PHYSICAL GEOGRAPHY

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Plate I

A. W. Clayden, *photo.*

Stratus.

PHYSICAL GEOGRAPHY

BY

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PREFACE

THERE are many books on Physical Geography in the English language, but most of them are intended for the use of junior classes in schools and are accordingly of a very elementary character. It is with the object of providing for the needs of somewhat older students that the present book has been written.

Certain preliminary matters which are often treated under the head of Physical Geography, but which do not properly belong to that subject, are omitted. It is assumed, for instance, that the reader understands the principles of the barometer and the thermometer, and that he has some acquaintance with the motions of the earth and the moon. But no one who has been through a general elementary course in science, such as is now given in most secondary schools, should meet with any difficulty.

A few words as to the arrangement adopted may be of use, especially to teachers who may wish to know how far the book will be serviceable in their own classes. The book falls naturally into three sections dealing respectively with the atmosphere, the ocean and the land. Logically the study of physical geography may begin either with the atmosphere or with the land, and it is mainly a matter of convenience which of these courses should be followed. The two branches of the subject are to so large an extent independent that, if desired, the section on the land may be read before the section on the atmosphere. But the study of the atmosphere should always precede that of the ocean.

In planning a course, however, there is one consideration that should be borne in mind. It is easy to arrange a useful series of laboratory lessons upon the atmosphere or the ocean; but in the case of the land comparatively little laboratory work is possible and the most valuable form of practical work consists of excursions in the field. Consequently there are advantages in dealing with the atmosphere during the winter months and with the land during the summer. But much depends upon the circumstances under which the course is given.

With regard to the order of the individual chapters, considerable latitude is permissible both in the section on the ocean and in the section on the land. But in the section on the atmosphere the mode of treatment requires that the chapters should be taken in the order in which they appear. According to my own experience the method adopted is the simplest and easiest for the student.

In a book of this kind the question of units is always one of some difficulty. In English-speaking countries climatological data are usually recorded in English measures, and I have accordingly used English measures throughout. But owing to the recent adoption of the C. G. S. system, except as regards temperature, in the Daily Weather Report of the Meteorological Office, I have added a short appendix upon the units employed in that publication.

For the photographic illustrations I have many friends to thank. Pls. I–IV are taken from the beautifully illustrated *Cloud Studies* of Mr A. W. Clayden, Principal of the Royal Albert Memorial College at Exeter, and for permission to reproduce them I am indebted to the author and his publisher, Mr John Murray. Pl. X, fig. 1, is from a photograph by Mr T. J. Roberts. The photographs of Himalayan glaciers, Pl. XII and Pl. XIII, fig. 1 were supplied by Mr T. D. La Touche, formerly of the Geological Survey of India. Pl. XIV, showing some of the Antarctic glaciers, I owe to my colleague at Cambridge, Mr C. S. Wright, who was with Captain Scott's last expedition. Pl. XVI, fig. 1 was given to me by Dr A. W. Rogers, Assistant Director of the South African Geological Survey. Pl. XVI, fig. 2 and Pl. XVII are from illustrations in *The Duab of Turkestan* by Mr W. R. Rickmers, published by the Cambridge University Press. Both of the figures on Pl. XVIII and fig. 2 on Pl. XIX are from photographs by the late Dr Tempest Anderson, whose photographic studies of volcanoes are so well known. For permission to reproduce them I am indebted to his executors. Two of these figures and also Pl. XIX, fig. 1 have already appeared in the *Text Book of Geology* by Mr Rastall and myself, and both Mr Rastall and our publisher, Mr Arnold, very kindly consented to the use of electros from the blocks prepared for that work. Pl. XX is from models in the Sedgwick Museum.

But most of all I am indebted to Prof. S. H. Reynolds of Bristol University, who supplied eleven of the photographs which appear in the plates. To him and to the other friends who have helped me I offer my heartiest thanks.

The maps of isobars and isotherms at the end of the volume are based chiefly upon the maps of Buchan, Mohn and the National Antarctic Expedition of 1901–1904.

PHILIP LAKE.

CAMBRIDGE,
9 *November*, 1914.

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CHAPTER I

INTRODUCTION. COMPOSITION OF THE ATMOSPHERE

Introduction. The planet upon which we live consists of a large core surrounded by two thin coverings or envelopes. The outer envelope is gaseous and is called the atmosphere. The inner envelope is liquid, and is known as the hydrosphere: it is formed by the oceans, but unlike the atmosphere it is incomplete and does not entirely cover the globe. The core itself is solid, at least externally, and is often called the lithosphere.

The specific gravity of the earth as a whole is 5·67 but the specific gravity of the rocks which form the outer part of the lithosphere averages less than 3. The interior, therefore, must be considerably denser than the exterior. It is, moreover, certain that in the interior the pressure must be very great and there is evidence that the temperature is also very high. The interior must, accordingly, be different from the exterior, and many writers limit the term lithosphere to the outer part of the earth, where the rocks are more or less similar to those which are visible at the surface, and separate the inner portion under the name bathysphere, centrosphere or barysphere.

Beyond the facts that the materials are dense, that the pressure is great and the temperature probably high, but little is known as to the nature of the bathysphere, and very different views have been held concerning its constitution. Some writers believe it to be solid; others imagine that it is molten; and some have even supposed that it may be gaseous, though owing to the enormous pressure it does not behave like a gas on the surface of the earth. Whatever the condition of the bathysphere may be, tidal observations show that the earth as a whole is as rigid as a solid ball of steel of the same size.

