowns may be the coldest part but, with less dense woods, the air chilled and made heavy by contact with the cold leaves can sink to give a temperature minimum near the ground.

All vegetation with horizontally-spreading leaf canopies, from garden flowers to deciduous trees, show some of these characteristics. On the other hand, more vertical structures—like standing ripe corn—allow the warmth of the sun to penetrate more effectively during the day. This, plus the fact that air is held trapped between the stalks and not constantly changed by turbulence, means that high temperatures can develop in and just above the corn. At night, the cooling effect is also distributed downwards within the plant.

In the case of grass, however, day temperatures are kept comparatively low by the cooling due to evaporation of transpired moisture. At night the stagnant air between the grass stalks is cooled by this means as well as by radiation. For this reason low temperatures and frost are especially likely over grassy surfaces and more so over long grass than short. Heavy dew is most common on long grass. The combination of low temperature and high humidity in and above long grass may also lead to the appearance of shallow *ground fog*, often only knee-high, during suitable (calm clear) nights; this small-scale weather effect is seen especially over riverside meadows in early evening and is often indicative of more general radiation fog later in the night (Plate XX).

Micro-Meteorology

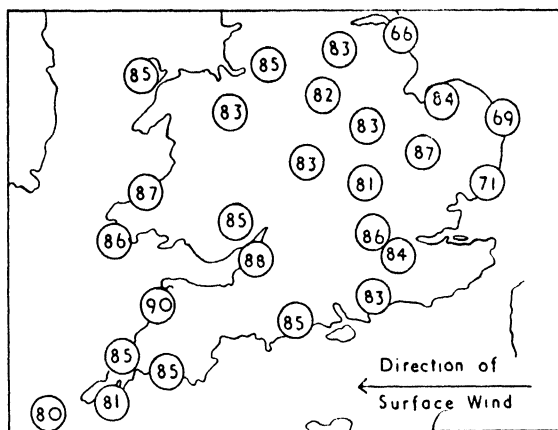
These are but a few examples of meteorological variations within small spaces—like the air layer near the ground—which may be of high significance, for example, for problems of agriculture. Their study is known as *micro-meteorology*, a growing subject that deserves a book to itself (and indeed more than one has already been written, to which the interested reader is referred). For our present purpose these micro-influences are not only interesting in themselves but also affect the general weather, particularly by providing some of the favoured heat sources—bare rock surfaces, tarmac roads, built-up areas generally, fields of ripe wheat—for convectional activity.

Coastal Weather

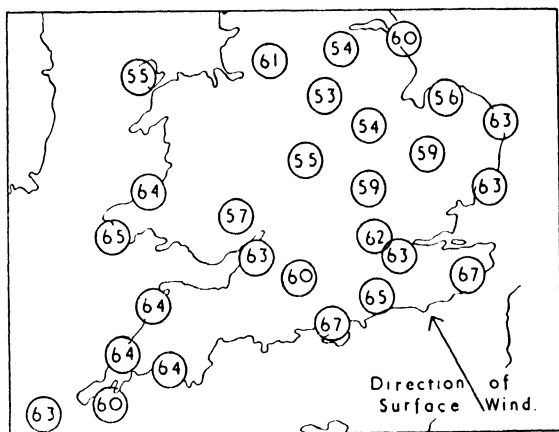
Meteorologically speaking, the surfaces most contrasted in their properties and effects are land and water, for reasons fully discussed in Chapter 3. Over the open sea, where the diurnal temperature change is negligible, there is a complete absence of all those diurnal variations—of relative humidity, of wind and turbulence, of convectional activity—that are so characteristic of inland districts. At sea there are no orographic influences as there are inland and the frictional resistance of the surface to the wind is very much less. Fundamental differences in weather and climate between land and sea result from these contrasts, as we have seen, but, in the present context, special interest attaches to coastal areas which are transitional between land and sea and must show in their local weather some of the characteristics of both.

While in detail much depends on the lie of the coast with regard to prevailing winds, in general our coastal districts share to some extent the equable character of marine weather. During the day they are cooler than inland especially in late spring and early summer (when sea temperatures are rising only slowly compared with day-time inland temperatures) but may be slightly warmer in late autumn and early winter (when the fall in sea temperature lags behind that of the land). At night coastal areas are warmer than inland at all times of year and especially in autumn and winter. Two examples of the effect of the neighbouring sea on coastal temperatures (Fig. 62) speak for themselves. It is worth remembering that the North Sea is always cooler than the English Channel and our western seas.

In climatological language, coastal stations have a smaller *range* of temperature, both diurnal and annual, than those inland. This is reflected in a smaller likelihood of frost and snow. As we should expect, humidities tend to be higher on windward coasts than inland. Winds off the sea blow more strongly and more steadily than further inland, as every holiday-maker knows, and the worst gales are experienced on exposed coasts. In unstable air, windward coasts enjoy no evening clearance of shower clouds. Fog in coastal districts is usually advection fog from the sea (Plate VII) and it may well be slow to shift. Radiation fog cannot form over the coast but occasionally drifts there from inland. Various orographic effects may contribute to the weather of hilly coasts. High cliffs cause forced ascent. With moist stable air, this may result in what is graphically known as *upslope fog*, over the high ground: Beachy Head has a reputation for fog of this type. In other conditions, cliffs may squeeze orographic drizzle from stratocumulus clouds or trigger off instability showers or even thunderstorms. Many of these coastal effects can be found on the synoptic charts accompanying Chapter 5.



(a)



(b)

Fig. 62. (a) Maximum temperatures in southern England, 23rd August, 1955. (b) Minimum temperatures in southern England, 12th August, 1953

Land and Sea Breezes

These winds, the most characteristic local weather feature of coastal areas, are parts of thermal mechanisms exactly similar to those associated with thermal lows and highs. They occur only under quiet general conditions allowing a good temperature contrast to develop between adjacent land and sea strips. On calm sunny days, the land warms up more strongly than the sea and, in effect, an elongated thermal low develops above it, with a complementary high out to sea. In the resulting circulation, air moves from land to sea at some height, and there is a return flow from

sea to land nearer the surface (the *sea breeze*). At night, pressure is higher over the cooled land, the circulation is in reverse and the low-level flow is from land to sea (the *land breeze*). Yachtsmen in coastal waters and even inland waters like the Norfolk Broads know these breezes well. Such winds are generally too transient to be adjusted to the Coriolis force and are usually more or less onshore or offshore in their directions.

The sea breeze sets in only when the land is sufficiently warmer than the sea (an excess of only a few degrees F. may be necessary). This happens regularly on tropical coasts where sea breezes are all-year phenomena but in temperate latitudes they are confined almost entirely to summer. The breeze appears late morning and fades late afternoon or evening. In temperate regions sea breezes are generally light (up to 10 m.p.h., occasionally 15) and shallow (not felt above 500 or sometimes 1,000 feet) and may penetrate a few to 20 or 30 miles inland (depending on the relief). Tropical sea breezes, however, are stronger (often over 20 m.p.h.) and deeper (up to 4,000 feet) and may be felt twice as far inland. Where there is an existing light wind, the sea breeze reinforces it if it is onshore or more or less cancels it if it is offshore.

The land breeze is a weaker affair altogether because the temperature contrasts are less at night (unless it is reinforced by other effects). It is shallower, rarely reaching more than 1,000 feet even in the Tropics, and extends only a few miles out to sea.

The most important effect of the sea breeze is to moderate what would otherwise be high afternoon temperatures. This is not necessarily welcomed by our own seaside visitors, who are not normally overburdened with warmth, but in low latitudes it plays a large part in maintaining tolerable conditions. Sea breezes off the relatively cool North Sea bring a noticeable freshness to our east coasts and probably help to explain why Skegness is so bracing. Off our southern and western seas, sea breezes have less effect. The transport of cool moist air over warmed land surfaces encourages convection, indicated often by a line of stationary cumulus over the coast. Under other circumstances, sea mist or fog may be carried in on the breeze. Malta, which is elongated in the north-west-south-east direction, characteristically develops two sea breezes on summer days, one from south-west and the other from north-east: these meet at a convergence line along the spine of the island which is marked again by a line of cumulus. The Florida peninsula also has a double sea breeze, from east and west, and the resulting convergence is thought to explain why this area has more thunderstorms than any other part of the United States.

The Effect of Smaller Water Bodies

Smaller water bodies show some of these effects on a reduced scale, although, of course, the shallower the water the less influence it has. In-

land seas and large lakes modify the weather of their shores and often develop their own systems of onshore and offshore breezes. Rivers are too small to produce local winds but, under calm conditions, influence the temperature and humidity of the immediately adjacent air. An example of this effect is illustrated in Fig. 63 which shows the results of two temperature measuring journeys across the Thames and its neighbouring meadows near Reading at different times of the day.

Weather in Hilly and Mountainous Regions

The shape of the land also contributes to the creation of local weather variations partly by the sheer mechanical effect of relief on the flow of air and partly through the temperature contrasts that so easily develop in terrain of rapidly changing aspect and elevation. Naturally these local effects are best expressed in real mountain country but they are quite well represented among the gentler contours of, say, southern England.

Aspect and Warmth

Aspect is perhaps the most obvious local factor. A cat carefully selects the foot of the sunlit wall for its afternoon nap. A field with a slope to the south is likely, other things being equal, to be the warmest on the farm. Alpine valleys lying roughly east-west have a south-facing side which is well blessed with sunshine and a north-facing side which may be in perpetual shadow for much of the winter. Not only does this unequal endowment with sunshine result in strong day-time temperature contrasts (which lead to other local effects), but the entire life and economy of the valley are adjusted to it. The farms and villages keep to the sunny side while the shady side is left under forest. In north-south valleys the distribution of sunshine is less unfair.

Relief and Airflow

When air moves over a large obstacle of any kind, its flow is not smooth but ruffled by vertical whirls or eddies which are especially persistent on the windward and leeward sides. Such eddies develop, for example, in windy weather, near large buildings such as aircraft hangars (and thereby increase the problems of airfield planning). They occur too on the flanks of hills and cliffs, and more than one small boy on the beach has released his kite or model aeroplane into the air only to see it crash down inexplicably, time and time again, at the foot of the cliff.

In addition to local eddying there is a general inducement to the uplift of air in hill and mountain country to which we have already referred as

NORTH

SOUTH

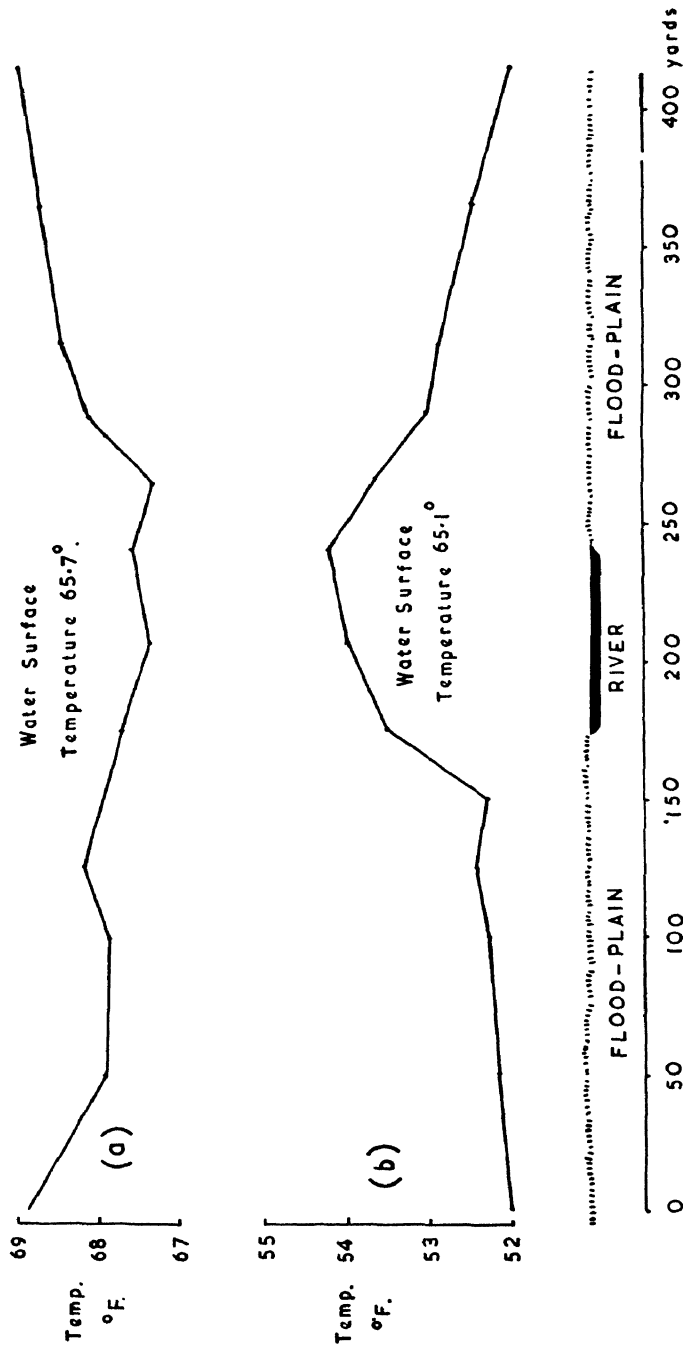


Fig. 63. The effect of a river on air temperature. (a) In the afternoon. (b) Late in the evening

forced or orographic ascent. Glider pilots make use of these up-currents (*slope soaring*) just as seagulls do above the cliffs. If the rising air is moist, cloud readily forms. Thermal influences are at work here, too, because mountain slopes, especially if south-facing, may warm up strongly in the sun and act as heat sources for convection (Plate XVII). Mountain cumulus may develop very rapidly from mere wisps and add to the hazards of mountaineering. If the air is initially unstable, cumulonimbus giving heavy showers and thunderstorms often form during afternoon and evening. With stable air, uplift is limited and upslope fog or a thin capping of cloud are the likely results.

Sometimes a lone cloud seems to float as if attached to the mountain top. It is aptly termed a *banner cloud*. It occurs when moist stable air is forcibly lifted and cooled to condensation point at about the height of the mountain top; further downwind the air sinks again, becomes warmed and no longer saturated, and here the banner ends. Although the cloud itself is stationary there is constant movement of air throughout it. Well-known banner clouds include the Levanter cloud above Gibraltar, the so-called 'Table-Cloth' of Table Mountain and the much-photographed banner of the Matterhorn.

Such orographic disturbance of air flow may be felt at elevations many times that of the high ground causing it. Glider pilots have provided ample evidence of this but the effect becomes clearly visible to all in the beautiful *lenticular* (lens-shaped) clouds of medium levels (Plate XVIII). These *hill-wave clouds* are like banner clouds but form in moist stable air well above the summit. They are represented even at high levels: cirrus caused by hill waves has a fibrous or striated appearance, often with a gently arched form highly suggestive of its origin. According to F. H. Ludlam, most of the fair-weather cirrus we see is orographic, caused by hills as modest in elevation as the Cotswolds or the Chilterns.

Under similar conditions, but with stronger winds, a whole train of standing waves may form in the lee of prominent mountains. These *lee waves* have been explored by glider pilots and are often marked by a succession of regularly spaced lenticular clouds. The best example in this country has been described by Professor G. Manley and is produced by Cross Fell, which overlooks the Eden Valley in a steep escarpment. With a strong north-east wind (the 'Helm Wind') a hill-wave cloud (the 'Helm Cloud') forms above Cross Fell (but sometimes sits on the summit as a banner cloud) and anything from one to five lee waves, each capped by a cloud (or 'bar') appear over the valley.

The Föhn Wind

The *Föhn* is the name given in the Alps to air that has been drawn over the mountains in response to the prevailing pressure pattern, has deposited

most of its moisture as orographic rainfall on the windward side, and now descends, dry and subjected to considerable adiabatic warming, on the leeward side. We can see with the help of Fig. 64 that when near-saturated air is forced up by high mountains, it cools mainly at the Saturated Adiabatic Lapse Rate: but when it descends on the lee side, now bereft of most of its moisture and no longer saturated, it warms at the Dry Adiabatic Lapse Rate, *i.e.* about twice as quickly. The air is now warmer than before its ascent (the difference being the latent heat of condensation of the original moisture).

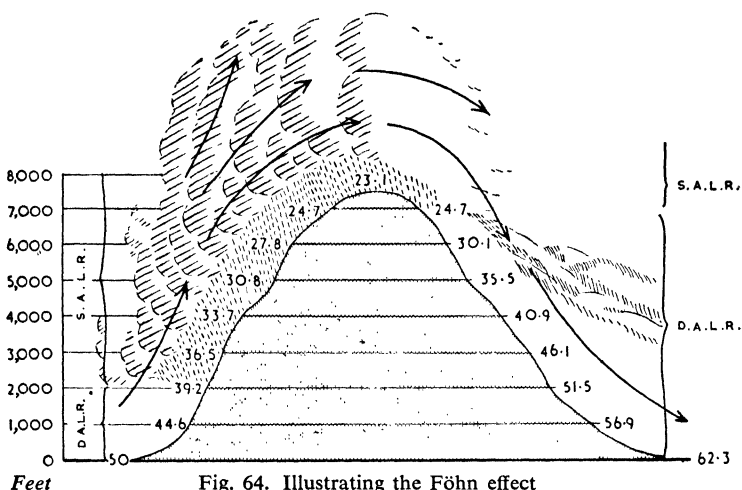


Fig. 64. Illustrating the Föhn effect

The sudden arrival of a strong Föhn is a remarkable phenomenon. It may raise the temperature in a valley by 40° or 50° F. almost abruptly and can clear snow by melting and sublimation within a day: it greatly increases the danger of avalanches. In this extreme form the Föhn probably represents initially warm air further warmed by this orographic process. Thus in the Alps the South Föhn (*i.e.* warm Mediterranean air) is distinguished from the North Föhn (cold Central European air which even after descent may be only a little warmer than the air it displaces). The Föhn is really a lee eddy (warmed air would not descend to replace colder air unless it were part of a forced circulation). The *Chinook* ('snow-eater') of the Rockies and the *Zonda* of the Andes are similar winds.

Mountain and Valley Breezes

Under quiet conditions, thermal circulations analogous to land and sea breezes over coasts occur in mountain country due to local heating and cooling. During the day the higher slopes warm up rapidly in the sun and

the adjacent air becomes warm, much warmer than the air at the same level high above the neighbouring plains. There is a gradual ascent of air in contact with the sunlit slopes, and fresh air is drawn in from the plains to give a general up-valley wind. A return wind aloft completes the circulation. Within the valley, air rises up the sides and there is a compensating down-current above the valley floor. Large-scale and small-scale effects combine (see Fig. 65) to produce an upslope wind—known as a *valley breeze* or *anabatic wind*—which lasts most of the day.

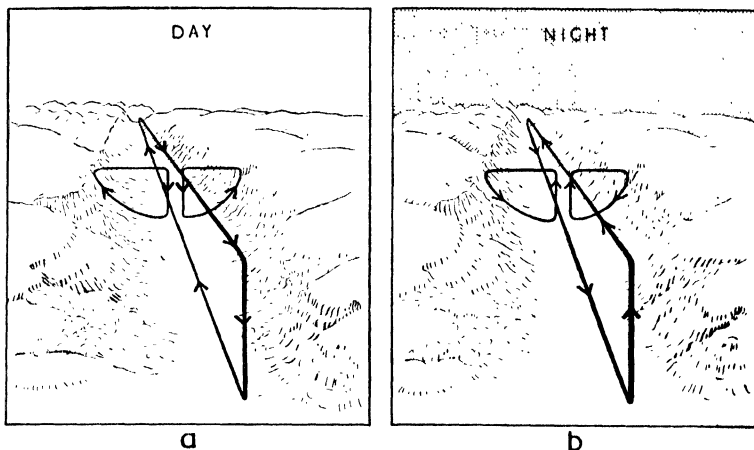


Fig. 65. Mountain and valley breezes. (a) Day-time. (b) Night-time. The arrowed lines indicate the longitudinal and lateral components of the circulations

At night the system operates in reverse. The upper slopes cool first and the air chilled by contact with them drifts down from the valley sides to the floor and generally down-valley to the neighbouring plains. Lateral and longitudinal flows together give a downslope wind which is called a *mountain breeze* or a *katabatic wind*. Stronger in general than the valley breeze, even so the nocturnal wind rarely exceeds about 10 m.p.h. High mountains are not needed for the creation of katabatics, they develop wherever slopes are appreciable. Well developed examples have been studied at the foot of the Chiltern slopes, and they are often conveniently observed by the drift of smoke from bonfires or chimneys. A special case of the mountain wind is found only in Alpine country. Where snowfields and glaciers cover the higher slopes there is a constant supply of cold heavy air to feed the down-valley wind. This so-called *glacier wind* reinforces the normal mountain breeze at night and slides down under the valley breeze during the day.

These periodic winds are not to be confused with the funnelling or canalisation of the general wind through narrow valleys (the same effect,

of course, occurs in city streets). The fierce cold north wind that sometimes blows down the Rhône Valley—the *Mistral*—is basically a funnelled wind of this kind.

Frost Hollows

Katabatic winds inevitably result in the accumulation of cold air over low ground, any adiabatic warming during descent being far outweighed by further cooling due to radiation and contact with cold surfaces. All valleys are vulnerable to this cold air drainage, some more than others, depending on their size, orientation and other factors, and many have acquired reputations as *frost hollows*. Probably the coldest known spot in Europe is a deep limestone sink-hole in the Austrian Alps—the *Gstettneralm*—on whose floor a minimum temperature of nearly -60° F. has been registered. On one occasion a temperature of -20° F. was recorded on the floor while 29° F. was recorded on the rim some 500 feet higher up, a local temperature inversion of 49° F. Such an inversion is the norm rather than the exception for the vegetation is adapted to it: only a few hardy grasses and plants of Arctic type grow on the valley floor, stunted pines are found on the middle slopes and a normal pine forest around the rim. The 'cold pole' of the earth seems to be a frost hollow in Siberia (Oimekon) where -94° F. has been reported.

Nearer home, E. L. Hawke has investigated temperatures in a narrow constricted dry valley in the Chiltern hills near Rickmansworth. Here no month is quite immune from air frost and a minimum temperature just below 0° F. has been recorded. At the same time some high maximum temperatures are on record, for the same configuration that makes a valley a frost pocket on clear nights may also make it a sun-trap on clear days (though this would not apply to a valley opening to the north). In Mr. Hawke's valley, the diurnal temperature ranges, which may exceed 40° F. and even reach 50° F., may be the largest in the British Isles. Less extreme examples are numerous in areas of well-defined (though not necessarily great) relief. Night temperatures at valley floor sites are commonly 6° , sometimes 10° or 12° , lower than those at hill-top or plateau stations not far away. To farmers and market-gardeners, such variations may mean the difference between suffering frost damage and escaping it. Nor must it be forgotten that the colder the air the nearer it is to saturation, and valleys are therefore especially liable to become enveloped in local mists and fogs which do not affect the higher ground (Plate XIX).

Relief, Rainfall, and 'Rain Shadow'

On the windward slopes of high ground the orographic effect is always added to whatever other rain-giving mechanism may be in operation.

Even stable air may be induced to part with a little moisture (orographic drizzle) from stratiform cloud on forced ascent. With conditionally or potentially unstable air, mountains provide the stimulus to the release of often heavy precipitation. In the case of fronts, orographic uplift reinforces frontal uplift and, moreover, mountain ranges tend to hold back the passage of fronts so that considerable and prolonged rain falls on the windward slopes. The heaviest rainfall occurs where warm, moisture-laden air is forced up over mountains (as in the summer monsoon of India). In this country, the most phenomenal falls—like the Exmoor downpour of 15th August, 1952—have been partly orographic in origin. The proportion of precipitation that falls as snow and the duration of snow are, of course, greater over high ground.

On the lee side of mountains there is a relative deficiency of rainfall, a 'rain shadow.' The air has lost much of its moisture on the windward side and is now descending and becoming effectively drier (the Föhn effect). Even frontal rain may be interrupted on the lee side and there is a temporary clearance of cloud. Thus it is that the distribution of rainfall largely reflects the relief of the ground, as a glance at a rainfall map of the British Isles will confirm (see also Fig. 66).

While the pattern of increased rain on the windward and 'rain shadow' on the leeward side applies to substantial highland masses, it is otherwise with low hills and rolling country. With these smaller scale features the lee side is the wetter. The rain is in effect blown over the hill crests by the stronger winds and dropped in the quieter air just on the lee side. This applies even more in the case of snow which, as is well known, piles up in drifts on the lee side.

Town Weather

The townsman, no less than those who live on hill-tops or valley floors or by the sea, experiences distinctive local weather. It could hardly be otherwise, for in the town artificial surfaces replace field and forest, an agglomeration of buildings obstructs and diverts the wind and atmospheric pollution significantly alters the composition of the atmosphere. A walk through the town often reveals a constantly varying and always interesting micro-meteorological pattern. We see this in the lingering of snow on north-facing roofs and the persistence of hoar-frost in the shadows of buildings. By turning the corner from a street lying parallel with the path of the wind to one at right angles to it, we pass suddenly from near-gale to comparative calm. Sunniness, temperature and humidity vary in street, square and park.

The heat-conserving properties of urban surfaces have already received mention. The artificial heating of buildings and (see later) the smoke pall help to keep the town warmer than the nearby countryside (by as much as

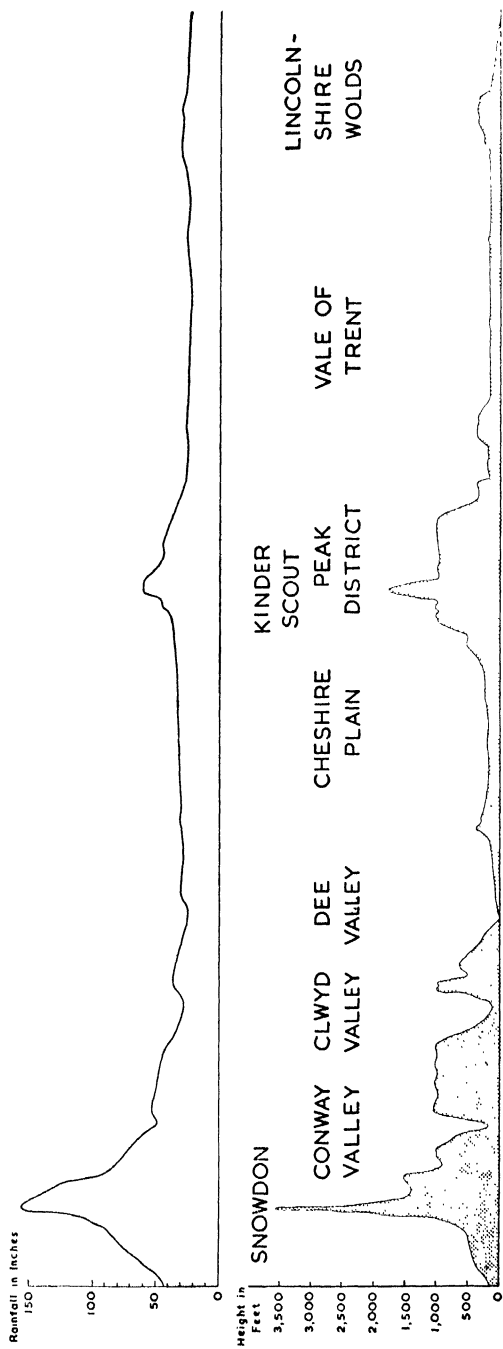


Fig. 66. Relief and rainfall profiles across England and Wales

8° or 10° F. on clear radiation nights). During the day, a town is a collection of heat sources, encouraging convection. Isolated cumulus clouds are sometimes seen sitting motionless above power stations, factories, or blast furnaces (Fig. 67) and examples have been reported of large clouds building up over urban areas while the surrounding country remains cloud-free. Occasionally it seems that the warmth of the town is capable of setting off the thunderstorm mechanism and unleashing an 'urban rainstorm.'

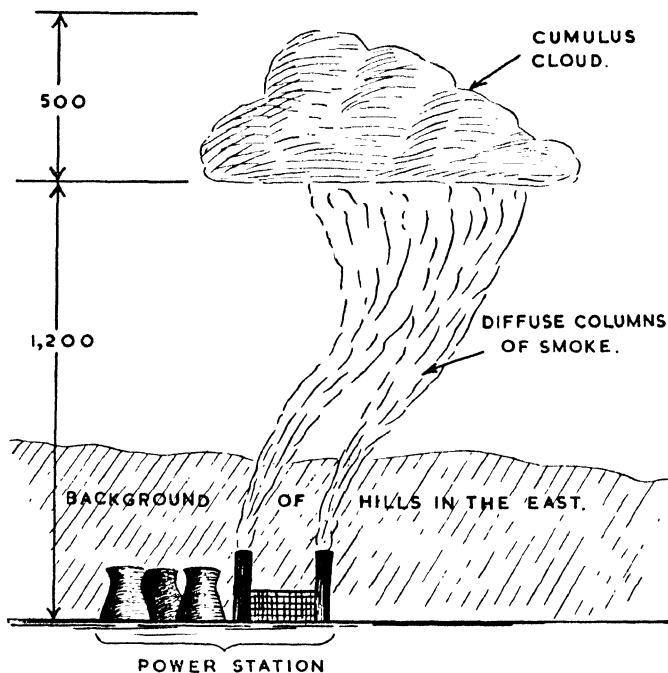


Fig. 67. Cumulus cloud above Brimsdown Power Station

Atmospheric Pollution

The most striking aspect of town weather is the polluted atmosphere which is the penalty we pay for our habit of burning fuel inefficiently. Atmospheric pollution consists of solid matter, part of it relatively heavy (ash and grit) and quickly deposited, and the rest sooty and tarry specks which remain suspended in the air: these, together with certain gases, make up what we call *smoke*. The gases include sulphur dioxide (the most important), carbon monoxide and carbon dioxide. Industrial chimneys, railways and (especially) power stations eject most of the grit and ash, industry and railways together produce more than half the smoke,

though domestic grates yield twice as much smoke per ton of coal burnt (and also discharge it at low levels). All chimneys are responsible for sulphur dioxide but industry produces twice as much as domestic fires: electric power stations emit rather more than all domestic grates put together. Motor vehicles contribute much carbon monoxide and carbon dioxide at street level.

Under normally windy conditions this atmospheric waste is diffused upwards by turbulence and removed by the stronger winds at height. But when the air is stagnant the pollution remains in increasing concentration. This has many implications. Those concerned with health or cost to the community we leave to Chapter 11. On the meteorological side, the smoke cover helps to maintain high temperatures at night by cutting down heat loss by radiation, but this is the only benefit it confers. It also cuts down light and sunshine (Central London has less than half the mid-winter sunshine of nearby rural areas) and, worst of all, it encourages 'smog' (Plate XIV).

'Smog'

A high concentration of smoke will of itself reduce visibility and compel the switching on of lights at midday. But this occurs in precisely those anticyclonic (or col) conditions, with the accompanying temperature inversion suppressing upward movement, that are most ripe for the development of natural radiation fog. Suspended smoke particles provide an abundance of condensation nuclei. The dark fog, once formed, defies penetration by the weak winter sun and may persist for days. The notorious London fog of December 1952 lasted four days: it was quite shallow, under a low inversion, and Box Hill and similar heights exceeding 500 feet stood up clear above it. Smog is thus an unholy combination of a water and a smoke fog and is far dirtier, more persistent and much less tolerable than natural country fogs.

Fog density often varies locally within large towns. The warmer (and therefore drier) central districts may be less foggy than the suburbs. Many a bus journey across London has illustrated this. Sometimes the fog is warmed out at street level, leaving a characteristic *high fog* above the rooftops.

Different towns present these features in different degrees. Other local factors may be involved and valley-floor towns differ from hill-top towns or coastal resorts. Aspect may become important: in many Alpine towns, in Bergen, at the head of its deep fiord, and, for different reasons, in New York, parts of the town see little of the sun during the winter. Residential towns must be distinguished from industrial towns (especially by the much-reduced pollution) and in 'garden cities' perhaps we are nearest the natural weather of the locality.

PROBLEMS OF WEATHER FORECASTING

It is not the aim of this chapter to teach weather forecasting. Many months of training are required to produce a professional forecaster and many years of experience to make him a good one. On the other hand, the amateur can sometimes achieve fair success at short-period forecasting by keeping a weather eye open and a few principles in mind. This may be useful and it is good fun, if nothing else. Some remarks later in this chapter are designed to encourage it. But, contrary to some often vocal but not very well informed opinion, the amateur, with his restricted view of the weather, is less likely to produce consistently good forecasts than the official forecaster with his synoptic view and his tools of the trade. At another extreme, some people assume that the forecasting problem is one capable of exact solution and that if the forecast goes wrong—as it sometimes does—that the forecaster must be incompetent and the organisation inefficient. The main aim of this chapter is therefore to give some idea of the problems facing the forecaster and of what to expect from his forecasts.

As a great meteorologist, Sir Napier Shaw, pointed out years ago, most scientists who make predictions do so on the outcome of carefully controlled laboratory experiments, but the poor forecaster is expected to predict the results of vast atmospheric experiments over which he has no control and whose workings he only dimly understands. It is not long since forecasting was entirely empirical, a matter of rough-and-ready rules based on experience. With the impetus of two world wars and the new knowledge provided by advances in observational techniques, some guiding principles have now been added to rule-of-thumb methods.

It is sometimes said that forecasting is an art not a science. This is another way of saying that a good forecaster draws on hidden reserves of experience and 'feel' to make decisions for which the scientific justification is not obvious. The subjective element in forecasting, which still looms large, may never be dispensed with entirely. It is unlikely that weather will ever be mathematically predictable as are, for example, the motions of the stars and planets. In the atmosphere there are too many influences at work, some perhaps far away in space or time. Even the most modern techniques can be only approximate.

It is unfortunately also true that forecasting is especially difficult in the British Isles because our position is marginal in so many senses. Stuck between continent and ocean in the restless middle latitudes, we are a battleground for so many invading air masses, and disputed territory for various pressure systems. A good share of local influences—geological, coastal, orographic, urban—helps to complicate the issue. The most confident forecasting claims are made by meteorologists in the wide uniform spaces of the continental interiors. The British forecaster has learnt to be cautious.

Understanding the Forecast

In our present state of knowledge, the official forecast is not a prophecy but a statement of the forecaster's estimate of future developments. We get the best out of the forecast if we understand it properly. As the distinguished meteorologist Sir David Brunt has written, 'Efficient weather forecasting is in fact a partnership between the official forecaster and the recipient . . .' The confidence that the forecaster feels in his own estimate is implied in his choice of words. In a clear-cut situation the forecast is definite ('The fine weather will continue . . .'). At other times there is a hesitancy ('. . . a chance of scattered showers') which is an honest way of indicating a low degree of confidence in an uncertain situation.

Most terms used in the official forecasts have their everyday significance but some have definite meanings and are carefully chosen. 'Showers,' for example (brief spells of precipitation with clearances between) mean something different from 'occasional rain' (when skies remain overcast throughout): the first we associate with unstable Pm air, the second with weak fronts or thick stratocumulus cloud. Sometimes various descriptive terms are used for expected temperatures which relate them to the normal for the time of year. Thus 'very hot' in summer means more than 20° F. above normal, 'cold' means more than 15° below. These and other terms are explained fully in the useful Meteorological Office booklet *Your Weather Service*.

Forecasts are the more reliable the shorter the time ahead and the more limited the area covered. Short-distance route forecasts for aircraft achieve a high degree of success. Local area forecasts (within 25 miles radius of the station) must be more generalised and may be upset here and there by small-scale influences. With regional forecasts it becomes more difficult to satisfy everybody everywhere.

In general, 12-hour forecasts are pretty accurate and 24-hour forecasts reasonably so. With more extended periods accuracy falls off sharply. In a forecast chart (prebaratic) for 24 hours ahead, such as appear in some morning papers, the average error in the placing of fronts is about 115 miles, which means a few hours in time. But 12 hours beforehand, a fore-

caster will often be prepared to time the arrival of a front within an hour or two. Some weather events cannot be forecast in any detail. A forecaster may be quite confident that frequent showers will occur over a given area but he cannot be expected to foretell the exact time and location of each shower. Every now and again, an unexpected development wrecks even a short-range forecast. Our psychological make-up seems to be such that we remember the occasional spectacular failure and are slow to forgive, especially if it affects us personally or the public at large on an important occasion.