The geographer, however, is concerned with the surface of the earth and not with its interior, and in general he has only to consider the atmosphere, the hydrosphere and the visible portions of the lithosphere, or in other words, the air, the ocean and the land.

Owing to the effects of perspective vertical heights usually seem to us considerably greater than equal distances measured along the ground. Partly for this reason and partly because we see so small a part of the earth's surface at a time, few people realise how shallow the ocean and the atmosphere are, compared with the size of the whole earth. If the earth is represented by a globe a foot in diameter, the ocean on a corresponding scale would nowhere be more than $\frac{6}{1000}$ of an inch in depth. The atmosphere has no well-defined limit (see p. 5), but assuming it to extend to a height of 200 miles, its thickness on the same scale would be about $\frac{3}{10}$ of an inch.

Composition of the Atmosphere. Except for the water vapour present, the composition of the atmosphere near the surface of the earth is practically uniform throughout the globe. Wherever a sample is taken, the air, after it has been freed from water vapour, has very nearly the following constitution :—

Nitrogen	78·03	per cent	by volume.
Oxygen	20·99	,,	,,
Argon	0·94	,,	,,
Carbon dioxide ...	0·03	,,	,,
Hydrogen	0·01	,,	,,

It is only in enclosed spaces, or in the immediate neighbourhood of a volcanic vent or of a factory chimney, or in other places where gases of various kinds are poured out, that any important variation in the composition of dry air is found. The uniformity is due partly to the winds and partly to the rapidity with which gases diffuse and mix with one another.

Up to a height of some 20,000 feet the composition remains practically the same. But at still greater altitudes it is supposed, on theoretical grounds, that there must be differences. In the higher parts of the atmosphere the proportion of the lighter gases will increase and the proportion of the heavier gases will diminish. It has been calculated that carbon dioxide almost disappears from the atmosphere at a height of 12 miles, oxygen at 68 miles and nitrogen at 80 miles. Water vapour, owing to the fact that at low temperatures it condenses to the liquid or the solid form, is confined to the lower part of the atmosphere, and water in any form hardly extends beyond 7 or 8 miles. Above 80 miles the atmosphere consists almost entirely of hydrogen.

These figures, however, must be looked upon as approximations only. The calculations are based on certain assumptions and data which are reasonable but which are not necessarily exact.

Although the composition of dry air at the surface of the earth is almost constant, there are great differences in the amount of water vapour present. The pressure and the temperature also vary greatly from time to time and from place to place. In the study of the atmosphere from the geographical point of view there are accordingly three main factors to be considered, viz., the pressure, the temperature and the humidity.

CHAPTER II

ATMOSPHERIC PRESSURE AND ITS INFLUENCE ON WINDS

Barometric pressure. Since the time of Guericke it has been known that the air has weight. Under ordinary circumstances a cubic foot of air weighs about an ounce and a quarter.

As the air has weight it follows that the atmosphere must press upon the surface of the earth, and the pressure at any point will depend upon the amount of air above. It will be less on the top of a mountain than at the foot. But even at sea-level the pressure varies from day to day.

The pressure of the atmosphere is measured by means of the barometer. The greater the pressure the higher the mercury in the barometer stands. The pressure of the atmosphere is in fact the same as the pressure exerted by a layer of mercury as deep as the mercury column in the barometer is high. When, for example, the barometric reading is 30 inches the atmospheric pressure is equal to that of a layer of mercury 30 inches deep. It is usual therefore to express the pressure of the atmosphere in inches or millimetres of mercury. Since the density of mercury is known it is easy to convert this into pounds per square inch. The average pressure at sea-level is about 29·9 inches, which is equal to about 14·7 pounds per square inch.

Measurement of heights by means of the barometer. When a barometer is taken to the top of a hill it will always be found that the reading is less at the top than it is at the bottom, because the amount of air above is less and therefore the pressure is less. Consequently the barometer may be used for measuring heights. Near the level of the sea a layer of air 9 feet thick is approximately equal in weight to a layer of mercury $\frac{1}{100}$ inch thick. Therefore for every 9 feet that the barometer is raised the barometric reading will be $\frac{1}{100}$ inch less. Thus we get a rough and ready rule for determining heights. Read the barometer both at the top and the bottom of the hill; take the difference in hundredths of an inch and multiply this by nine. The result will be the height in feet of the top above the bottom.

But the rule assumes that the air is always and everywhere of the same density, which is far from being the case. Air is readily compressed, and when compressed into a smaller space its density is naturally increased. At sea-level the air has the whole pressure of the atmosphere above it; at a height of 10,000 feet the pressure is much lower. Therefore at 10,000 feet the density is much less than at sea-level and 9 feet of air is no longer equivalent to $\frac{1}{100}$ inch of mercury, and the rule becomes altogether inaccurate. Moreover, the density of the air is affected also by the temperature. In order to determine heights with greater accuracy the mean pressure and mean temperature of the layer of air between the two stations must be known. The following table[1], which is small enough to be put into the case of a pocket aneroid, will give results correct to within about 1 per cent.