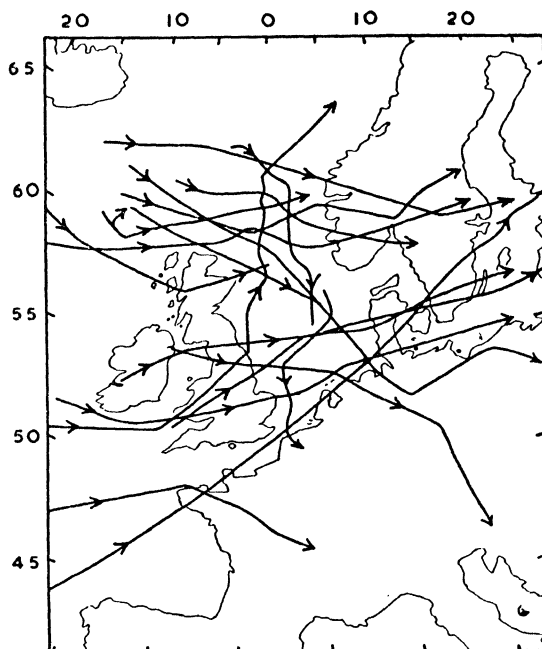


Fig. 68. The tracks of depressions over Western Europe during January, 1956

Forecasting from Surface Charts

Forecasting from the surface chart, which is still the main synoptic tool, is largely a matter of *extrapolation*. This assumes that weather systems will continue to move as they have been moving in the recent past, unless there is good reason to think otherwise. This method, the earliest form of synoptic forecasting, is still the basic one, especially in middle latitudes. Here our thinking is guided by the Norwegian model of frontal behaviour. Experience and theory have together evolved a number of general 'rules,' e.g. that young and mature frontal depressions move in the direction of

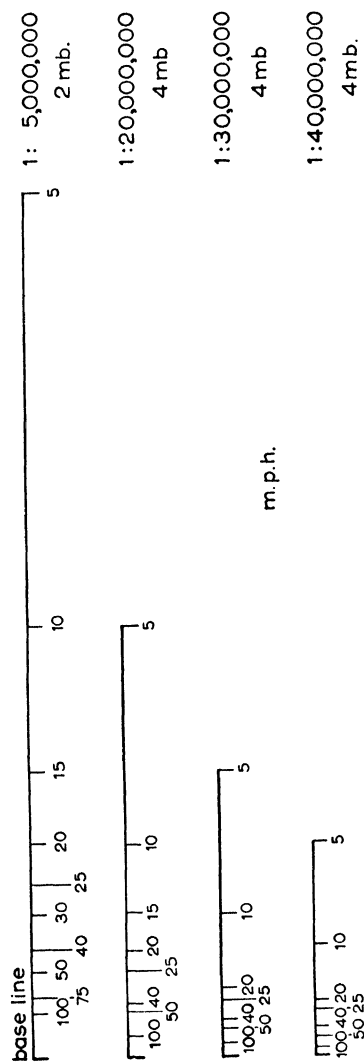


Fig. 69. Geostrophic wind scales (correct at 55°N) for various scales of charts used in the Daily Weather Report

the warm sector isobars, or that older lows slow up after occlusion. Unfortunately, not all depressions behave according to the precepts laid down in textbooks. No techniques are applied slavishly. The forecaster must always keep a hawk eye on the possibilities of new developments.

A large part of the forecasting problem concerns the travel of fronts and pressure systems. We generalise loosely when we say that our frontal lows move from south-west to north-east. In fact they have certain favourite tracks, preferring, for example, to avoid high ground and repelled by strong anticyclones. But any particular depression may take almost any track as Fig. 68 shows. How can such aimless wandering be forecast? One well-tried method is to plot the successive positions of the centre at, say, 6-hourly intervals and extrapolate to fix its future position (assuming no change in speed or direction is expected). This may be used intelligently for curved tracks and diminishing or increasing speeds.

No forecaster relies entirely on one method. Several techniques rest on the fact that the geostrophic wind speed is inversely proportional to the isobar spacing. This makes it possible to devise a *Geostrophic Wind Scale* which is a most valuable forecasting tool. Fig. 69 shows Geostrophic Wind Scales suitable for various scales of chart used at different times in the *Daily Weather Report*. The scale 1 : 5,000,000 (approximately 80 miles to the inch), formerly used in the *Daily Weather Report*, is the best scale for detailed studies of British weather: the weather maps of Chapters 5 and 6 were originally plotted on this scale. (Copies of 1 : 5,000,000 British Isles outline charts are obtainable from the Meteorological Office.) The smaller the chart scale, the less accurate is the use of the wind scale. The method is quite simple. Lay the appropriate scale across the isobars (at right angles to them) so that one isobar cuts the base line and read off where the next isobar cuts the scale to obtain the geostrophic wind speed in that region. The same scale can also be used for estimating the direction and speed of travel of fronts: lay it along a front and the isobar spacing gives the speed of movement of the front (at right angles to itself); warm fronts move at speeds roughly two-thirds that indicated by the wind scale.

The forecaster also makes considerable use of pressure changes or *tendencies* (the change, up or down, during the last three hours). Lines of equal tendency—*isallobars*—can be drawn on the chart and show at a glance the present pattern of pressure change. Fig. 70 shows the isallobars drawn for the frontal depression of Fig. 43. A line drawn from the area of greatest pressure rise (largest positive tendencies) behind the depression to that of greatest pressure fall (largest negative tendencies) ahead of it gives the likely direction of movement. Tendencies provide evidence of the intensification or decay of systems and a falling pressure at a lone weather ship is often the first clue to fresh cyclogenesis out in the Atlantic.

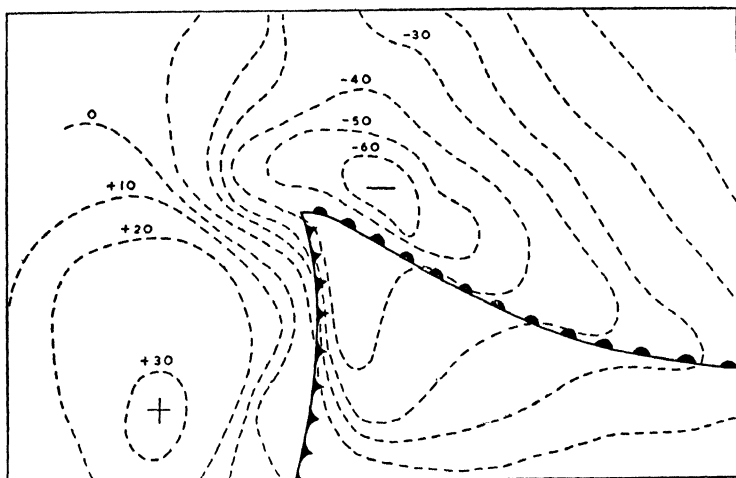


Fig. 70. Isallobars of a frontal depression

The Use of Aerological Diagrams

The introduction of the tephigram (by Napier Shaw in 1925) and of similar diagrams on which aircraft and (later) radiosonde ascents could be plotted gave the forecaster a much-needed three-dimensional view of the weather situation. This enabled him in particular to forecast the development of weather within the air mass due to movement or diurnal changes, especially as regards instability and the likelihood of convection. In Chapter 4 we used the temperature-height diagram to depict the development of convection cloud. The forecaster must foresee such possibilities long before, usually on the basis of an early morning ascent made when very different conditions prevail. An example will show how this is done.

Fig. 71 shows an ascent made from the radiosonde station in south-east England early on a summer morning in 1955. There is an unmistakable nocturnal surface inversion but, above this stable layer, the air is conditionally unstable for considerable depths and furthermore is extremely moist throughout. Here is a situation ominous indeed should day-time warming remove the surface inversion.

The surface temperature is bound to rise, but how much? With the versatile tephigram there is a sound technique for forecasting maximum temperatures but this cannot be applied to the temperature-height diagram. The forecaster could reasonably have accepted a representative maximum of the previous day, *e.g.* 83° F. (London Airport), since the air mass was the same. The dew-point too will rise, as the deposited dew evaporates

and adds moisture to the air. It would be reasonable to increase it to, say, 65° (as it is 100 feet above the surface).

With a dew-point of 65° and a dry-bulb temperature of 83° carry out the usual construction (see page 72). Cloud base is reached at 3,800 feet. The tops could extend to over 40,000 feet! In fact, any temperature above about 76° F. could set off this powerful convection. Under such conditions not only showers but thunderstorms are more than likely. In fact 14th July, 1955, was notable for severe and widespread thundery activity south-east of a line from the Wash to the Isle of Wight. Six people were killed by lightning that day (three were children struck while sheltering under an oak tree); at the Royal Ascot meeting, 44 people were injured (one died later) when lightning struck a metal rail (also see Chapter 11).

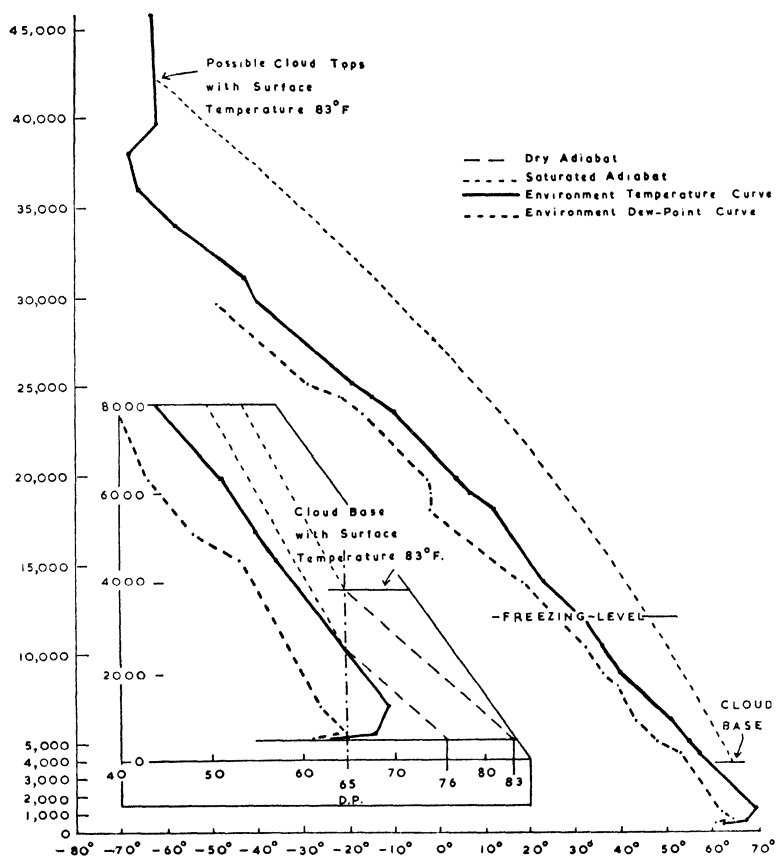


Fig. 71. An example of the use of the temperature-height diagram in thunderstorm forecasting. Crawley, Sussex, 14th July, 1955, 0212 G.M.T.

Upper Air Charts

Modern forecasting gives increasing prominence to the upper air where patterns change more slowly, more steadily and more predictably than those of the surface chart. In the United States, the emphasis is on the use of contour charts. Much attention has been paid to the long waves in the westerlies and the eminent Swedish meteorologist G. C. Rossby has devised formulae (based on various assumptions) for predicting their movement.

British practice favours the use of thickness patterns. Techniques developed largely by Dr. R. C. Sutcliffe (active in forecasting research at Dunstable) are based on extrapolation methods plus physical reasoning on the nature of convergence and divergence. Under some conditions the surface systems are 'steered' by the thermal winds and certain patterns of thickness lines denote to the trained eye developments like the appearance or deepening of lows and the building up of highs.

These new methods are inevitably outside the ken of the non-professional and beyond the scope of this book. In the British Meteorological Office, such techniques are used in conjunction with well-tried surface methods in order to build up the prebaratics that are the basis of our short-term forecasting. The two procedures do not invariably give the same answer. In the end it is the wise judgement of experienced senior forecasters that must decide the issue.

Statistical Forecasting

Among more speculative methods are those concerned with finding in past statistics the clue to future developments. This is largely a matter of studying records over a long period to discover if there is any tendency for weather patterns to recur at the same time of year. It is a fascinating pursuit, in which enthusiastic amateurs can usefully join. And such annual recurrences—or *singularities*—undoubtedly exist with a regularity that cannot be blamed on chance. No doubt the general circulation tends to arrange itself in roughly the same way at roughly the same time each year, though why is not known.

At one time the so-called 'Buchan Spells' aroused great interest. In 1869 Alexander Buchan, a Scottish meteorologist (and a very good one), used 10 years of Scottish data to pick out 6 'cold periods' and 3 'warm periods.' These were not meant to apply outside southern Scotland and Buchan never claimed that his spells always occurred between the same dates.

The fullest studies of singularities in this country have been carried out by Dr. C. E. P. Brooks (to whose book *The English Climate* the interested reader is referred). He gives a list of such singularities some of which have occurred in most of the 52 years studied. While their existence is not in doubt, their value for forecasting is another matter.

For example, there is a well-established stormy period (unsettled depression weather) in early January; it occurred in 45 out of 52 years. Its average dates of beginning and ending are 5th January and 17th January, respectively, but it has begun as early as 1st January and as late as 18th January and it has ended as early as 6th January and as late as 22nd January. What this amounts to is that there is likely (not certain) to be a stormy period during the first three weeks of January. This is not information to be scorned but it has little application to day-to-day forecasting. It is better to keep an eye on the current weather situation than on the past statistics.

On a larger time-scale, the records reveal any number of fluctuations or *cycles* which, with periods measured in years and even centuries, belong to the study of climate rather than of weather. Some of these are associated with the cycle of sunspot frequency which reaches a peak every 11 years on average, with double and quadruple cycles thrown in. At sunspot maxima, there is increased radiation in the ultra-violet wave-lengths but we are not at all sure what effect such changes in solar activity have on the atmosphere. Statistical analyses of pressure records point to the existence of sometimes regular pressure oscillations with periods of 72 and 48 days and their sub-multiples down to 6 days. Their origin is not known. The Meteorological Office is examining these periodicities to see if they have any value for long-range forecasting.

Forecasting by Analogues

This method also rests on the belief that meteorological history repeats itself. A long series of synoptic charts (60 years, or so) is classified so that any particular pattern can be readily sought out: a punched card system is used and selection of the required pattern is mechanical. A chart is found that most closely resembles the current one and the ensuing development is taken as a guide to what will happen in the future. Although much success has been claimed (by Krick and his associates in the United States, and by Baur in Germany), the method has not found favour in this country. The weather will behave in the same way twice only if all the circumstances are identical and this is hard to establish. The technique has been described as a kind of mechanical memory. Actually, an experienced forecaster carries his own analogue system in his mind but is not rigidly bound to it.

Mathematical Forecasting

It has been the pipe-dream of many a meteorologist that, ultimately, tomorrow's weather shall be mathematically predictable. Over 30 years

ago, the late Dr. L. F. Richardson wrote a book called *Weather Forecasting by Numerical Process*. His ideas were doomed to failure at the time if only for the reason that the calculations (even if soundly based) would take so long that the computers could hardly keep up with the weather. Today, however, we have electronic computers that do calculations often in less time than it takes to set out the problem on paper.

Sir Graham Sutton, Director of the Meteorological Office and himself a distinguished mathematician, has recently dispelled any false hopes that the new calculating machines may have raised. The computers can do only what they are told to do, however quickly they do it. For reasons discussed earlier in this chapter, it is unlikely that we shall ever know enough for the complete mathematical solution of the forecasting problem.

This is not to suggest that mathematical techniques will not play an increasingly important part in forecasting. Indeed, the British Meteorological Office was the first national organisation to experiment with numerical forecasting and is to have its own computer. The method is based on a much simplified 'model' atmosphere which is taken to represent part of the real atmosphere. By ignoring a good deal, the pressure changes at a number of points can be computed by means of complex formulae. For a 24-hour forecast the task takes one to two hours on an advanced computer. The end-product provides the skeleton of the prebaratic and it still requires the skill and experience of the forecaster to 'put in the weather.' Encouragingly, the prebaratics so far produced by the mathematical 'model' have been about as good as those evolved by more usual methods.

Long-Range Forecasting

There are no miracle methods in short-range forecasting: still less are they available for extended-range forecasts (up to a week ahead), and long-range (more than a week). The methods being tried are really extensions of the familiar short-range techniques. Beyond the fourth day or so, detailed accuracy is not to be expected, except in unusually static situations. At long range only the broad trends of pressure, temperature and precipitation can be indicated.

The Americans have been experimenting with 5-day and even 30-day forecasts based on contour charts for the 700 mb level (about 10,000 feet) which show the average flow over those periods. The long waves are moved on by simplified formulae as on ordinary charts. They look also for broad fluctuations in the strength of the zonal circulation (Chapter 7). Empirical and subjective procedures inevitably enter the picture. British forecasters have been critical of the mean 30-day chart which, they say, can have little real significance. The Germans use similar techniques for

10-day forecasts. The Russians attempt 7-day forecasts partly on their ability to predict the tracks of the anticyclones that so largely dominate their winter weather.

British practice, more cautious, persists with 4-day forecasts based largely on thickness patterns. Dr. J. M. Stagg (Assistant Director of the Meteorological Office), recently reviewing the situation, has concluded that there is as yet 'no ready-made procedure or combination of procedures suitable for long-range forecasting in the British Isles.' The methods being used are more likely to be successful in large continental areas where conditions are more uniform. In complex coastal regions like western Europe, and the British Isles in particular, such is the variety of influences guiding or distracting the development of weather that even the safest looking forecast can go suddenly and unpredictably astray.

Forecasting for the Amateur

Amateur weather forecasting is no doubt as old as human society. People whose occupations keep them out of doors, on the land or at sea, acquire what seems to be an instinct for the immediate weather prospect. Their abilities claim our respect even if they have never heard of warm fronts or unstable air masses, for they have learned by experience to read the signs that these present. They know the significance of certain wind directions though they may be quite ignorant of Buys Ballot's Law. At the same time, too much faith must not be placed in them. It is possible at any one time to see perhaps 50 or even 100 miles of sky in any direction: this may mean a few hours of weather. Sky watching and an intelligent appraisal of winds are useful aids to local, short-term forecasting. But the front far out in the Atlantic that will provide tomorrow evening's rain is beyond even the countryman's keen perception though it does not escape the synoptic eye of the official forecaster.