Mean temperature, Fahr.	30°	40°	50°	60°	70°
Mean pressure 27 in... ..	9·7	9·9	10·1	10·3	10·6
,, ,, 28 in... ..	9·3	9·5	9·8	10·0	10·2
,, ,, 29 in... ..	9·0	9·2	9·4	9·6	9·8
,, ,, 30 in... ..	8·7	8·9	9·1	9·3	9·5

Read both barometer and thermometer at the two stations. The mean of the barometer readings may be taken as the mean pressure of the intermediate air and the mean of the thermometer readings as the mean temperature. Look up in the table the number corresponding with this mean pressure and mean temperature, and use this number instead of 9 in the rule already given.

Example:

	Pressure	Temperature
Upper Station	28·00 in.	45°
Lower Station	30·00 in.	55°
Mean	29·00 in.	50°
Difference	2·00 in.	

Factor for mean pressure 29 inches and mean temperature 50° = 9·4.

$200 \times 9{\cdot}4 = 1880$ = Difference of altitude in feet.

Height of the Atmosphere. If the atmosphere were of the same density throughout, it would be easy to calculate its height; for since $\frac{1}{100}$ inch of mercury corresponds with about 9 feet of air 29·9 inches of mercury will correspond with 26,910 feet of air, *i.e.* rather more than 5 miles. But because the pressure decreases upwards the density also decreases and the same weight of air occupies a larger space. At a height of about $3\frac{1}{2}$ miles the pressure is half what it is at sea-level, that is to say half the atmosphere (measured by weight and not by volume) is below and half above. At 7 miles the pressure

[1] From the *Pocket Altitude Tables*, by G. J. Symons.

is about one-quarter of the pressure at sea-level, at $10\frac{1}{2}$ miles one-eighth and so on. If we imagine the atmosphere divided into horizontal layers about $3\frac{1}{2}$ miles thick the lowest layer contains half the atmosphere (by weight), the second a quarter, the third an eighth, the fourth a sixteenth and so on. If such a law as this, which is approximately true near the earth's surface, holds at all altitudes, there can be no definite limit to the atmosphere, and it must extend upwards in an extremely attenuated form until the attraction of the earth is overpowered by that of some other heavenly body. But long before this point is reached the pressure must be far lower than has ever been attained experimentally, and the behaviour of gases at such low pressures is not accurately known.

There is, however, definite evidence that the atmosphere extends beyond 100 miles. Meteorites or shooting stars are small solid bodies which travel through space at great velocities. In space they are cold and invisible to us, but so great is their speed that when they enter the atmosphere the friction of the air and the compression of the air in front make them white-hot and they become visible, and this is sometimes found to be the case at heights greater than 100 miles. They have even been observed at an altitude of 188 miles.

Barometric variations. Although the average pressure at sea-level is about 29·9 inches there are considerable variations from day to day, and in England the barometer seldom remains steady for any length of time. It has long been observed that there is some connection between the fluctuations of the barometer and the changes in the weather ; but the connection is not so simple as is usually supposed. It is often said that when the barometer stands high the weather will be fine, and when it stands low the weather will be wet ; and accordingly many barometers are marked with the words Fine, Change, Rain, etc., but very little reliance can be placed on these indications.

Another common belief, for which there is more justification, is that a rising barometer foretells fine weather and a falling barometer wet. But even this rule has many exceptions.

There is indeed a very close connection between the weather and the barometric pressure. But, to take London as an example, it is not the actual pressure in London that determines the weather there, but the distribution of pressure throughout the surrounding region. If the barometer stands at 29·5 inches the weather may be fine or it may be wet. It will depend to a very large extent upon the pressure in the places round about.

Isobars. The distribution of pressure is determined by taking simultaneous readings of the barometer at a number of different places. These readings are telegraphed to a central office—in England to the Meteorological Office—and

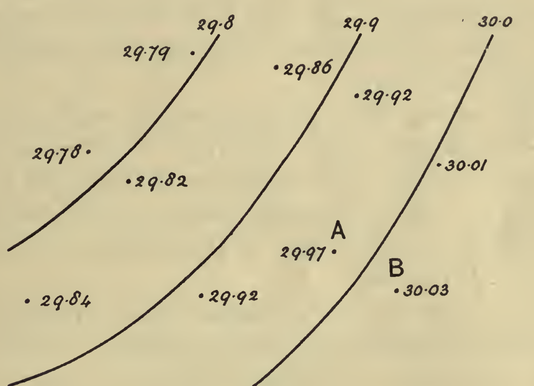


Fig. 1. Method of drawing isobars.

after certain corrections are plotted upon a chart. In order to show the distribution of pressure clearly, isobars or lines of equal barometric pressure are drawn through all places where the pressure is the same.

The method of drawing the isobars is illustrated in Fig. 1. In this figure the readings of the barometer at the different points shown are given. It is usual to draw isobars for each tenth of an inch, and in this figure isobars for 29·8, 29·9, and 30·0 (as shown) would be drawn. None of the places of observation have these readings; but it is assumed that for short distances the change in the pressure is gradual and fairly regular. If then it is 29·97 at *A* and 30·03 at *B* the isobar for 30·0 will pass half way between these points. In this way we obtain the isobars as shown. It should be noticed that all the numbers on the right of the 30·0 isobar are above 30·0, and all the numbers on the left are less than 30·0; and similarly with the other isobars.

But in a single town there may be two barometers, one on the top of a hill and the other at the bottom. The barometer on the hill will always read less than the barometer at the foot, and there is no reason why one reading should be taken as the pressure for that town rather than the other. It is evident therefore that in comparing barometers some allowance must be made for altitude. On the top of Ben Nevis the barometric pressure will always be less than in London and no deductions concerning the weather could be drawn from that fact.

For this reason the pressures plotted on the chart are not the actual readings taken. In every case they are "reduced to sea-level," that is, an allowance is made for the altitude of the place; and the number plotted is the pressure which there would be at that place at the bottom of a pit reaching to the level of the sea.

Isobars on an ordinary weather-chart are therefore lines drawn through places where the barometric readings, with all necessary corrections, including reduction to sea-level, are the same.

Isobaric charts have been prepared daily in England and other countries for many years, and comparison with the weather for each day has gradually shown that particular forms of isobar are usually associated with particular types of weather. It is on this general principle that most weather-forecasting by the various meteorological offices is based.

Relation of wind to isobars. Of all the weather conditions the one which is most evidently and directly dependent upon the distribution of pressure is the wind, and the reason for this is simple. If the pressure in one place is high and in the surrounding area low, then the air will be squeezed outward, as it were, from the higher pressure, and will flow towards the lower. Accordingly we always find that the winds blow from high pressure to low pressure.

If this were all, the wind would blow straight from the high pressure to the low pressure at right angles to the isobars. But the direction is modified by the rotation of the earth. In the northern hemisphere every wind is deflected to the right of the course which it would take if the earth were still; in the southern hemisphere the deflection is to the left. The cause of the deflection will be explained later: its consequence is that the winds are not at right angles to the isobars but are inclined as in Figs. 2 *a* and 2 *b*. We can accordingly formulate the following rule:

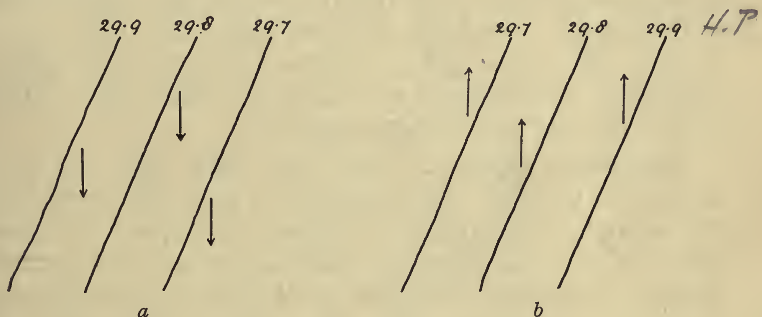


FIG. 2. Relation of winds to isobars in the Northern Hemisphere.

Stand with your back to the wind; if you are in the northern hemisphere there will be high pressure to your right and low pressure to your left; if you are in the southern hemisphere the high pressure will be to your left and the low pressure to your right.

This is known as Buys Ballot's law. It must not be supposed, however, that the actual centre of highest pressure will be exactly to your right or to your left. It is more likely to lie a little backwards also.

When the wind is known, Buys Ballot's law enables us at once to infer the approximate direction of the high pressure. If, for example, the wind in England is from the north, then there is high pressure over the Atlantic towards the west or north-west ; if the wind is from the south, the high pressure is over the continent towards the east or south-east.

When the air is almost still, with only very light and gentle breezes, the direction of these breezes may show no apparent relation to the isobars on the ordinary weather-charts. They are due to local differences of pressure, too small to appear upon the chart. In the valleys of a mountainous district, or in the streets of a town, the direction of the wind is influenced by the valleys or streets and is not determined entirely by the pressure. But when no such local causes interfere, Buys Ballot's law is almost universally true.

The amount of the deflection is not always the same. Sometimes it is so great that the winds are almost parallel to the isobars ; sometimes it is not more than about 45°. No general rule can be given, but the amount of the deflection appears to be influenced by the form of the isobars, that is, by the type of pressure distribution.

Barometric gradient. Since the winds are due to differences of pressure, the strength of a wind will depend upon the amount of the difference. In order to express this numerically the term barometric gradient has been introduced. The barometric gradient is the rate of fall of the barometric pressure measured in a direction at right angles to the isobars. The unit of barometric gradient is a fall of $\frac{1}{100}$ inch in 15 nautical miles. If, for instance, the isobars 29·5 and 29·6 are 30 nautical miles apart (in a direction at right angles to the isobars) the fall is $\frac{10}{100}$ inch in 30 miles, which is equal to $\frac{5}{100}$ inch in 15 miles, and the gradient will accordingly be 5. This is above the average and such a gradient would be called high or "steep."

On any weather-chart it will usually be found that where the isobars are close together and the gradient is therefore high, the winds are strong, and where the isobars are far apart and the gradient low, the winds are light. In general, the higher the gradient the stronger are the winds ; but no exact relation between the two has yet been found and the same gradient does not always produce winds of the same force.

Deflection of the winds : Hadley's explanation. The first reasonable explanation of the effect of the rotation of the earth upon the direction of the winds is due to Hadley.

At the equator everything that appears at rest upon the earth is really moving eastward with the earth at a rate of about 1000 miles an hour. At 60° from the equator, the eastward velocity is only 500 miles an hour.