Weather Lore

Both popular experience and popular misconceptions find expression in the proverbs of weather lore (of which there is a famous collection by Inwards: other examples are contained in the 'weather rules' of the Shepherd of Banbury). Some weather sayings are well-founded, others have amusement value only. The phase of the moon has no prognostic value and St. Swithin's Day is meteorologically no different from any other.

On the other hand, rhymes about red and grey skies have more value, assuming that our weather travels broadly from a western to an eastern quarter. Red skies are seen when the sun is low and the atmosphere dry

and hazy with many dust particles (as in anticyclonic conditions). 'Red sky at night' suggests that these desirable conditions lie to the west and are on the way, 'red sky at morning' that they have already passed away eastwards. The rainbow, seen from the ground when the sun is low in the sky, is sometimes substituted for the red sky in these jingles, none of which is wholly reliable. 'Rain before seven, fine by eleven' works with many cold fronts and weak warm fronts with narrow rain belts. And when depression follows depression, it is true that 'A nor'wester is not long in debt to a sou'wester.'

There are many more examples, some mentioned later in this chapter. The interested reader will find it diverting to examine the sayings of his locality and test their validity, as well as puzzling out their explanation (if any).

Certain animals and plants are sometimes regarded as weather prophets. There is no question that some living things react to existing meteorological conditions—we do ourselves—but there is little evidence that their behaviour is influenced by weather yet to come. Certain flowers, like the pimpernel, close their petals by day when the relative humidity exceeds about 80 per cent. Seaweed—reputedly the forecaster's standby!—reacts visibly to changes in humidity. So do certain chemicals, which conveniently change colour. None of this is necessarily prognostic except in so far as it indicates the arrival of a type of air from which certain kinds of weather can be expected.

Some Sky Signs

Ignoring such doubtful aids, what are the prospects of success for the amateur forecaster? They vary, of course, from situation to situation, but some knowledge of the broad weather situation and of local conditions provides a good beginning. If we know that we are in Pm air we are not deceived by the clear skies of early morning. The convection clouds will build up sooner or later and the earlier this happens the cloudier the day is likely to be. Since we have no upper-air data we cannot say how much they will grow. We can only wait and watch their vertical development. If they remain flattened (or if they flatten after an earlier build-up) then there is an inversion aloft: this may be anticyclonic (Plate XIII) and a fine or fair day is indicated: or there may be an advancing warm front (in which case look for the tell-tale cirrus wisps). But if the cumulus show much vertical growth (less is necessary in winter than in summer) and especially if the tops become frayed (Chapter 4), showers are likely (Plates III and IV).

When clouds appear like rocks and towers,
The earth's refreshed by frequent showers

If we know that Tm or Pmr air is in possession, a dull early morning need give no cause for pessimism. It generally means no more than a thin layer of stratus or stratocumulus and, in spring, summer and autumn, it is likely that breaks will soon appear (the patch of blue sky 'big enough to make a sailor's pair of trousers'). This will be followed by a more or less complete dispersal of cloud and a warm day. If there is other cloud above, the lower layer is less likely to break up. Another bad beginning that actually portends well is early morning mist or fog for (inland) this usually means anticyclonic conditions and, except in winter, should clear during the morning to give a fine day.

If a shallow summer depression is spreading up from the south, or in the stagnant air of a col, we must keep in mind the possibility of heat thunderstorms. A useful warning is often provided by the cloud type *altocumulus castellatus*, a medium cloud with marked vertical (castle-like) growth, which indicates instability at those levels (Plate XII). With more day-time warming, it only needs convection currents from the surface to penetrate to this unstable layer for convectational over-turning to be powerfully reinforced through considerable depths. Many heat waves break up in this way. With large cumulus and cumulonimbus about, it is often rash to predict the usual evening dispersal of convection cloud. They have the habit of persisting and thunderstorms often break out anew in the evening. The reason for this is not always clear. It may be that in the evening the cloud tops lose heat by radiation while the lower parts remain warm so that the lapse rate within the cloud increases and it becomes less stable than before. Or sometimes, as F. H. Ludlam has suggested, katabatic downflow from nearby hills causes local convergence and pushing up of air over valleys and plains.

How reliable is *altocumulus castellatus* as a thunderstorm warning? Investigation has shown that, on 7 out of 10 occasions when this cloud was observed in the London area, thunder occurred somewhere in south-east England on the same or the following day. Strong winds aloft do not favour thunderstorm development. If large cumulus are observed but seem to lean over (meaning that they are being torn by rapid changes of wind with height), thunderstorms are less likely.

The most easily identifiable weather system is the warm front. In an American study the typical cloud sequence (as described in Chapter 6) was observed before 84 per cent of 135 frontal depressions. It is the sequence that matters, cirrus itself is not enough (it may be orographic). Fast-moving cirrus is a likely herald of the coming front (see Plate IX) and the faster it moves the quicker we may expect further changes to occur. After a few hours the cirrus should give way to the cirrostratus. This is reliably labelled by the characteristic halo which is a useful prog-

nostic sign ('The moon with a circle brings water in her beak'). Some studies made in London suggest that there is about a 70 per cent chance of rain within 24 hours of the observation of a halo (some haloes, of course, appear behind a cold front). Thickening altostratus is perhaps the surest sign of an approaching warm front.

The Use of the Barometer

The possession of a barometer, or better still a barograph, increases the scope of amateur forecasting. Some means of assessing wind direction is also a help. With the barometer it is, of course, the changes (barometric tendencies) that matter, not the absolute readings. If identification of the warm front cloud sequence is reinforced by a falling barometer and a backing wind, we are the more confident in our prediction of rain. The quicker the pressure fall, the more vigorous—and short lived—is the weather change.

Long foretold, long last,
Short notice, soon past

is often applicable to pressure changes. Persistent systems like warm anti-cyclones announce their coming by slow changes of pressure.

A falling barometer and a freshening east wind generally spell a good deal of rain ('Rain from the east, twelve hours at least'). An appeal to Buys Ballot's Law tells us that pressure is low somewhere to south or south-west of us and quite probably there is most of a frontal depression yet to come. By application of the same law we can see where we are in any pressure distribution.

Limitations of Single Observer Forecasting

A little practice will make it quite clear that forecasting by the individual has its limitations. Even if cirrostratus thickens into altostratus according to plan, even if pressure falls, we cannot be sure that rain will follow. It may not do so if the front is weak, and we cannot tell beforehand what air mass lies behind an advancing front. A cold front may be followed by the usual marked wind veer, rapidly rising pressure and cold showery Pm air from a north-westerly quarter: or, as sometimes happens, by only a slight wind change, little rise of pressure and mild Pmr air from the south-west. In the latter case there is often a large trough of low pressure yet to come, which may well bring more precipitation than the front itself. We cannot know whether a frontal low is occluded or not. There are many marginal situations in which our unaided efforts are insufficient. We need to know the synoptic pattern.

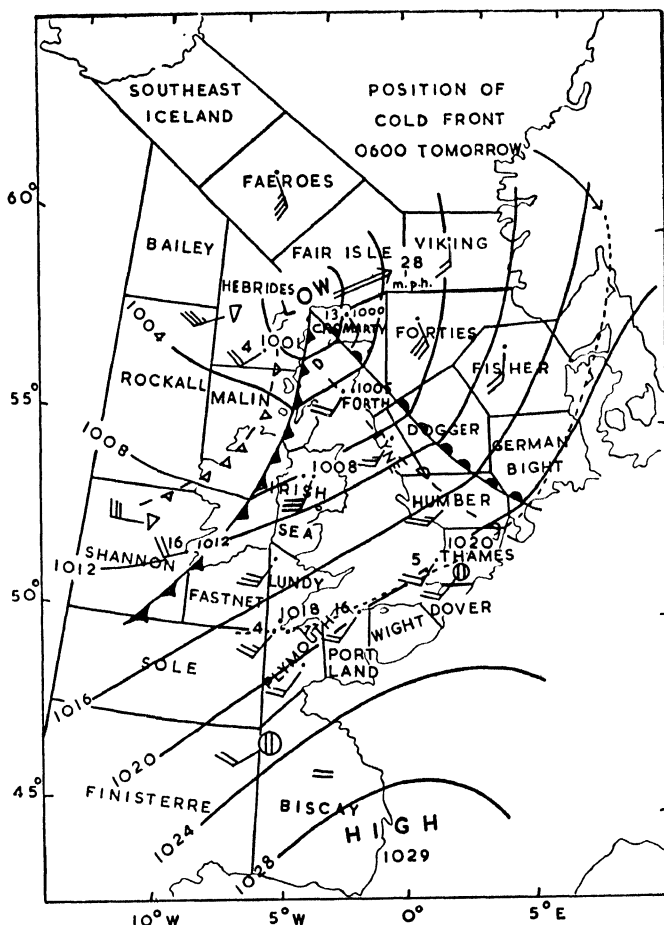


Fig. 72. Synoptic chart for around mid-day, 9th May, 1956, constructed from the 1.40 p.m. shipping forecast. The frontal positions given in the bulletin (broken lines; symbols not filled in) are for 6 a.m. and must be adjusted (full lines; solid symbols) to fit in with the observations, which are for 11 a.m.

It is best to be opportunist about this and take advantage of all possible information. The *Daily Weather Report*, if available, gives the immediate history of the present situation and shows broadly what to expect. The B.B.C. general weather bulletins on the Home Service help but give no synoptic details. The 'prebaratics' published in some morning papers give the opinion of the forecasters at Dunstable (the best opinion available at the time of issue) regarding the situation at noon.

While no substitute for the defunct Airmet broadcasts, the new weather bulletins for shipping issued on the Light Programme make it possible to

build up a very rough and ready synoptic chart (if we are prepared to be up very early or very late or to be home after lunch). The bulletins give the synoptic situation of a few hours previously, including the position and expected movement of pressure centres and fronts, and observations of wind, visibility, weather and pressure at 8 coastal stations (as well as the usual shipping area forecasts). By dint of hasty plotting (using the symbols given in the Appendix) and a little juggling, it is possible to construct a chart like that of Fig. 72. This example, plotted from the 1.40 p.m. bulletin, gives a picture for roughly the same time as the newspaper prebaratic and can be used to check it.

LIVING WITH THE WEATHER

Living with the weather is rather like living with a relative of moody and capricious character whose presence in the house is not of our own choosing. We live with the weather more successfully the better we understand it. As in human relationships, understanding leads at least to toleration and to some degree of adjustment. Even crusty relatives have their benevolent moments and our weather in particular is as often pleasant as it is otherwise. We must learn to make the most of its benefits and to protect ourselves as well as we can against its nastier moments. Through our increasing understanding of the weather we are already learning to predict its behaviour. Finally, understanding can lead to a measure of control, though the opportunities at present seem severely limited. It is from this point of view that we now look at the weather again, in its good and bad moods, and examine its impact on the life of the community.

Weather as a Friend

Good weather finds a place in our memories less readily than bad. Yet all of us can surely remember some enjoyable occasion or period in our lives that was made even more delectable by perfect weather. We all owe gratitude at one time or another to a summer anticyclone that blessed a certain much-needed holiday or speeded a period of convalescence. Harassed mothers thank the sunshine that allows the toddler to play out in the garden instead of being a nuisance in the house. The successful garden-party, village fête or Sunday outing owes much to the kindness of the weather. In good weather, other things being equal, life goes according to plan.

Townsmen (unless they are also gardeners) sometimes tend to forget that what is good weather for them is not necessarily good weather for the farmer. The farmer needs dry weather in spring for sowing and in mid-summer for haymaking; he hopes for sunny spells for ripening corn and gathering the harvest; he waits for quiet weather for spraying with insecticides and weedkiller and prays for a rainless period to follow. For reasons such as these, the notification of fine spells (one of the special services of the Meteorological Office) is so useful for his planning of

operations. Equally, however, rain (in moderation) is necessary, for germination and sprouting in spring and later for the growth of lush grass and the achievement of good crop yields. Rain, of the right type and amount and at the right time, is an essential part of good weather on the farm.

Weather as an Enemy

It is however bad weather that makes news. We are perhaps less vulnerable to weather catastrophes today than in earlier times but even now no year passes without the weather extracting its toll in human lives and material damage somewhere in the world. In some regions such events are frequent and expected and there is some human adjustment to their possible occurrence. Our own country enjoys an equable climate and there is no hint of danger in our average climatic statistics. But our everyday weather is very variable and we should expect and certainly get all manner of unseasonable occurrences. Unfortunately, the organisation of our lives is keyed to that mythical mean and the occasional (though inevitable) extreme often catches us woefully unprepared.

Destructive Winds

Strong winds, no more than a nuisance to most of us, are a positive worry to the farmer. They chill and desiccate bare soil (occasionally causing soil erosion in dry areas like the Fens), they damage crops and endanger animals on exposed hill farms. Calamitous winds, a regular accompaniment of tropical storms, occur more rarely in our latitudes and then only with the most intense depressions (Chapter 6).

A wind of gale force is defined statistically as one of Force 8 or more on the Beaufort Scale, that is, an average speed of over 38 m.p.h. though this would include gusts reaching probably 50 m.p.h. Gales occur frequently on our surrounding seas and exposed coasts, averaging about 30 a year on western coasts but less than half this on more sheltered north-eastern coasts. The Meteorological Office was initially called into being in 1855 for the purpose of providing gale warnings for shipping and this remains one of its most vital functions. The forecasting of gales presents no special problems, being based on the expected development and movement of pressure systems.

Few winters pass without a severe gale and reports of uprooted trees, fallen telegraph poles, damaged buildings, dislocated traffic and the occasional casualties that these must entail. In the gale of 15th January, 1954, in Sunderland, a seven-ton furniture van was blown 60 feet across a road and overturned. Gusts capable of such feats must reach 70 or 80 m.p.h. On exposed coasts, winds of well over 100 m.p.h. have been experienced, rivalling those of a moderate tropical hurricane. In the remarkable



PLATE XIX. Local weather: valley fog in Edale, Yorkshire



PLATE XX. Local weather: shallow ground fog over Stevenage, Hertfordshire

Orkneys gale of mid-January, 1952 (which caused £500,000 worth of damage) the anemometer recorded 127 m.p.h. and then blew away. Sometimes, violent squalls of wind associated with cold fronts (line squalls) or thunderstorms occur, even inland. During the very severe north-westerly gale of 17th December, 1952 (occurring in the rear of an occluded depression with a central pressure as low as 961 mb), an extraordinary gust of 111 m.p.h. was registered by the Dines anemometer at Cranwell, Lincolnshire.

The destructiveness of tropical hurricanes is due only partly to the direct effect of the winds. These are formidable enough, being capable of lifting houses bodily from their foundations or of sweeping a train off the line (as once happened to a rescue train in Florida). During the New England hurricane of 1938 (Chapter 7) 275 million trees were destroyed or damaged. When Hurricane Hazel passed near New York City, the Empire State Building set up a rythmical 16-inch sway. But the most terrible disasters have been caused by the *storm waves* raised in coastal waters and the flooding of low coastal plains. The worst hurricane catastrophe in the United States occurred in 1900 when Galveston, Texas, was overwhelmed by 15 feet of water and more than 6,000 people were drowned, while damage to property exceeded 20 million dollars. In an even worse inundation in the Bay of Bengal (1876) the death roll was estimated at 100,000.

The more localised havoc wrought by tornadoes is due entirely to their tremendous winds (Chapter 7). These have been known to lift an 80-ton steel railway coach and to drive objects into and through wood or iron with the force of a bullet. On average nearly 250 people die each year as a result of 'twisters' in the United States.

Only limited precautionary measures can be taken against hurricanes, among them the building of strong structures, the avoidance of floodable land for housing, the provision of emergency food and water supplies and the evacuation of people and livestock from vulnerable coastal areas when warnings are issued. The United States storm warning system is very efficient. Tornadoes spring up without warning and little can be done about them, except to take cover and hope for the best.

Heavy Rains and River Floods

Every now and again—though not often enough apparently to provoke us into counter-action—there occurs in this country serious river flooding due to excessive rainfall. This may take the form of long periods of unrelenting rain brought by frequent depressions or of exceptionally heavy downpours of 'cloudburst' type. Local factors like relief and the state of the ground are also involved.

Many of our rivers flood pretty regularly during our wetter winters. In the Thames valley above London, many places registered their highest-ever water mark during serious flooding in November, 1894, due entirely to persistent depression rain: over 8 inches fell between 23rd October and 17th November. More widespread were the floods of March, 1947 (see Frontispiece). This flooding, which marked the end of that long and bitter winter, resulted from a combination of heavy frontal rain, melting snow and a frozen impervious ground. The Thames, Severn, Trent, Yorkshire Ouse were among many rivers that rose inexorably and drowned large areas, including 100 square miles of valuable Fenland. The cost to the country was about £12 million. Flood, with its grim sequel famine, is part of the history of the overcrowded valley lands of Monsoon Asia.

Cloudbursts cause more local and sudden flooding. They may occur in lowland country—a Lincolnshire thunderstorm in May, 1920, unleashed 4 inches of rain in 4 hours and drowned 23 people in the town of Louth—but the effects are always accentuated by the steep slopes and narrow valleys of mountainous regions. Cloudbursts generally occur on the lee slopes of high ground: here the up-currents that hold the raindrops in suspension are checked and, robbed of this support, the cloud spills its content in a great deluge. Such a downpour befell the Dolgelly district of Wales when probably 3 inches of rain fell in two short evening storms in May, 1944. Water piled up to depths of 20 to 30 feet in the narrow bottle-neck valleys tributary to the Mawddach. Young trees were uprooted, a stone bridge was washed away, road surfaces torn up, a small hydro-electric plant smashed to pieces and transported ballast dumped in fan-shaped deltas.

Another example of very heavy rain in a thundery summer depression aggravated by orographic uplift caused the unprecedented floods of August, 1949, in the Tweed valley. Over 5 inches of rain in a day fell on rivers already swollen and ground already saturated by considerable rain throughout the preceding week. In these Border floods over £1 million worth of damage was reported in Berwickshire alone.

Perhaps no recent natural disaster has shocked us as much as the flooding that devastated the Devonshire resort of Lynmouth on 15th August, 1952. The immediate cause of this catastrophe was a vigorous Atlantic depression with very moist and unstable warm sector air which was forced aloft by the high ground of Exmoor. As a result, 17 square miles of country received over 8 inches of rain and 11 or even 12 inches may have fallen in places. Much of Exmoor is peat-covered with an ironstone layer beneath the surface which impedes drainage and the ground was already waterlogged due to considerable rain earlier in the month. The phenomenal downpour of the 15th, rejected by the soaked soil, cascaded via tributaries into the twin rivers East and West Lyn. These both fall 1,500

feet in under 4 miles and become torrential under only moderately wet conditions. Lynmouth lies in a narrow valley where the two rivers join and thus suffered the full impact of the torrent. The culvert that normally leads the West Lyn beneath part of Lynmouth into the East Lyn soon became choked with debris and the river took its old course through the town. 28 people were drowned, 17 bridges destroyed and much of the town laid waste.