Looking down upon the North Pole as in Fig. 3 every part of the earth is moving in the direction of the arrow, completing the circle in a day. In three hours, or an eighth of a day, everything "at rest" at *A* will move to *A′* and everything "at rest" at *B* will move to *B′*. If in addition to this movement we give to a particle at *A* an impulse towards the pole *N*,

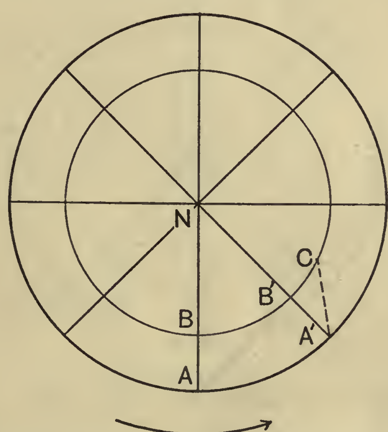


FIG. 3. Hadley's explanation of the deflection of a south wind.

that would bring it to *B* in three hours; it will be carried the distance *AB* towards the pole, but it has also an eastward motion sufficient to take it to *A′* in the same time. Consequently, it moves northwards the distance *AB* and eastwards the distance *AA′*, which is equal to *BC*. Therefore instead of reaching *B* it reaches *C*, and its actual movement is from *A* to *C*.

But during this time the point *A* of the earth's surface has moved to *A′*, and the point *B* to *B′*; and consequently the final positions of the starting point, of the point to which the particle was aimed, and of the particle itself are A', B', and C. The apparent movement of the particle has been from A' to C and instead of moving as it was aimed, directly to the pole, its movement on the earth has been in a north-easterly direction.

If on the other hand we give a particle at B (Fig. 4), an impulse which will carry it due south to A in three hours, it will move the distance BA in that direction and also the distance BB' eastward. It therefore reaches the point C, AC being equal to BB'. But in the three hours B has moved to B' and A to A'; and the final positions of the starting

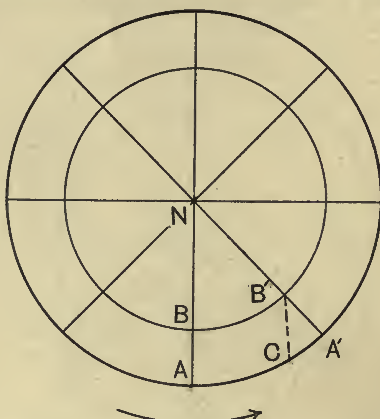


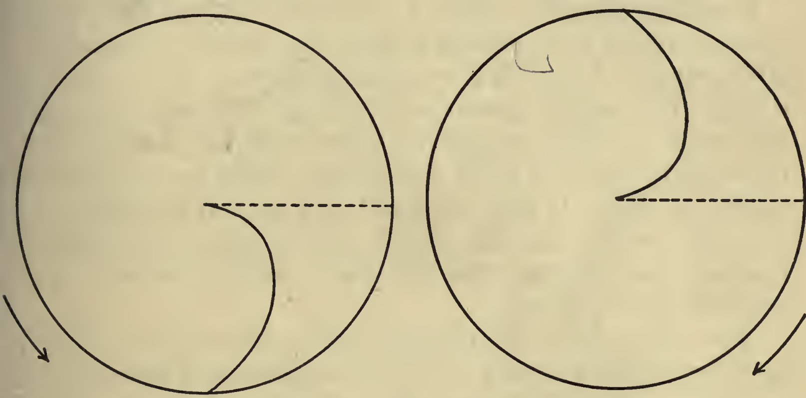
FIG. 4. Hadley's explanation of the deflection of a north wind.

point, the point towards which the particle was aimed, and the particle itself, are B', A', and C. The movement of the particle on the earth has been in a south-westerly direction $B'C$, instead of due south BA.

In both cases the particle has been deflected to the right of its intended course. It can be shown in the same way that in the southern hemisphere there will be a deflection to the left.

Experimental illustration. It is easy to illustrate experimentally the apparent deflection of the course of a particle moving from the pole to the equator. Fix a circular piece of card upon the table by a pin passing through the centre.

This may be taken to represent a view of the globe looking downward on the pole. Turn the card round the axis formed by the pin and at the same time draw a pencil point from the pin in a straight line towards some other fixed point on the table, as indicated by the dotted lines in Fig. 5. The line traced upon the card will be a curve similar to those in the figure. If the card is turned in the direction shown by the arrow in Fig. 5 *a*, it represents a view of the globe looking down on the north pole, and the curve will deviate towards the right. If the card is turned in the opposite direction



a *b*

FIG. 5. Curve traced by a pencil on a rotating card.

(Fig. 5 *b*) it represents a view of the globe looking down on the south pole and the deviation is towards the left.

It should be noted that this method cannot be used to illustrate the deflection of a particle moving from the equator towards the pole. On the earth such a particle, in addition to its poleward movement, has an eastward movement due to the earth's rotation. If, in our experiment, we draw a pencil line from the circumference of the rotating card towards its centre, the pencil does not share in the rotation, and a somewhat complex curve is produced which does not represent the movement of any wind upon the earth.

Incompleteness of Hadley's explanation. Hadley's explanation is not complete. According to the principle which he expounds a wind which was nearly east or nearly west would suffer very little deflection. But observation shows that all winds, whatever their original direction may be, are equally affected, and moreover the deflection is always greater than his explanation would lead us to expect.

The effect of the earth's rotation has been more fully investigated by later mathematicians. It depends upon the laws of centrifugal force, and the more important principles can be illustrated by a very simple experiment.

Centrifugal force. When a weight on the end of a string is swung round and round in a circle, the weight is continually attempting to fly off and accordingly exerts a pull upon the string. This pull is known as centrifugal force and the faster the weight is swung round the stronger is the pull, until, if the velocity becomes great enough, the string breaks and the weight flies off. It may be stated as a general principle that when the weight and the length of the string (or radius of rotation) are constant, the centrifugal force increases as the square of the velocity.

Pass the end of a string through a short piece of tube, the edge of which should be rounded to prevent it from cutting the string. Hold the loose end of the string in the left hand and the tube in the right, and swing the weight round and round. It swings round the end of the tube as a centre.

When the weight has attained a steady rate of movement, keep the right hand still for a moment and pull the string with the left so as to shorten the swinging part. The weight is drawn inwards and, as it approaches the centre of rotation, its velocity increases. It will also be found that it pulls more strongly outwards—so strongly in fact, when the string becomes very short, that it is impossible to hold the hand steady. The centrifugal force is increased.

Repeat the experiment but, instead of pulling the string inwards, allow a little more to go out through the tube so as to lengthen the swinging part. The weight flies outwards and, as it recedes from the centre of rotation, its velocity decreases.

It will be found, moreover, that the pull on the string diminishes, that is, the centrifugal force decreases.

These experiments show that when a body is rotating at a uniform speed round a centre, (1) if it is forced towards the centre its velocity increases and the centrifugal force also increases, (2) if it is forced, or allowed to fly, away from the centre its velocity decreases and the centrifugal force also decreases.

Ferrel's explanation. Fig. 6 is a view of the globe from above the North Pole, the circumference is the equator and the centre *N* is the pole. The axis of rotation passes through

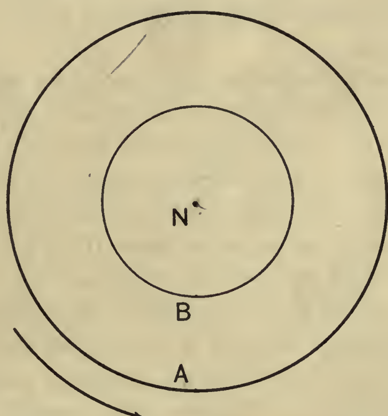


FIG. 6. Ferrel's explanation of the deflection of winds.

N at right angles to the paper; the direction of rotation is shown by the arrow. Anything that moves from the equator towards the pole is approaching the axis of rotation, as the weight did when the string was shortened. Anything that moves from the pole towards the equator is receding from the axis of rotation, as the weight did when the string was allowed to lengthen.

A particle at rest upon the earth at *A* is really moving round the axis at a rate of about 1000 miles an hour. If we attempt to force it directly towards *B* it approaches nearer to the axis of rotation and its eastward velocity becomes more than 1000 miles an hour. But *B* has an eastward velocity less than 1000 miles an hour, and therefore the particle when it reaches the latitude of *B* is in front of *B*. It has been deflected to the right.

On the other hand a particle at rest in latitude 60° N has a velocity round the axis of about 500 miles an hour. If we attempt to force it directly towards *A*, it recedes from the axis of rotation and its eastward velocity becomes less. But the eastward velocity of *A* is more than 500 miles an hour, and therefore when the particle reaches the latitude of *A* it is behind *A*. It has been deflected to the right.

These results are similar to those of Hadley, but calculation shows that the amount of the deflection is very much greater than he supposed.

Next we have to consider the case of an attempted movement in a due east or due west direction. A particle at rest at *B* in latitude 60° is moving round the axis in an eastward direction at a rate of 500 miles an hour. Give it an impulse towards the east. Its eastward movement is now increased and its rate of rotation becomes more than 500 miles an hour. Its centrifugal force is therefore also increased and it moves away from the axis of rotation towards the equator. Therefore instead of travelling due east it moves to the south of east. It is deflected to the right of its intended course.

Instead of attempting to move the particle eastward give it an impulse towards the west. This is in opposition to its movement when "at rest" and its rate of rotation becomes less than 500 miles an hour. Because its rate of rotation is diminished its centrifugal force decreases and accordingly it moves towards the axis, *i.e.* towards the pole. Instead of travelling due west it moves in a north-westerly direction. Again the deflection is to the right.

In all cases therefore in the northern hemisphere a moving particle is deflected to the right of the course which it would take if the earth were still. It can be shown in a similar way that in the southern hemisphere the deflection is to the left.

The following is a fuller statement of Ferrel's explanation of the effect of the earth's rotation on due east or due west movements.

If the rotating earth were a true sphere with a smooth and frictionless surface no loose particle could remain at rest excepting at the poles or the equator. Fig. 7 is a section of the earth through the poles, *NS* being the axis of rotation. Every particle, such as *A*, upon the surface, moving with the earth, is acted upon by two forces, the force of gravity in the direction *AB* towards the centre, and the centrifugal force in the direction *AC* at right angles to the axis of rotation. The latter is a very small force compared with the former. If the earth is a true sphere, *AB* will be at right angles to the surface and can have no tendency to make the particle move in any direction. But the force *AC* is oblique to the surface. It is not enough to lift the particle but it is clear that if there is no friction it will cause the particle to slide towards the equator.