Such disasters cannot be forecast but we need not be quite at the mercy of our rivers. As regards periodic winter flooding, analysis of past records shows what the river is capable of doing. The building of houses on land liable to flood must be forbidden. The construction of bridges must be planned to withstand the river at its worst. Relationships can be established between rainfall and river flow which make prediction possible. Taming a river is more than an engineering problem, it is part of regional planning. For example, the afforestation of steep slopes helps, by slowing up run-off. Large-scale projects like the Tennessee Valley Authority embrace flood-control as a matter of course. In Britain our problems and opportunities are smaller but there is still need for more adaptation to the possible behaviour of our rivers.

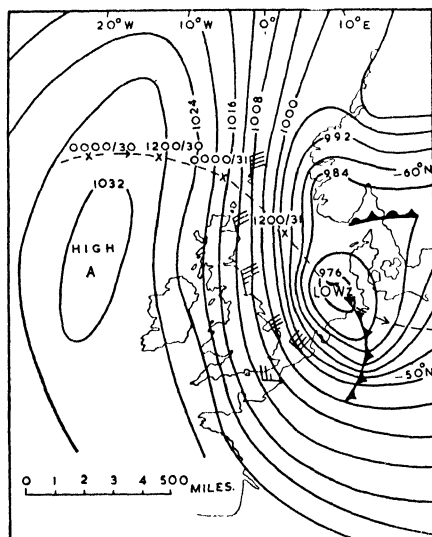
Storm Surges

In another mood, nature sometimes troubles us with floods in flat coastal districts, particularly on our east coasts. The tidal fluctuations at sea-level depend on the relative positions of sun and moon and, for a given spot, can be forecast for years ahead with great accuracy. But meteorological conditions occasionally intervene to give water levels much higher than the predicted ones. These *storm surges* are caused mainly by high winds, especially those with a long fetch, in a particular configuration of coastline and sea.

The situation becomes significant for our vulnerable east coast when a deep depression has moved across to Scandinavia and northerly or north-westerly gales blow the length of the North Sea. These conditions become dangerous only if they coincide with particularly high tides. Then there may be tremendous piling up of water in the bottle-neck of the southern North Sea, which can be relieved only by flooding to east and west. On 1st January, 1922, sea-level at Southend was 11 feet above the expected value but this occurred at low water and no harm was done. But in a similar situation on 7th January, 1928, water level rose nearly 6 feet above predicted high water and serious flooding affected the Thames estuary and the London area.

The worst storm surge in recent memory was that of 1st February, 1953. A breakaway depression from the Atlantic had crossed Scotland, all the

while deepening and giving violent gales (which had already had tragic consequences in the loss of the ferry boat *Princess Victoria* in the northern Irish Sea). The low then veered south-eastwards, its central pressure now only 970 mb, while high pressure built up westwards (Fig. 73). Very fierce



Left: The synoptic situation that morning (the broken line shows track of the depression)

Below: The extent of the flooding (black), the figures representing the surge (in feet) at high water

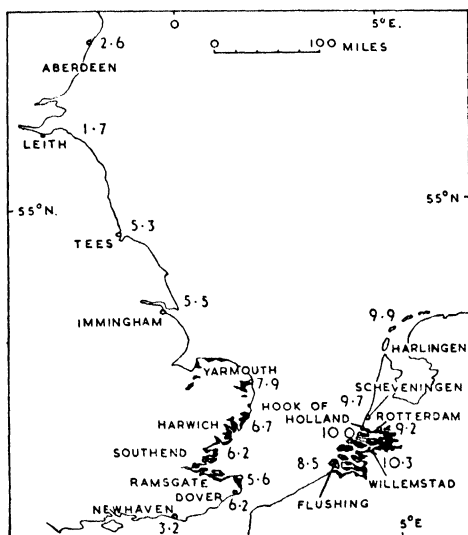


Fig. 73. The storm surge of 1st February, 1953, in the North Sea

northerly gales whipped up enormous seas at roughly the time of unusually high spring tides. On our east coast, water rose up to 8 feet above predicted high water in places and, in one of the estuaries of the even more exposed Dutch coast, it was fully 17 feet above the expected height.

Both to east and west the swollen seas submerged dykes and sea walls. Once over the tops, the waves, lashed by the wind, pounded away at the weaker inner slopes and breached them at many points. In eastern England flooding of low-lying areas from the Humber to the Thames estuary caused nearly 300 deaths, grievous loss of crops and livestock and great damage to property. The Dutch suffered even more. Nearly 1,800 people were drowned and more than 50,000 evacuated: 9 per cent of all agricultural land was inundated.

North Sea surges can be predicted if a reliable forecast of the pressure and wind situation is available; this must be considered in relation to the expected time and height of high water. A rapid and effective warning system is the prime necessity. At the same time we should look to our sea defences and avoid building on vulnerable low-lying sites.

Hail and Lightning

Hail and lightning are the most dramatic features of thunderstorms but they affect us relatively little. Much of the hail we experience is small, soft and harmless. But once every few years a really vicious hailstorm occurs somewhere in our eastern or midland districts, most often with a summer heat thunderstorm. Hailstones an inch or two across, hard and weighty projectiles, fall with great force, damaging crops and shattering glasshouses, skylights and windows. Hailstones of 4 inches diameter may fall in more continental regions where severe storms have been known to kill sheep and even people. In the United States and Germany, insurance against hail is common. Anti-hail measures are described on page 241.

Lightning activity as such is not recorded in detail in this country but a rough estimate suggests that any square mile may expect to receive about six flashes to ground in an average year. The lightning risk is therefore slight. Most lightning fatalities occur in the open: the most dangerous spots during a thunderstorm are on exposed hill-tops, under isolated trees and near wire fences. Indoors, especially in towns, the risks have been much reduced by the widespread use of the lightning conductor. In England and Wales the chances of being killed by lightning in a year are about 1 in 4 million. In the United States they are 3 in a million, still small enough odds.

Lightning can cause minor injuries, from shock to burns. It is a fairly common cause of fires, including forest fires. In this country it is the most frequent reason for breakdown of the electricity service, striking over-

head wires and causing powerful surges of current which may damage their insulation. It is not, of course, true that lightning never strikes the same spot twice: tall structures like the Empire State Building and the Eiffel Tower are struck repeatedly.

Drought and the Water Problem

Drought is an insidious enemy. It is recognised by the farmer and the water engineer while the man-in-the-street, less aware of danger, is still marvelling at the continued fine weather. For statistical purposes, an *absolute drought* is a period of at least 15 consecutive days none of which has a hundredth of an inch of rain (the criterion of a 'rain-day'). If this tiny amount falls the drought is officially 'broken' but thirsty crops and depleted reservoirs are no better off. A *partial drought*—statistically at least 29 consecutive days with some slight rain but averaging less than a hundredth of an inch a day—is serious enough.

In this country droughts occur under prolonged anticyclonic conditions (especially the Azores High in summer) or when winds blow persistently from a dry source (*e.g.* easterly winds in late winter). Grouping absolute and partial droughts together, most of England is liable to get one or two in a year, the likelihood being greatest in the south-east, least in the north-west. A persistent drought eventually becomes apparent to all, in the sun-baked soil in the garden, the dusty roads, the withered look of grass and crops, the shortage of vegetables in the shops, the outbreaks of heath and forest fires and, sooner or later, the appeals to restrict the consumption of water.

The longest absolute drought on record in this country was 60 days, reached in 1893 on parts of the South coast and the longest partial drought (Hailsham, Sussex, also 1893), lasted 114 days. 1921 was another very dry year with rainfall only about half the average over large areas. The most extensive absolute drought occurred in the summer of 1947: only small areas failed to record it during August and the first half of September. One part of Kent had 50 consecutive rainless days. Although the records have been scanned and various cycles in the occurrence of very dry years have been claimed, we know of no sound basis for forecasting them.

These periodic droughts only underline our ever-present—and now rapidly growing—water problem. It is not a problem of over-all shortage of water but of its mal-distribution. Britain has been built the wrong way round. The high west and north extract an undue share of the available water in orographic rainfall, so that parts of North Wales, the Lake District and northern Scotland get over 100 inches annually, while the Thames Estuary gets only 20. Large parts of south-east England actually have insufficient rainfall for the water needs of most crops. The complex

relationship between the weather and the loss of water by evaporation and transpiration from soil and plants has been much studied, in this country notably by Dr. H. L. Penman of the Rothamsted Experimental Station. Investigations show that, south-east of a line from South Lincolnshire to Dorset, in 8 years out of 10, the actual summer rainfall is unlikely to meet the water needs of crops if they are to grow to their best. The deficiency can be made good by supplementary irrigation: this is now practised in many areas and pays handsome dividends.

But this regional problem is only part of our national problem. Irrigation will mean yet another growing demand on our water supply. Domestic consumption is steadily on the increase and the requirements of industry are mounting even more rapidly. We are already drawing on our underground reserves of water faster than they are being replenished by rainfall. In London and the south-east water may soon be a scarce commodity. There is no easy way out of this situation. Schemes have been mooted for controlling the Thames by a number of reservoirs: these would eliminate most flooding and the flood water that formerly went to waste would help maintain an adequate dry-weather flow. But lowland England does not provide good sites for large impounding reservoirs and the sacrifice of productive land would in any case be enormous. There are ways of re-charging our underground reserves which merit consideration. The major problem is transferring water from the over-endowed west to the thirsty east. Nothing less than a national water grid may be necessary with all its financial and administrative implications.

The Fight against Frost

Frost is a weather hazard we are not likely to escape in any year. The number of days with frost per annum may be only 10 in our south-western coastal areas but increases rapidly inland to an average of about 70 in parts of East Anglia and 100 over much of Scotland. The occurrence of frost is very dependent on local conditions, as we have seen in Chapter 9. *Air frost* (screen temperature below 32° F.) must be distinguished from *ground frost*: this is said to have occurred when the grass minimum temperature has fallen to 30° F. or below, on the basis that plant tissues do not suffer injury until the temperature falls well below freezing point. On frosty nights the coldest air is at the ground: a ground frost alone may spell danger to only the most vulnerable of young plants, but an air frost (which means even lower temperatures near the ground) may mean destruction to most plants.

Damage to plants—as well as discomfort to human beings—due to low temperatures is always greater if the wind is fresh or strong. Frosts are therefore classified according to the combined effect of temperature and

wind and the descriptive terms that figure in the weather forecasts ('slight,' 'hard,' etc.) have definite meanings. For example, a temperature of 27° F. signifies a 'slight frost' in calm air but a 'hard frost' at 10 m.p.h. The exact definition of each term is best shown by a diagram (Fig. 74). For the 'radiation' frosts that occur at night under clear skies, calm is a pre-

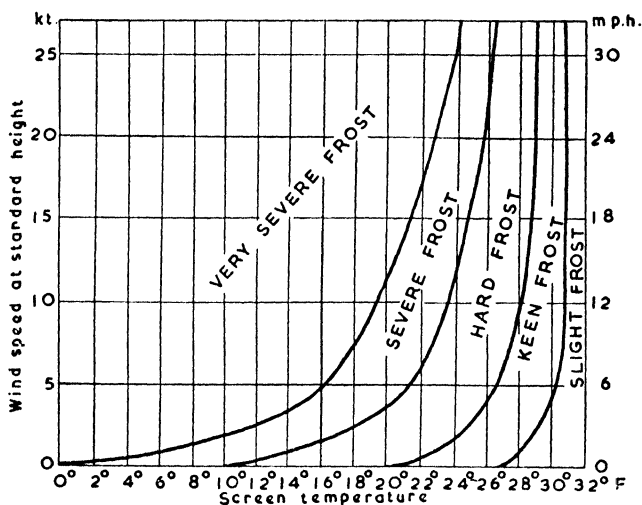


Fig. 74. Classification of frost diagram

requisite: but persistent day-and-night frosts ('advection' or 'wind' frosts) due to very cold (Arctic or Pc) air are often accompanied by strong winds.

The farmer is not worried about winter frosts, in fact these aid the plough in breaking up the soil. He most fears the late spring frost that catches young plants at a susceptible time after earlier clement weather has encouraged some growth. Fruit crops are especially sensitive to late April and May frosts. Frost is a nuisance to others besides farmers and fruit-growers. It means much extra work for railwaymen for anti-freeze paste or powder must be applied to the bearings of outdoor mechanism. It used to bring to a standstill all building operations involving the casting of concrete, but the development of new anti-freeze mixtures here too makes the builder more independent of adverse weather.

Freezing weather presents the motorist with a number of difficulties and anxieties; mist or fog may freeze on the windscreen so that he can no longer see through it clearly, ice may form in the carburettor or petrol pipe, but worst of all the water may freeze in the radiator or cylinder jacket and, expanding, crack them, necessitating a very expensive repair. Even when the engine is working the radiator may freeze up if the air

driven through is very cold and this blocks the circulation, so that the water round the cylinder heads boils while the bottom is simultaneously frozen. Wise motorists reduce the area of the cooling surface by blanking off the bottom of the radiator with a piece of cardboard or a radiator muff, but the remedy for all these ills is an anti-freeze mixture (generally ethylene glycol) which lowers the freezing point of the coolant to an extent depending on the concentration. The motorist who uses it in accordance with instructions can enjoy unworried sleep though his car stands all night in the open, and a little rubbed on the windscreen (glycerine will do) will delay ice formation on the windscreen for quite a long time even in freezing fog.

Not least, our traditionally bad plumbing systems make severe and prolonged wind frosts a headache to householders (though a boon to plumbers). The sensible householder, however, lags water pipes in the loft, ensures that no taps are left dripping (since this is the surest way of blocking a waste-pipe with ice) and drains the tank if he is leaving the house unoccupied during cold weather. New plastic linings for water pipes have been designed, which take up the expansion pressure when water freezes (just as a large cork will do if left floating in the water tank). Even more desirable, of course (and perfectly feasible, since countries with severer climates than ours have achieved it), would be a house design that makes these precautions unnecessary.

There are three lines of counter-attack in the battle against destructive frost. It can be forecast, it can be minimised by protective measures, or it can be forestalled by planning beforehand. Most forecasters have their favourite formulae for predicting night minima and the possibilities of frost. Anyone with a reasonably exposed thermometer can build up one of these empirically from past records, remembering that they have only local application and are not infallible. Table 6 for Reading, was constructed from values of the temperature drop from maximum to succeed-

TABLE 6

Expected Fall of Temperature (Maximum minus Minimum) under Various Conditions (Reading)

SKY	WIND						
	<i>Calm or Light</i> (< 7 m.p.h.)			<i>Moderate</i> (7-18 m.p.h.)			<i>Fresh</i> (> 18 m.p.h.)
	<i>Autumn</i>	<i>Winter</i>	<i>Spring</i>	<i>Autumn</i>	<i>Winter</i>	<i>Spring</i>	<i>All-Year</i>
<i>Clear</i>	20	15	23	15	11	18	4
<i>Partly</i>	15	11	18	10	8	11	4
<i>Cloudy</i>	9	7	11	6	5	6	4

Note: Autumn = Sept. - Nov., Winter = Dec. - Feb., Spring = Mar. - May.

ing minimum averaged for different nocturnal weather conditions. To obtain the minimum, simply deduct the appropriate figure from the maximum (or, say, 3 p.m.) temperature, but it is still, of course, necessary to make an estimate of the expected cloudiness and wind.

The risk of radiation frost exists in Polar air from autumn to spring, whenever evening skies clear and the wind drops. Frost warnings are given prominence in all official forecasts.

Once forewarned, the fruit grower in particular can take precautionary measures. These seek either to prevent heat loss at night or to add heat artificially. The first category includes the use of protective coverings—the gardener's usual method. Artificial smoke canopies (produced from oil burners) are less popular now than they were: they are too easily shifted by light winds. Even flooding with water has been tried but only a few plants take kindly to this treatment. Among methods of adding heat are continuous spraying with water (the latent heat released as this water freezes on leaves and stalks keeps the temperature around 32° F.) and bringing down warmer air from above through turbulence created artificially by giant electric fans. Most effective is direct heating by strategically placed burners: this is a sound proposition where fuel is cheap (oil in California, brown coal and briquettes in Germany) but costly in Britain.

The best way to protect fruit is by careful choice of orchard sites in the first place. Unfortunately, in many cases it is too late. The battle against frost has already been half-lost if the orchard lies in a frost-hollow.

Snowfall and Avalanches

Despite our addiction to Christmas cards portraying snow scenes, snowfall makes up only a small proportion of our total precipitation outside highland regions. It takes 10 or 12 inches of snow to provide the equivalent of 1 inch of rain. Children have scant opportunity of snowballing in low-lying parts of the south-west, where snow falls on only about 5 days a year and much of this amounts to no more than a few flakes. The chances increase eastwards (10–15 snow days on average in south-east England) and northwards (30–35 in northernmost Scotland). The number of snow days increases rapidly with elevation: highland villages may have 40 or 50. Snow rarely lies long in the south but persists for 100 days or more in parts of the Scottish Highlands. A persistent snow-bed in a deep north-east facing corrie below the summit of Ben Nevis was once thought to be permanent—until it disappeared in 1935.

Snow is not common at Christmastime, although the heavy snow-storm of Christmas 1927 is on record in support of the Christmas cards. It is surprising how many places have most snow days in March or at least as many then as in January or February. April, too, can provide quite a

severe snowstorm, like the one that brought over 6 inches of snow in parts of southern England on the night of 25th–26th April, 1950.

A good fall of snow is a huge delight to the children but may be a great nuisance to the community as a whole. It places an extra burden on the municipal authorities, who must clear snow (by laying salt) from city streets and sprinkle gravel where frozen compacted snow makes roads especially perilous. Drifts may be piled up (sometimes 10 or even 20 feet deep—see Plate XXI) by the wind, blocking roads and railway lines (especially in cuttings). In countries more accustomed to heavy snowfall—and along some of our own trans-Pennine tracks—snow-fences are used, but these must be carefully sited, not necessarily alongside the lines or the eddies they themselves set up may increase drifting. Visibility during a heavy snowfall is practically nil and this, too, adds to the hazards of transport. Lonely villages may be cut off for days and losses among sheep on hill farms are often grievous.