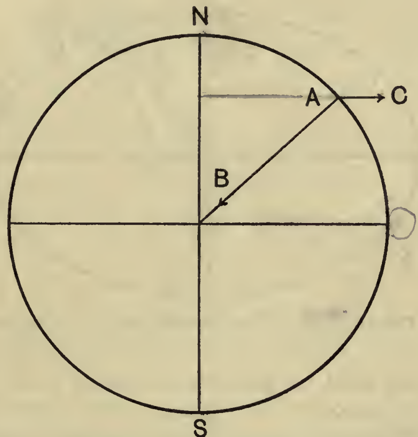


FIG. 7. Effect of centrifugal force upon a spherical earth.

At the pole there is no centrifugal force. At the equator the centrifugal force is at right angles to the surface and is directly opposed to the force of gravity. It therefore has no tendency to make the particle move on the surface and, because it is so small compared with gravity, its only effect is to diminish slightly the weight of the particle.

But the earth is not a sphere. A section through the poles is not a circle as in Fig. 7 but an ellipse, as shown, with the ellipticity very greatly exaggerated, in Fig. 8. In an ellipsoidal earth the force of gravity is not, in general, directed exactly towards the centre, but it is nearly so, and is not at right angles to the surface, except at the poles and the equator. Elsewhere it may be resolved into two components, one *AD* at right angles to the surface the other *AE* tangential. The former presses the particle against the surface, the latter tends to move it towards the pole.

The centrifugal force is at right angles to the axis, in the direction *AC*. It may be resolved into two components, one *AF* at right angles to the surface, the other *AG* tangential. The former makes the particle a little lighter than it would otherwise be, the latter tends to move it towards the equator.

When the particle is at rest on the rotating earth the forces *AE* and *AG* must balance each other. If we now give it an eastward motion relatively to the earth we are increasing the velocity of rotation which it already had. The centrifugal force *AC* increases, *AG* therefore increases and becomes greater than *AE*. The particle therefore slides towards the equator and instead of moving due east it moves to a point somewhat south of east.

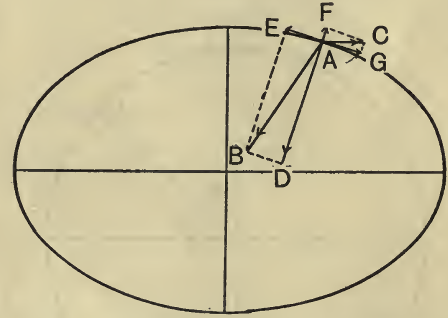


FIG. 8. Effect of centrifugal force upon an ellipsoidal earth.

If on the other hand we give the particle a motion westward, this is opposed to its direction of rotation. Its velocity round the axis becomes less. The centrifugal force decreases and the force *AG* therefore decreases and becomes less than *AE*. The particle therefore slides towards the pole and instead of moving due west it moves to a point somewhat north of west.

In both cases the deflection is to the right. In the southern hemisphere the deflection will be to the left.

The law of deflection of moving bodies on the surface of the rotating earth is often spoken of as Ferrel's law, because Ferrel was one of the earliest writers to develop it fully, particularly in its application to the winds ; but he was not actually the first to enunciate the principle.

CHAPTER III

INFLUENCE OF ATMOSPHERIC PRESSURE UPON THE WEATHER

Isobar shapes. The connection of other weather conditions with the distribution of pressure is not so simple as in the case of the wind. On any chart showing the isobars at a given time we can, with a fair degree of certainty, indicate the direction of the wind at any spot. All that we have to bear in mind is that the wind will blow from high pressure to low pressure but (in the northern hemisphere) will be deflected to the right. From the barometric gradient we can even form a very good idea of the strength of the wind, although we cannot actually calculate it.

But with regard to rainfall and temperature no such certainty is yet attained. They are undoubtedly in some way affected by the distribution of pressure. When for example the isobars are more or less circular in shape, with the highest pressure inside, the weather is usually fine. But it is not always so. In general it may be said that certain shapes of isobars are usually associated with certain types of weather, but exceptions are far more numerous than in the case of the winds.

In the following pages the more important isobar shapes with their associated weather are described, but it should always be remembered that on occasions the weather may be different from that which is characteristic of the shape. Some of the apparent abnormalities may be due to the fact that our observing stations are not close enough to enable us to draw the isobars with perfect accuracy; but in other cases this explanation does not appear to be sufficient.

Cyclones

General description. A cyclone is roughly circular or oval in form, with the pressure lowest near the middle and gradually increasing outwards in all directions. The isobars are therefore circular or oval also, with the isobars of lowest value inside (Fig. 9).

The size of cyclones varies greatly. In our latitudes they may be 200 or 300 miles across or they may be as much as 2000. In the tropics the diameter does not often exceed a few hundred miles.

The height is relatively small. In a later chapter it will be shown that, in England, the disturbances in the lower atmosphere do not extend to a greater altitude than seven or eight miles; and long before this limit is reached, the cyclone has

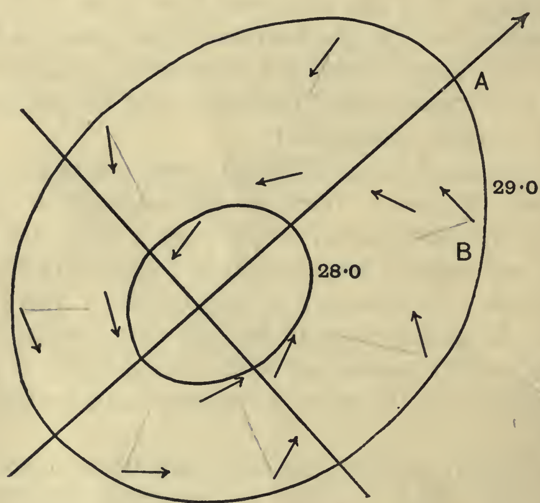


FIG. 9. Isobars and winds in a cyclone.

lost the characters which distinguish it at the earth's surface. A cyclone is therefore a disc and the thickness of the disc is small compared with its diameter. In proportion to its size an ordinary cyclone in north-western Europe is thinner than a penny.