On the whole, however, snow affects us relatively little at least in lowland Britain. In other climates, the life of the community is adjusted to it, in ways of transport, in the style of dwellings and so on. In high mountain regions, snow converts the accessible slopes into winter sportsgrounds. Mountain dwellers live with the perennial danger of *avalanches* (Plate XXII). These occur usually when unconsolidated powdery snow accumulates rapidly in great quantity and is displaced by strong winds, or when a rise of temperature half-melts and loosens the snow masses. A medium-sized avalanche releases something like 2,000 tons of snow (three times as heavy if wet) in less than a minute: combined with the displacement of air (the ‘avalanche wind’) the destructive force is terrifying.

Aircraft Icing

In Chapter 3 we pointed out that many cloud droplets are in the supercooled state. This is a barely stable condition. In the laboratory, water can be supercooled, without ice forming, as long as it is not disturbed: if it is agitated, freezing is instantaneous. Similarly, in the atmosphere, if supercooled droplets are disturbed, as by a moving aircraft, freezing takes place on it. This is the root of the serious problem of aircraft icing. Ice may form on wings, propeller blades, windshield or in the carburettor intake. The general effect is to add weight and so alter the aerodynamic properties of the machine that its flight is impeded, or, in extreme cases, prevented altogether.

Icing is partially countered by the use of de-icing fluids or pastes, or by a device for pulsating the leading edge of the wing so that the ice does not adhere, or by applying engine heat through some kind of air or liquid circulation. At the same time pilots are taught what action to take in various

icing situations. Icing may be severe in large cumulus and cumulonimbus and thick frontal cloud. An aircraft is also in danger when flying in cold, freezing air into which rain is falling from warm front cloud above: this freezes rapidly on to the cold aircraft as glaze.

Fog, Smog and Smoke Abatement

Most inland districts get 40 or more days with fog in an average year, though usually only between 5 and 10 of these are 'thick fogs' (visibility less than 220 yards) seriously disrupting road and rail traffic. Inland fogs are mainly of radiation type forming in conditions already described (Chapters 4 and 9) usually forecast with confidence. The forecaster has to decide whether his predicted night minimum temperature will or will not be lower than the likely dew-point (or 'fog-point'). The only practical way yet devised of dispersing natural fog is to evaporate it by warming. This enormously expensive method (known as FIDO—Fog Intensive Dispersal Of!) was resorted to on certain British airfields during the war, when lines of oil burners were lit along runways for emergency landings.

There is nothing natural about smog. It is a man-made evil, a direct result of air pollution, something we have meekly accepted as a by-product of nineteenth century industrialisation. We have tolerated it far too long. The case against pollution has been argued strongly and persistently by the National Society for Clean Air (a voluntary organisation which deserves more support than it gets) but it took a major smog disaster to shock us as a community into at least considering an attack on the pollution menace.

The first indictment is on the count of health. Air pollution partly accounts for the pallor of the townsman and at the same time blackens his lungs. By blotting out the sunshine, it has helped in the past to produce diseases like rickets in young children. It is considered by competent authorities to be at least a contributory factor in respiratory ailments like sinusitis, catarrh and bronchitis, and is perhaps as responsible as cigarette smoking for the genesis of lung cancer. These long-term, slow-acting influences are reflected (with others) in a definite excess of urban over rural mortality in this country.

The menace to health is periodically highlighted by a killing smog. There is nothing new about these but only recently have they been subjected to close investigation. In December, 1930, a 5-day fog trapped beneath a low inversion in the narrow, highly industrialised Meuse Valley, Belgium, caused 63 deaths and many illnesses. In the blasting and smelting town of Donora, Pennsylvania, under similar physical conditions, 20 people died and half the population of 14,000 became ill through a 4-day smog in October, 1948. Then came the great London disaster of Decem-

ber, 1952, with its 4,000 victims (Fig. 75). Enquiries showed that most deaths were caused by diseases like bronchitis and pneumonia, mainly among older people already suffering from respiratory disorders. They were precipitated by irritant substances like the oxides of sulphur known to have been present in the air in high concentrations. Animals suffer too,

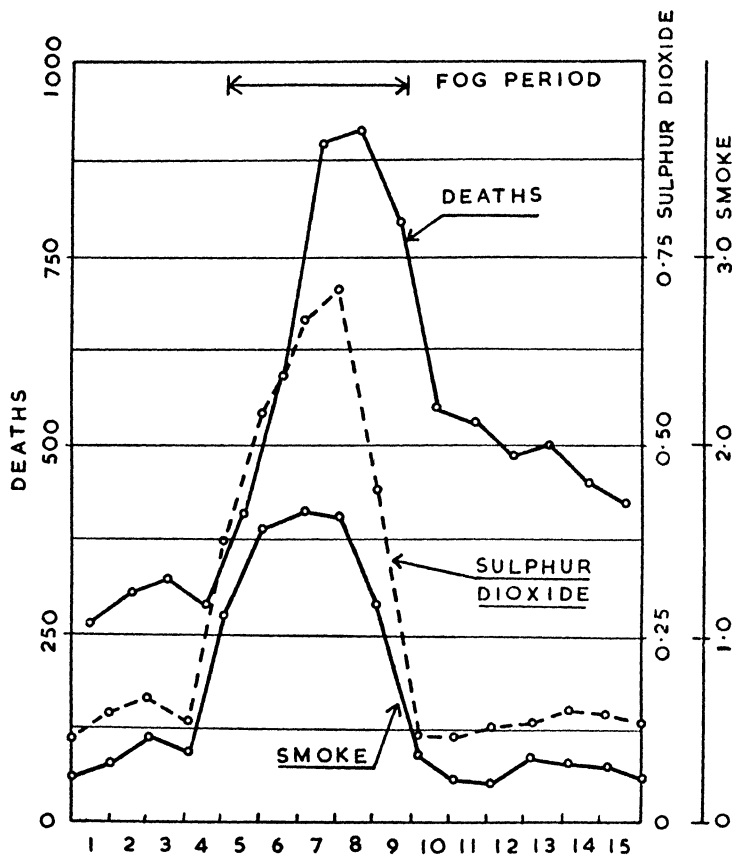


Fig. 75. Smoke intensity and smog deaths in London during the first 15 days of December, 1952

as was demonstrated by the casualties among prize cattle at the Dairy Show (which coincided with the 1952 London smog).

Polluted air is also injurious to plant life: evergreens suffer most and there are few coniferous trees in the central areas of London, Leeds or Manchester. Since the town spreads its aerial sewage with the wind, crops in rural areas to leeward may also be affected. There is also perpetual damage to materials of many kinds. The sulphuric acid in the air attacks

building stones, especially limestone, as the semi-permanent scaffolding round several of London's public buildings testifies.

The community suffers all this and pays for it too. We pay directly in extra bills for laundry and domestic cleaning, in more frequent repair and redecoration of buildings, in protection or replacement of materials, in extra lighting and medical services: we pay indirectly in the reduced efficiency of agriculture and transport. We pay too for an enormous waste of coal which we cannot afford. After careful consideration, the Committee on Air Pollution which reported in 1954 (the 'Beaver Report'), set the total cost as between £275 and £300 million per year: this works out to between £5 and £6 per annum per head of population—including children!

There is no doubt that pollution can be banished. St. Louis, described as a 'city of decay' in 1940, was a clean city ten years later. Pittsburgh is another transformed American city. The advance towards clean air demands, among other things, the improvement of furnace and boiler plant, the widespread availability of smokeless fuels and the encouragement of smoke-free domestic heating appliances. The spread of 'smokeless zones' is an important step forward, both material and psychological, while, in the planning of new towns, provision for district heating and the careful siting of industry (in relation to winds) ensure a good start. Some of these proposals were advocated in the Beaver Report, and a number later embodied in the Clean Air Act.

Weather in Wartime

During war our values regarding weather are apt to be strangely inverted. A clear night is good weather for marauding bombers, bad for their target. For a vulnerable town, low cloud or fog is relative protection. Some towns, during the last war, were intentionally hidden by artificial smogs produced by a battery of oil-burners. Sea fog in the English Channel allowed the German battleships *Scharnhorst* and *Gneisenau* to slip through under the very noses of the British Fleet and Air Forces. Bad weather, by grounding aircraft and bogging down tanks and heavy equipment, gives the advantage to light forces, especially guerillas, as so many 'colonial' wars have shown. Polar Continental air was the staunchest ally of the Russians against Napoleon's army in 1812 and the Germans in 1941. Gas operations during the First World War hung precariously on the vagaries of the surface wind. Today, not only jet aircraft but atomic warfare in general and the problem of radio-active fall-out in particular depend greatly on the upper winds.

The forecast problem intensifies in wartime. The full story of the achievements of forecasting personnel during the last war has yet to be

told but C. K. M. Douglas (himself a distinguished forecaster) has described some of the preparations for the D-Day landings in 1944. The problem, jointly tackled by the Meteorological Office, the Admiralty and the United States Air Forces Weather Service, was to find a short period of relatively quiet winds and suitable cloud conditions for bombing to coincide with suitable tides, during a generally unsettled summer season. Postponed from the original date (5th June) because of the passage of a cold front, the invasion was fixed for the following day when a small ridge of high pressure was expected over the area. This in fact gave the required fair period for the landings, but the margin was so tight that the Germans were taken by surprise.

The Conscious Control of Weather

The foregoing pages have reminded us of some of the ways in which weather affects our lives. It is just as true that we, in our complex twentieth century activities, affect the weather, though at present only on a small scale. Sometimes this is unintentional: the pollution of the air and the creation of smog are cases in point. Another is the formation of *vapour trails* by high-flying aircraft when, under certain conditions, the moisture in the exhaust from the engine freezes into ice-crystal cloud. This is true man-made cirrus: 'sky-writing,' once a popular means of advertisement, is different, being at comparatively low levels and using artificial smokes. A nuclear explosion also creates weather, in the form of the now-familiar mushroom cloud and heavy local precipitation, which is incidental to its main purpose. (There is no evidence, however, in spite of popular support for this belief, that atomic tests cause bad or unusual weather outside the immediate locality.)

At the same time we are all consciously creating 'weather' for one purpose or another. We retreat from inclement out-door weather and produce tolerable conditions of temperature and air movement within the four walls of a house (not always very successfully). The sensible housewife who is not blessed with a refrigerator leaves her milk or butter inside an earthenware container kept saturated with water, thus making use of the cooling effect of evaporation. Weather is controlled to some extent in a greenhouse, where the long-wave outgoing radiation is trapped by the glass, and to an almost complete extent in an artificially-heated, air-conditioned greenhouse. Double glass walls, or glass walls lined with cellophane, add to the heat-conserving properties of greenhouses (because of the insulating character of the air space between), just as double windows check the heat loss from houses in cold climates.

A good example of weather control on the farm is the planting of shelter belts and windbreaks. These afford valuable protection for livestock, crops

and buildings and help to prevent soil erosion by checking the blowing of light soils. Cutting down the wind speed means more than reducing mechanical damage to crops, it also keeps temperatures a little higher and slows down evaporation and transpiration. A suitable shelter belt confers these benefits effectively to a distance of up to 20 times the height of the barrier in the downwind direction and also to a much smaller extent upwind.

The Ministry of Agriculture and Fisheries and the Agricultural Branch of the Meteorological Office will give advice on the right sort and lay-out of sheltering trees or hedges. These must not be too dense or they defeat their own purpose by setting up eddies with damaging downdraughts on the lee side: walls are not advisable for the same reason. An effective windbreak is of medium density, allowing some air to filter through. Shelter belts should be planted at right angles not necessarily to the prevailing wind but to the damaging wind: this may be north-easterly in eastern England or canalised along a valley.

Rain-Making

The prospect of increasing the natural rainfall offers the greatest challenge so far to the would-be weather controllers. There is nothing new in the idea of rain-making for in primitive societies this is a routine task performed by the tribal medicine man. But modern rain-makers appeal to meteorological rather than mystical principles and probably with more success. Scientific experiments began in the United States in 1946 ('Project Cirrus') under Doctors Irving Langmuir and Vincent Schaefer) and soon after in Australia. Work has continued in those countries and begun in others, mainly in semi-arid regions. Our own Meteorological Office, mindful of the rainfall shortage in the eastern counties, has recently started a pilot scheme.

As a result of this early work, Bowen in Australia has concluded that a 5 to 10 per cent increase in annual rainfall is feasible. In America, Irving Krick and his associates have claimed very substantial increases especially in mountainous regions. There, too, an Advisory Committee on Weather Control submitted an interim report in February, 1956, and seemed satisfied that excesses of actual over expected rainfall of up to 17 per cent had been achieved by rain-making projects.

Rain cannot, of course, be conjured out of a cloudless sky, nor will every cloud be induced to part with its moisture. We saw in Chapter 3 that, as regards most rainfall in temperate regions, only where ice particles and supercooled water droplets exist together in the cloud can the Bergeron process operate. A cloud extending to these levels is likely to rain of its own accord. A cloud not reaching freezing level will not rain



PLATE XXI. A record snow drift at Barras railway station, Westmorland, April, 1947. The two men are standing on drifted snow 15 or 20 feet above the railway line



PLATE XXII. An avalanche near Davos, Switzerland. The detachment areas can be seen on the upper slopes on the left of the picture

at all. But a cloud that pushes above freezing level (*i.e.* contains super-cooled droplets) but not high enough for the formation of ice crystals has possibilities, provided it can be 'seeded' with freezing nuclei of some kind.

The first attempts at 'seeding' consisted of dropping pellets of solid carbon dioxide ('dry ice') into suitable cumulus clouds from aircraft. This very cold substance cools the cloud so much (to below -40° F.) that many ice crystals form spontaneously and the Bergeron mechanism can then operate. Much more effective as freezing nuclei are crystals of silver iodide, which induce freezing at temperatures as high as 23° F. This too may be dropped from aircraft but it has proved more economical to generate the chemical on the ground in the form of vapour (by burning fuel impregnated with a silver iodide solution) and rely on natural up-currents to carry it into the cloud.

Other seeding methods are necessary for the 'warm clouds' of tropical regions where coagulation is the main rain-producing process (Chapter 3). Here the problem is to provide a promisingly thick cloud with large droplets or large condensation nuclei. This was done with some success in Australia by spraying water into the base of the cloud from an aircraft. In Pakistan, salt solution has been sprayed into clouds both from the air and from ground sprayers. Another method of injecting clouds with large nuclei has been tried in Tanganyika (in order to maintain reservoir levels) where naval rockets, modified to incorporate salt particles, have been fired to burst and scatter their contents at suitable heights.

Various difficulties face the rain-makers. Not least is that of being sure that a seeded cloud would not have grown sufficiently to have precipitated of its own accord. However, there is impressive evidence in many cases that seeded clouds have developed rapidly and given rain while others not so treated have remained unchanged. When silver iodide is released from ground generators it is not certain in what concentration and condition it finally reaches the clouds. The Meteorological Office project aims to tackle this problem. It needs careful statistical checks of rainfall over a long period of seeding operations before their effects can be assessed. Nevertheless, results to date are encouraging and it seems that the rain-makers will be able to coax enough extra rainfall from the clouds to be valuable for thirsty crops, for replenishing reservoirs or for increasing river flow for purposes of irrigation or hydro-electric power.

Defence Against Hail

Paradoxically, rain-making techniques have recently been used for the prevention of another type of precipitation—hail. In the past, anti-hail measures have relied on freezing the supercooled droplets of a cumulonimbus cloud into many small ice particles which cannot then contribute

to the growth of large hailstones. This has been done by disturbing the droplets through an explosion. 'Hail-shooting' with cannon was practiced in the French vineyard region of the Beaujolais long before the hail-forming mechanism was properly understood. Later, rockets were tried and even the bombing of the supercooled layer from aircraft.

Now the potential hail-cloud is induced to yield its moisture before hail has time to form. A new technique of seeding with silver iodide was introduced in the Beaujolais in 1950. The chemical is incorporated into the explosive of rockets fired to burst in the supercooled part of the cloud. A para-military organisation, involving mobile rocket guns, and the closest contact with the meteorologists are necessary parts of the technique.

THE WEATHER IN HARNESS

The story goes that the idea of his engine came to James Watt when he saw the steam lift the lid of his mother's kettle boiling on the hob. What followed from Watt's invention we all know. But the man who first thought of harnessing the energy of the weather is unknown and unsung: human memory does not go back far enough to record who first rigged a sail to drive his ship or fixed a wheel in the flowing river to grind his corn. The energy of the weather is so apparent and its power so enormous that it is at first sight surprising that after this early and anonymous start so little progress has been made and so little of man's ingenuity has been exercised in harnessing natural forces. Watching a thunderstorm display at night, seeing a great river running wild in flood, listening to the wild shriek of the wind in an Atlantic gale and feeling the great liner heel over with its pressure on the hull—these are experiences that fill us with awe at the spectacle of uncontrollable power unleashed. But look at the sails of an old-fashioned windmill leisurely turning the heavy mill-stones with giant strength or listen to the faithful and contented hum of the dynamos at a hydro-electric power station and reflect on the mighty work being done for us by that infinitesimal fraction of meteorological energy that we have drawn off and harnessed for useful ends. And remember it is a gift.

Energy, Income, Capital

Watt's discovery had immediate results and led to the Age of Steam, and the invention of the Internal Combustion Engine led even more swiftly to the Motor Age. Both inventions created a rapid demand for mineral fuels, coal, and oil, both of which are exhaustible and their eventual exhaustion should cause more anxiety than it does. To be sure we find new coal and oil fields and we invent new techniques for more economical use, but this does not alter the fact that we are burning up a rich capital endowment of fuel reserves that have been stored in the earth for millions of years and are only being renewed very, very slowly. We have burnt more coal in the first half of this century than in all preceding centuries put together and the consumption of mineral oil in 1956 was 3½ times that in 1936, which was 5 times that in 1916; and in 1896, when the

internal combustion engine was in its infancy they hardly knew what to do with the volatile fraction and burnt it for safety.

By contrast with this squandering of capital resources we have made all too little progress towards taking advantage of the regular income of energy that comes to us in a steady stream from the sun and which will keep coming unabated for as long as we continue to inhabit this planet.

Solar Energy

The sun turns towards us a shining disc whose diameter is 864,000 miles. Its area is 2,500,000 square miles, each square centimetre shining with a light of 50,000 candlepower. Although our little planet receives only one 2,000 millionth of this yet the average amount is calculated to be equivalent to 5 million horsepower per square mile of surface. The problem of economic use is one of conversion of radiant energy into utilisable forms such as heat or electricity, and it is to such problems that the ingenuity of engineers must be bent. However, the endowment is so extravagantly generous that a minute fraction of the total is enough for our needs and only a very low efficiency conversion is called for. Fortunately, too, some of this conversion is already effected by nature, for as was explained in Chapter 3 solar energy raises the temperature, causes the wind, evaporates water and raises it into rain clouds. In fact the sun is the prime mover in all meteorological phenomena and each of these indirect manifestations of its power constitutes a usable source of energy—heat, wind power and water power, which are, in fact, the most immediately promising and most fully realised at the present day. We will deal with each in turn.

Heat Energy

Various devices have been tried for collecting and concentrating the form of the sun's direct radiant energy—mirrors, lenses, etc. At Mt. Louis in the Pyrenees there is an experimental sun furnace which, by means of two large concave mirrors, magnifies sun energy 20,000 times, raising a temperature of 3,000° C. at the focal point. But it is only at a small point. There is no great quantity of heat present, only that amount which originally fell upon the mirrors.

A much more profitable and promising line is the proper use, and subsequent 'magnification' if necessary, of the 'low temperature heat' that is being generated by the impact of solar energy wherever it is felt all over the earth's surface. Of course we use it and enjoy it as it comes, but little conscious attempt is made to store the surplus which is lost by radiation during the night or in the long winter months when outward radiation into

space exceeds the income, so that temperature falls. The crux of the problem, therefore, is how to check the nocturnal and winter leakage and to conserve as much as we can.

Heat Storage

Heat is harder to hold than water, but it can be stored. Only we cannot see it and perhaps that is why so little conscious attention is given to the problem of its conservation which is, by its nature, more subtle than the problem of water storage. The reasons for the fall of temperature by radiation loss at night and in winter have been explained in Chapter 3. The air itself cannot retain heat; it takes its temperature from its surroundings, and heat loss occurs mainly from radiation surfaces. So the solution lies in deep storage within heat-retaining substances such as earth and water. As air temperature varies both seasonally and diurnally, there are two distinct time-scales involved and two distinct problems, (1) to collect day-time warmth for use at night and (2) to conserve summer heat for use in winter.

A good illustration of 12-hour heat retention is provided by the traditional design of houses in desert climates in which the diurnal range of air temperature is large. Windows are made small to allow little sun to penetrate by day, keeping the interior dark and cool; walls are thick to absorb the day-time heat, pass it slowly through to the interior and give it out into the house throughout the night. An elaboration of this device is the 'heat trap' consisting of a box of gravel or crushed rock covered with glass. Solar radiation passes freely through the glass and is converted to long-wave heat which warms the gravel. At night air is drawn through the hot gravel and goes on to warm the house. Another device makes use of the heat of fusion of chemicals. Sodium sulphate melts at 90° F. (a temperature at which human beings begin to suffer real discomfort) and in the process absorbs 100 British Thermal Units per lb. By night the chemical re-solidifies, paying back the 100 borrowed B.Th.U.s to warm the house.

Heat Reservoirs

If the water-board cuts off our domestic supply for a few hours while some repairs are made we can manage with what is in the tank; similarly the devices just described may enable us to do without sun-heat for the night hours. But if there is a long drought the water-board has to draw on the rivers and reservoirs to meet our needs; similarly the retention of summer heat for use in the winter months demands reservoirs of enormous heat storage capacity. Nothing of smaller capacity than the earth beneath us or large water bodies such as the sea, lakes or large rivers will do. Enough heat has to be stored to last for six months or so.

The earth has great reserves of heat at great depth. On the average the temperature in the earth's crust rises 1° F. for each 60 feet we go down and the rock temperature in deep mines makes the air in contact so hot that working conditions can only be kept tolerable by elaborate ventilation systems. In some volcanic regions temperatures just below ground are too hot to bear and water, trickling in, boils up in geysers and hot springs. This volcanic heat, which is used as a source of power in Italy, Iceland and the volcanic North Island of New Zealand, does not concern us here as its source is not meteorological, but the ground surface wherever the sun's rays fall on it is warmed and some of this warmth soaks slowly into the ground: this is true meteorological heat.

When we were considering the heat balance of the atmosphere in Chapter 3 we recognised that during the summer months more energy was received by the surface by day than was radiated away at night. The mean temperature of the surface thus rises slowly but steadily as summer goes on. The earth thermometers (at 1 ft., 2 ft., and 4 ft. down) show by their steady rise that a wave of warmth is slowly penetrating into the soil. The greater the depth the slower the rise and the longer it takes to travel down, so that the maximum temperature is not recorded by the 4-foot thermometer until August or September. By this time the surface temperature has already been falling for two months, initiating a cold wave that travels downwards in pursuit of the warm wave, causing the minimum reading at 4 feet in February or March.

Few 'meteorological' readings are kept at great depth, but those that are show quite clearly that even in middle latitudes there is very little seasonal variation of earth temperature below 20 feet. At these depths, curiously enough, the soil is actually at its warmest in winter and coldest in summer, for the time taken for the downward transmission is just about six months. Herein, of course, lies the explanation of the uniform temperature of deep cellars, familiar in those used for the storage of wines and cheeses. We should waste none of our sympathy for those of our Ice-Age ancestors who lived in deep caves in warm limestone rock. In domestic architecture today, especially in countries with cold winters, it is becoming more and more the practice to build houses upside-down with the bedrooms underground. They are much warmer in winter than the conventional first-floor bedrooms round which the winter wind may blow at a temperature several degrees below freezing point, chilling the sleeper and freezing the wash-basin pipes. Even in the very cold winters of continental climates freezing does not penetrate very deeply into the soil. How deeply can easily be found out and water mains must, of course, be laid deeper than this level.

Underneath us, then, and easily available is a great reservoir of winter warmth and summer coolness, which, in climates that suffer from tem-

perature extremes, could save us a great deal of money now spent on central heating and air conditioning if means can be found of drawing on it.

Heat Exchangers

The most obvious way of tapping the warmth or cold (according to the season) would be simply to pass water in pipes at the necessary depth, but the temperature difference between depth and surface is too small to be very useful. It is therefore necessary to magnify the difference and a device is available—the heat pump, essentially a refrigerator working in reverse—by which the ‘low-temperature heat’ of the earth can be ‘boosted.’

But earth is not a very good conductor of heat, consequently the zone from which the heat is being taken may chill so rapidly that local freezing is caused before warmth can come in from the surroundings. This unfortunate happening can be delayed by using brine instead of water in the pipes and underground tanks. It is more usual to take the warmth from a large water body such as a river, in which the chilled water is quickly removed and replaced by new supplies of warm water from further upstream. For thanks to its high specific heat and heat retaining qualities water is much warmer in winter than the air that overlies it or the land that borders it. The Festival Hall on the South Bank is warmed in winter by a heat pump using the Thames, and, if necessary, can be cooled in summer by working the pump in reverse.

There would seem to be a considerable future for heat-exchangers of this type but they are still in the experimental stage and the theoretical benefits of ‘free’ heat are difficult to realise at present because of the initial cost of construction.

Transport and Redistribution of Heat

It is one of the cardinal principles of climate that seas and oceans are slow to warm up and slow to cool down; hence they exercise a moderating effect on the temperature of climates in their vicinity. This effect is due to the high specific heat of water, to the high albedo of its surface and to the waste of latent heat by evaporation from the surface and to the redistribution of heat in depth by the turbulent mixing of layers. But despite its low efficiency as a receptor of solar energy the sea, by virtue of its great surface area, stores a prodigious amount of warmth in its surface layers in low latitudes. And this heat is carried by ocean currents, surface drifts and the winds that blow over the surface into higher latitudes, thereby constituting a gigantic central heating plant that warms the whole earth.

Sea Temperature as an Energy Source

Water masses, like air masses, take their temperature from the climatic conditions at their source regions and where water masses of diverse origin are brought cheek-by-jowl, marked differences of temperature occur, as they do along a front in the atmospheric circulation.

Wherever there is a temperature gradient there is an energy potential and the steeper the gradient the greater the potential. Volcanic regions, already referred to, present obvious opportunities but temperature gradients in the ocean can make a contribution and an interesting experiment is going on at Abidjan on the Ivory Coast of West Africa, making indirect use of sun heat stored in ocean water. Two power plants, each of 3,500 kilowatt capacity, are using the temperature differences between the surface water of the ocean (at 80° F.) and cold bottom water (at 46° F.). The thermodynamic cycle consists in evaporating under vacuum warm surface water and inhaling the steam into a condenser which is cooled by cold bottom water; on its way the steam passes through a turbine that drives the electric generator. The company developing this pioneer plant is appropriately called 'Energie des mers,' for it is drawing upon the inexhaustible source of solar energy stored in the limitless ocean.

Wind Power

Driven by sun power the circulation of the atmosphere has components both in the horizontal and the vertical planes. It is the vertical circulation that lifts the water in the clouds, and endows us with water power. The horizontal circulation provides wind power which is much more difficult to harness on a large scale, because, unlike water which flows continuously downhill, confined in valleys by which it can be guided into the maw of great turbines, the wind blows when and where it wills. So wind-power can only be used at present in small units, tapping the energy passing through the swept circle of the sails or rotors of the windmill, while all the rest goes by to waste.

It is true that hills and valleys deflect and guide the flow of air and where the hills stand athwart the prevailing winds they stow it against the hill-sides, making the hill-tops the windiest places, suitable, for example, for the old-fashioned iron forges. And winds blowing up a narrow valley are confined and strengthened in the gaps at their heads, but the degree of concentration is relatively small. However, if the wind is to be used as a source of power it is a wise precaution to take every advantage of the best sites. The factors governing the choice of site are of two types on two scales—zonal and local. The world map of wind direction and force gives the first clue. Some parts of the world are windier than others, the trade

wind belts and the westerlies offer promising conditions, the anticyclones and the calms of the Doldrums are unpromising. But within the zones of promising prospects it is the local factor that determines the best site. On general principles we may suggest the most favourable types of localities are to be found on islands and east-facing hilly coasts in the trade wind belt, on west-facing hills and coasts in the westerlies and on the open plains and prairies where the wind blows unobstructed. But in any case, after consulting the existing records of the nearest available meteorological station, more detailed site studies, based on long and continuous observations, are necessary preliminaries to the final choice of site for the harnessing of wind power.

It is obvious that wind power, even in the best sites, is irregular and unreliable and inefficient. It is easily outmatched by other sources of power, if available. Its future would therefore seem to be brightest where alternative sources are expensive or unobtainable. In the Arctic or Antarctic, in deserts and steppes, on lonely islands and in such inaccessible places wind power may come into its own again, this time not as a direct drive of machinery but as a generator of electricity.

The Storage of Wind Power

The wind comes in fits and starts, so the storage of wind energy is a matter of always keeping enough ready cash, so to speak, in one's pocket to meet the day-to-day expenditure. Development must be limited for want of a truly satisfactory form of collection and storage of surplus energy accumulated during windy spells to tide over the inevitable calm spells. One device is to pump water into high-level storage tanks and this device has done good service in the past, but now that the present trend is to use wind to make electricity it would seem logical to 'accumulate' the electricity in 'accumulators.' This calls for research into suitable electrodes and electrolytes and for the improvement of batteries.

Water Power

The sun's heat evaporates surface water and raises the vapour into clouds from which rain and snow fall on hills and mountains. The form of the land leads this water into valleys which the rivers themselves have cut and thus the power is concentrated for our use. Our Saxon ancestors used the power of rivers to drive the mill wheels to grind their corn and in medieval times the rivers worked the tilt hammers for crushing rock and extracting ore and for working the forge bellows for the iron industry, etc. But these archaic methods are going out of use and nowadays we convert the water power into the more convenient and transmissible form

of hydro-electricity. Small installations in mountain streams still have their uses for lighting a few houses or a village or even for small industries, but the present tendency is for larger and larger scale works.

For giant installations two sets of factors must be favourable—a steep fall and a large and steady volume of water flowing. To some extent these factors are complementary to each other and in theory the same amount of power can be had from a large volume with small fall as from a small volume with large fall, the output being the product of the two. However, the former arrangement is much more convenient and the hydrographic factor is clearly the more important. This means that the regime of the rivers is of paramount importance and this in turn is mainly dependent on climates, especially on rainfall. Flashy little streams, quick to rise in spate and equally quick to run low in dry spells are most unsatisfactory, mainly for the reason that no efficient method yet exists for storing electricity in any quantity; it has to be used as it is made. But if we cannot store the electricity, we can at least store the water and this raises the question of storage to ensure regularity of flow and therefore of power output.

Rainfall and River Regimes and Water Storage

Rainfall fluctuates capriciously, but less disastrously than wind; our ways of using it, too, are less hand-to-mouth. If it stops raining we do not immediately run short of water; there is a reserve in the soil which is steadily delivered to the rivers which continue to flow for quite a while. But if the dry spell is prolonged we may begin to feel the consequences quite seriously and if this happens often we shall have to think about reservoir storage. From long-term records of rainfall we are able to calculate how often and for how long this state of affairs may be expected to occur and to plan for the worst. This is a wise precaution but the water engineer has to weigh the risks involved against the cost of reservoir construction as an insurance.

But quite apart from the chance fluctuations of weather (abnormal long spells of drought) the rainfall in many types of climate is very unevenly distributed throughout the year as described in Chapter 8. In Britain there is no dry season; in the London region, for example, each month has a normal expectation of about 2 inches of rain. But at Los Angeles, by contrast, no rain can be expected and certainly not relied upon between May and October. At Lahore there is a dry season lasting eight months from October to May and agriculture would come to a standstill were it not for the irrigation schemes which are man's response to the challenge of seasonal rainfall. Fortunately there is water coming down the great rivers of the Punjab from the snow and ice on the Himalaya and there is a contribution from slight precipitation along the foothills in the middle of

winter (see page 180), but the rivers would run very low and perhaps dry up altogether in the dry season, waiting for the main 'fill-dyke' which is the monsoon from June to September. By July the Indus is in full flood, a menace to banks and bridges.

How can we save the water income rising in excess of immediate needs for subsequent cashing when the funds run low in the dry season? Much attention has been given to water storage in tanks and reservoirs, by dams and barrages from ancient Egypt to the present day, but there is still much room for improvement. Floods and droughts tend to be treated as separate problems instead of complementary ones. We get rid of the flood with all possible haste and regret its loss when drought follows—an improvident and short-sighted policy even in a climate like Britain's with no regular dry season. The crux of the problem, of course, is where to store the water while we wait for the drought—where to find reservoir space without drowning valuable land or villages. And a concomitant problem is how to conserve the water in the reservoirs against evaporation loss, which, in a hot climate, may consume more than half of it. While the provision of more and more surface storage space must continue to exercise the minds of water engineers in these times of rapidly increasing consumption for agricultural, industrial and domestic uses, the possibilities of storage in the soil and in the rocks must be fully explored. Proper watershed and soil management in the gathering grounds and the re-charge of underground aquifers, many being exhausted by over-pumping, could conserve considerable quantities out of reach of evaporation. Some of these problems are dealt with in Chapter 11 (pages 230–231).

In the reservoirs themselves promising experiments have been made to check the evaporation by protecting the water surface with very thin films of floating oil. In Australian trials the use of cetyl alcohol, so thin as to be almost monomolecular, is said to have reduced evaporation losses by 30 per cent.

Enough Water for All

We cannot live without water. Early man went to live by the banks of rivers to satisfy his daily needs for his family, his flocks and his crops. Medieval woollen industries needed the water for scouring, washing and dyeing as well as its power for the mill wheels. But civilised man today expects the water to be brought to him, in pipes, and because it is so easy to turn on the tap he may forget his ultimate dependence for this cheap commodity on rain and rivers, *i.e.* on climate. For supplying building estates and even new towns it is generally assumed that water can be brought to the tap, nor do industries often feel the need, as in former

times, to go to the water—it will come to them. No doubt this new attitude of mind can be justified by the immense strides that have been made in the provision of piped supplies over the last hundred years, but this sort of outlook is creating a situation that threatens to place an intolerable strain on the present system. Droughts give us warnings and long dry spells, depleting the reservoirs, cause anxieties, but let it be clearly understood that the real menace is not diminishing supplies but increasing consumption. A crisis is approaching which will demand bold thinking and planning for the future on a much larger scale and of a quite different nature.

In England and Wales the consumption of water per head of population averaged 33 gallons per day in 1938; by 1948 it had risen to 40 and last year (1956) it had risen even more steeply to 46. In some areas the increase is even more alarming, *e.g.* Manchester 41 (1938), 53 (1948) and 60 (1956). But the heaviest users are the industries; Chemicals for example use 820 million gallons a day and will soon pass the 1,000 million mark. It is estimated that 600 tons of water are needed for every ton of coal consumed. Of course, much of this water can be used again, hot or cold, purified or unpurified, but we have not to look very far into the future to see a time when today's demand will be doubled.

In a book concerned with weather and climate we should not concern ourselves with this consumption aspect of water supply, but we are bound to examine how far the rainfall of Britain can be expected to meet the likely demands for all future time. The answer is that it would certainly suffice if we could use it all, for as a matter of simple arithmetic the average rainfall of England and Wales would provide 31 cubic miles of water a year—enough for 2,000 gallons a day for each man, woman and child of the population. This is a generous allowance even after subtracting evaporation losses, and we need not be unduly concerned with percolation which can be regarded not as loss but as use, for most of the water that soaks into the ground goes to nourish the vegetation or is returned to the rivers eventually. But although the total amount is satisfactory the distribution leaves much to be desired.

As a matter of geographical fact the places where the water is wanted are not the places where the climate provides it. Britain's heaviest rainfall is in the mountains of Wales, the Lake District and Scotland (see page 203); the heaviest demand is in the densely populated coalfields and regions of heavy industry in the foothill zone and increasingly in the industrialised lowlands of the south midlands and the London Basin—the driest parts of the country. Thus the re-deployment of our water resources is likely to become a question of increasingly vital importance. Water must be led from where it falls to where it is used. Rivers do this work to a large extent, rather inefficiently; pipe lines more efficiently but at some expense.

Transport and Distribution of Water

Rivers are the great distributors of the gift of rain. Gravity is the free force that makes them flow and only a very slight gradient is needed to keep the water moving. Often the gradient of a river, especially in hilly country, is extravagantly steep, so that water can be made to go further if it is led off along the hill-sides, as in mill-leats, creating in the process a head of water to drive the mill wheel. Irrigation canals are similarly led off along the backbones of spurs dividing tributary basins and from them tributary canals lead the fertilising water on to the fields. These devices are ubiquitous from China to Peru. And in canal systems the water channel is kept as high as possible up the hill-sides, but sooner or later when it is necessary to transfer water from one drainage basin to another, tunnelling through divides or raising the water by pumps may become an inescapable necessity.

In a small industrialised country like Britain with many river basins, each of them quite small, it is unlikely that the demand for water can neatly be met and balanced with the supply within each basin. Given a knowledge of the rainfall, qualified by an estimate of losses by evaporation and percolation, checked by gauging the rivers, the quantity of water available in each river basin is a value that can be known. The demand for water in the basin could also be estimated or planned. From these two calculations it should be possible to estimate the need for transfer from one region to another. The meteorological aspects of the problem are relatively simple; the legal and administrative difficulties are the stumbling block, but they must be overcome if population and industry are to continue to expand.

The cost of the works may be high, but water today is the cheapest, though one of the most valuable of commodities and, as has so often been said of other necessities, can we afford not to afford some form of water 'grid'?

Partnership with Weather

The saying 'knowledge is power,' though generally attributed to Francis Bacon more than three hundred years ago, only gives expression to a universal truth already said in Proverbs (xxiv. 5) 'a wise man is strong; yea, a man of knowledge increaseth strength.' Knowledge of weather increases our power a million-fold for it places a giant's strength behind our arm if we are wise enough to work in co-operation. To talk of controlling the weather is a foolish boast; even with atomic power at our command we are hopelessly outmatched by the stupendous energy of the atmospheric circulation deriving its authority from the sun. But while we can do little or nothing to divert the weather from its main purpose we can

deflect to our own advantage minute but still considerable portions of its strength, we can make the fullest use of the gifts it confers, and we can shelter ourselves from its inclemency and perils all the better when we know, from our studies, what to expect. Wayward the weather may be, it affects but need not control our lives; wisdom surely lies in making it our ally.

APPENDIX

Plotting Weather Maps from Daily Weather Report Data

When the *Daily Weather Report* again became available after the War it was a fairly simple matter to plot weather maps from the observations contained therein. Only a minimum of information (cloud types, for example) was given in code figures: wind direction was given by compass points and speed by Beaufort Force numbers, weather was described in Beaufort letters. But later, wind direction was given in degrees and speed in knots and then, since mid-1955, the system has been radically changed and the observations are now set out in completely coded form just as they appear on the teleprinter paper (see page 33). All this has made it progressively more difficult for the enthusiastic amateur to plot his own charts although there is no need for him to be deterred from so doing. Anyone who currently takes the *Daily Weather Report* will find an explanation of all codes and specifications in the Introduction which is sent out quarterly and this, together with a little booklet *Instructions for the Plotting of Weather Maps* (M.O. 515—obtainable from Her Majesty's Stationery Office), will answer all his decoding and plotting problems.

But for those who do not want to take this trouble, this appendix is supplied which, with a selection of codes and symbols, will enable them to interpret the examples of weather maps included in this book. The Beaufort Scale of Wind Force is reproduced and there is also a list of the more commonly used Beaufort Letters for weather.

Plotting Temperature-Height Diagrams

The plotting of these diagrams from the *Daily Aerological Record* also presents some problems. Earlier issues gave temperatures in °F. for pressure levels at 50 mb intervals and corresponding heights in feet were given for every 100 mb. Additional information was given concerning inversions and isothermal layers. From these data the curves are easily plotted. Later, however, corresponding heights in feet were given for only 10 pressure levels, widely separated, and the details of the ascent were shown only in terms of pressures. More recently there has been a change to the C.G.S. system, the 10 heights being given in metres and temperature in

LOW CLOUD C_L

<i>Code Figure</i>		
1		Cumulus (Cu)—little developed
2		Cumulus—considerably developed
3		Cumulonimbus (Cb)—without anvil
4		Stratocumulus (Sc)—formed from spreading of Cu
5		Stratocumulus—otherwise formed
6		Stratus (St) or Fractostratus (Fs) not of bad weather
7		Fractostratus ('scud')—under As or Ns
8		Cumulus with Stratocumulus
9		Cumulonimbus—with anvil

MEDIUM CLOUD C_M

<i>Code Figure</i>		
1		Altostratus (As)—thin, sun or moon visible
2		Altostratus, thick, or Nimbostratus (Ns)
3		Altostratus (Ac)—thin, unchanging
4		Altostratus—thin, in patches, continually changing
5		Altostratus—thin, in bands, or increasing
6		Altostratus—formed from spreading of Cu
7		Altostratus—thick or several layers or with As
8		Altostratus castellatus
9		Altostratus—several layers, chaotic sky, usually with Ci

HIGH CLOUD C_H

<i>Code Figure</i>		
1		Cirrus (Ci)—fine, not increasing
2		Cirrus—dense patches, not increasing
3		Cirrus—anvil-shaped, usually with Cb
4		Cirrus—tufted, increasing, thickening
5		Cirrus and/or Cirrostratus (Cs)—distant but increasing
6		Cirrus and/or Cirrostratus—closer and increasing
7		Cirrostratus—covering whole sky
8		Cirrostratus—not increasing, not continuous
9		Cirrocumulus (Cc)—alone or with Ci or Cs

PRESENT WEATHER

	DRIZZLE	RAIN	SNOW
Light Intermittent	•	•	*
Light continuous	••	••	* *
Moderate intermittent	• •	• •	* *
Moderate continuous	•• ••	•• ••	* * *
Heavy intermittent	• • •	• • •	* * *
Heavy continuous	•• •• ••	•• •• ••	* * * *
Precipitation within sight, distant from station			)•(
Precipitation within sight, near to but not at station			(•)

FOG

Mist	==	Fog in patches	==
	Thinning	No Change	Thickening
Fog: sky discernible	==	==	==
Thick fog	==	==	==

SHOWERS

• ▽	Slight rain	• ▽	Sleet
• ▽	Heavy rain	• ▽	Snow
• • ▽	Violent rain	△ ▽	Hail

THUNDERSTORMS

⚡	Slight or moderate
⚡	Heavy
•/⚡	With rain/snow
△ ⚡	With hail

Note: When precipitation, fog or thunderstorm has occurred in the past hour but not at the time of observation, the symbol is plotted to the right of the past weather symbol.

CLOUD AMOUNT

in oktas (eighths) of sky covered

- None
- ◐ 1 or less
- ◑ 2
- ◒ 3
- ◓ 4
- ◔ 5
- ◕ 6
- ◖ 7 or overcast with openings
- ◗ 8
- ⊗ 9 sky obscured

PAST WEATHER

- Fair
- ◐ Variable sky
- ◑ Cloudy
- ≡ Fog
- Drizzle
- Rain
- ✱ Snow or sleet
- ▽ Showers
- ⚡ Thunderstorm

HEIGHT OF CLOUD

Code

Figures

00 to 50 = cloud height in hundreds of feet

e.g. 02 = 200 feet

25 = 2,500 feet

51 to 55 — not used

56 to 80 — subtract 50 to get cloud height in thousands of feet

e.g. 58 = 8,000 feet

79 = 29,000 feet

81 = 35,000 feet

82 = 40,000 feet

83 = 45,000 feet, and so on, to

88 = 70,000 feet

89 = above 70,000 feet

FRONTS

Warm front



Cold front



Occluded front



Stationary front



Symbols are placed on that side of line towards which front is moving.

On M.O. working charts, warm fronts are shown by red lines, cold fronts by blue, occlusions by purple and stationary fronts by lines made up of alternate red and blue sections.

AMOUNT OF LOW CLOUD given in oktas.

DRY-BULB TEMPERATURE given in °F.

DEW-POINT given in °F.

BAROMETRIC PRESSURE given in mb: only the last 3 figures (including 1 decimal) are plotted, e.g. 98·7 for 998·7, 10·2 for 1010·2.

BAROMETRIC TENDENCY given in tenths of mb ± (Note: On M.O. working charts, symbols—the barometric *characteristic*—indicate in more detail the behaviour of the barometer during the preceding 3 hours.)

WIND. Arrows fly with the wind: half feathers represent 5 knots, full feathers 10 knots. Calm is indicated by circle surrounding station circle.

VISIBILITY

Code

Figures

00 to 50 = visibility in tenths of kilometres or half-furlongs, e.g. 21 = 2·1 km. (= 10½ furlongs)

51 to 55 — not used

56 to 89 — subtract 50 to get visibility in kilometres, e.g. 78 = 28 km.

BEAUFORT SCALE

<i>Force</i>	<i>Description</i>	<i>Specifications on land.</i>	<i>Specifications at sea.</i>	<i>m.p.h.</i>
0	Calm	Smoke rises vertically	Sea like mirror	< 1
1	Light air	Smoke drifts, but vanes remain still	Ripples form	1—3
2	Light breeze	Wind felt on face, leaves rustle, vane moves	Small wavelets, not breaking	4—7
3	Gentle breeze	Leaves and small twigs move constantly; light flag extended	Large wavelets, beginning to break, glassy foam, scattered white horses	8—12
4	Moderate breeze	Raises dust and loose paper, moves small branches	Small waves, fairly frequent white horses	13—18
5	Fresh breeze	Small trees in leaf begin to sway	Moderate waves, many white horses, perhaps spray	19—24
6	Strong breeze	Large branches move, telegraph wires whistle, umbrellas hard to control	Large waves, with extensive white foam crests; spray probable	25—31
7	Moderate gale	Offers some resistance to walkers: whole trees move	Sea heaps up, white foam begins to be blown down-wind	32—38
8	Fresh gale	Impedes progress: breaks twigs off trees	Moderately high waves: foam blown in well marked streaks	39—46
9	Strong gale	Removes roof slates and chimney pots	High waves, rolling sea, dense streaks of foam: spray may reduce visibility	47—54
10	Whole gale	Rare inland: trees uprooted, much structural damage occurs	Heavy rolling sea, very high waves with overhanging crests, great foam patches and dense streaks: much spray	55—63
11	Storm	Rare	Rare	64—72
12	Hurricane	Rare	Rare	73—82

SELECTION OF BEAUFORT LETTERS

Appearance of Sky

b blue sky, bc partly clouded, c cloudy, o overcast (uniform layer).

Precipitation

d drizzle, r rain, s snow, rs sleet, p (followed by r or s) shower, h hail.

Electrical Phenomena

l distant lightning, tl thunderstorm.

Atmospheric Obscurity

m mist, z haze, f fog, fe wet fog, F thick fog.

Ground Phenomena

w dew, x hoar-frost.

Combination of Beaufort Letters

Capital letter denotes intensity *e.g.* R = heavy rain.

Suffix subscript o denotes slight *e.g.* so = slight snow.

Letters repeated denotes continuity *e.g.* dd = continuous drizzle.

Prefix i denotes intermittent *e.g.* ir_o = intermittent slight rain.

Prefix j denotes weather near but not at the station.

Solidus / divides present weather from weather which occurred in preceding hour *e.g.*

r/tl = rain now but thunderstorm in last hour.

*C. Humidity, formerly expressed as relative humidity, has for some years now been given as dew-points.

If speed of plotting is important and only approximately accurate diagrams are sufficient, then the accompanying graph (Fig. 76), based on average conditions, will enable any pressure in millibars to be read off in round numbers of either feet or metres. It is necessary only to begin at the required pressure, move horizontally to meet the curve and then vertically, up for metres, down for feet.

If, however, accuracy is required, it will be necessary to draw for each ascent, with some care, a new curve relating pressure and height, based on the 10 pressure levels and corresponding heights that are given: the other significant pressures can then be converted into heights by reference to this curve.

Conversion Scales

Line conversion scales enabling rapid though approximate translation as between degrees Fahrenheit and degrees Centigrade, miles per hour and knots, millibars and inches of mercury, and miles and kilometres, are provided for the convenience of the reader in Fig. 77.

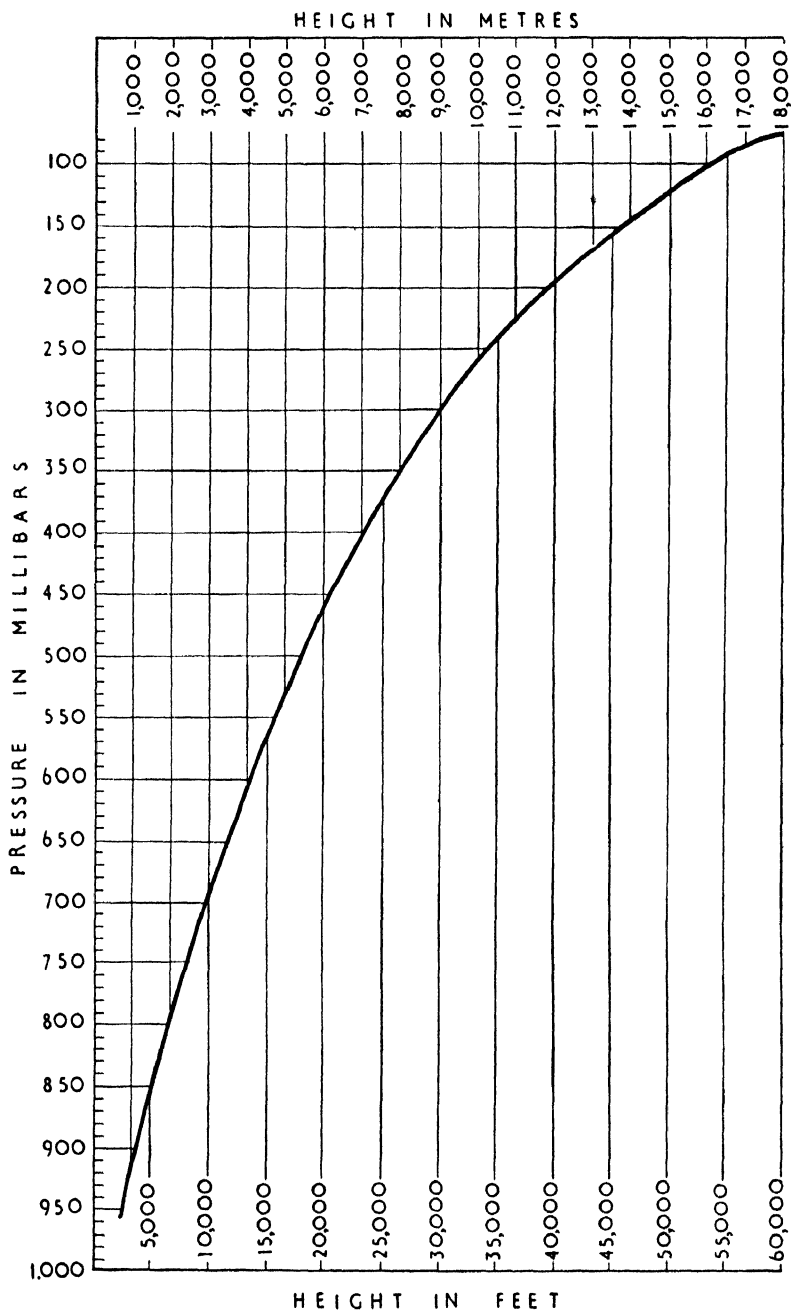


Fig. 76

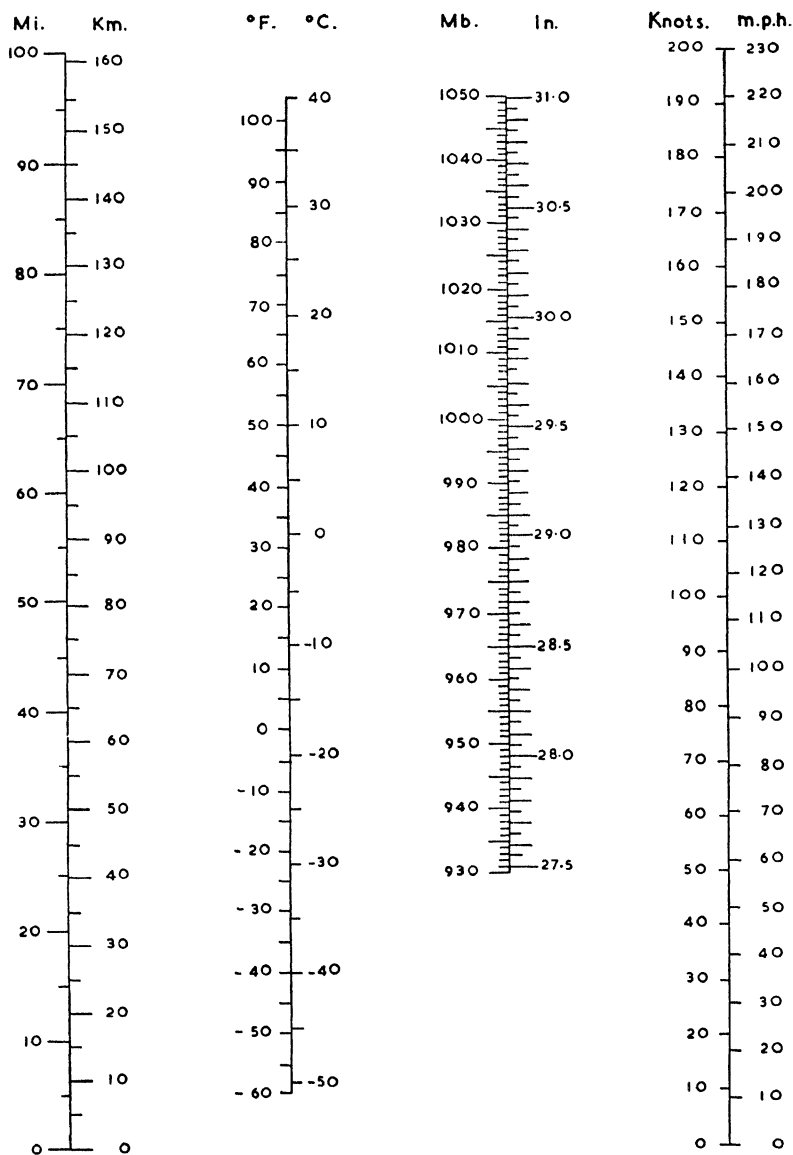


Fig. 77

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