A cyclone is seldom stationary. It usually moves approximately in the direction of the prevailing winds. Thus in north-western Europe most cyclones travel in an easterly direction, and it is very rarely indeed that they move towards the west. In the region of the trade-winds on the other hand the movement is usually westward. When, as sometimes happens, a tropical cyclone passes into the region of the westerly winds, it changes its direction and begins to travel towards the east.

The rate of movement is very variable. Fifteen miles an hour is rather slow for a cyclone in the neighbourhood of Great Britain; forty miles an hour is decidedly fast. Tropical cyclones generally travel much more slowly than this, their rate varying from about two miles an hour to about ten.

The rate of movement of the cyclone must be carefully distinguished from the velocity of the wind in the cyclone. The winds, as we shall see, blow round the centre of the cyclone and their velocity has apparently no more to do with the movement of the centre than the velocity of rotation of the earth around its own axis has to do with its revolution round the sun. In the slow-moving tropical cyclones the winds are usually much more violent than in the more rapid moving cyclones of temperate latitudes.

Since a cyclone moves, it has a front and a back. In the diagram, Fig. 9, the large arrow shows the direction of movement. A line drawn through the lowest pressure and at right angles to the direction of movement is called the trough. The lowest pressure is usually not in the middle of the cyclone, but behind that point. It is nevertheless commonly called the centre of the cyclone; the part of the cyclone in front of the trough line is called the front, and the part behind is called the back.

The following account applies more particularly to cyclones in north-western Europe. Tropical cyclones differ in several respects besides those already noticed and the more important differences will be pointed out later.

Movement of the barometer. In the position in which the cyclone is shown in Fig. 9, the barometer at *A* stands at 29·0 inches. But as the cyclone moves in the direction of the arrow the lower isobars pass over *A* in turn and the barometer falls, until the centre arrives. The isobars of the back of the cyclone then approach, one after the other, and the barometer rises. At a point such as *B*, which is not in the path of the centre, the barometer will fall until the trough line is reached and will then rise again; but the fall will be less than at *A*.

Winds. Since the lowest pressure in a cyclone is near the middle the winds necessarily blow inwards from all sides, but because they are deflected to the right (in the northern hemisphere) they do not blow directly towards the centre. Their directions at various points are indicated on the diagram, from which it will appear that they blow round the centre in much the same fashion as the water of an eddy flows round the depression in the middle of the eddy. In the front the

FIG. 10. Cloud in a cyclone.

wind is mainly from a southerly direction, in the back from a northerly direction.

On account of the fact that the lowest pressure is usually behind the geometrical middle of the cyclone, the isobars are closer together at the back than in the front, and the winds are accordingly stronger. Usually, moreover, the winds in the back of a cyclone make a smaller angle with the isobars than those in the front.

Rain and cloud. (Fig. 10.) The approach of a cyclone is often heralded by the appearance in the sky of long streaks of thin white cloud, in shape not unlike the feathered part of a quill pen, with one margin hard and the other soft. The direction of these clouds gives a good deal of information as to the approaching cyclone. They are really approximately parallel to one another, but owing to the effects of perspective they appear to radiate from some point on the horizon; and if this point lies north-westward the centre of the cyclone is likely to pass near to the observer.

It often happens that these streaks disappear altogether and the sky may again become perfectly clear, with no indication of approaching rain. But after some hours a thin white veil of cloud spreads upwards from the western horizon, giving the sky a somewhat milky appearance. If the moon is behind the veil it will be surrounded by a halo, or ring of light, not in contact with the moon but at an angular distance of about 22°. The halo usually seems white but may be coloured. If the sun is behind the veil a similar halo will be formed but, owing to the dazzling light of the sun, it is seldom seen. If, however, instead of looking at the sky direct we look at the reflection in a piece of blackened glass the halo can be seen as readily as in the case of the moon. A halo around the sun is very often coloured.

As the cyclone continues to approach, the veil of cloud spreads higher and at the same time thickens, until the sky becomes entirely overcast, and a light drizzling rain begins to fall. The rain increases, the wind grows strong and gusty; and in front of the centre there is heavy driving rain, and the wind is often of great violence.

The character of the weather after the passage of the trough depends upon the position of the observer with regard to the path of the centre.

If he is situated south of the large arrow in Fig. 10 he will enter the right-hand back quadrant of the cyclone. As the trough passes there are violent squalls of wind and heavy showers of rain, with short lulls between the squalls. During one of these lulls there will be a sudden shift in the wind, that is to say, in one squall the wind will blow from a certain point of the compass, in the next it will blow from a different direction. The shift may be only a few degrees or it may be much more considerable. Its suddenness is often dangerous at sea.

After the shifting of the wind, the clouds begin to break. At first only a few blue patches appear, but they rapidly extend and the sky clears. For some time, however, heavy masses of cumulus cloud, often in long lines, continue to blow over from some westerly point. As each of these masses passes overhead there is a squall with heavy rain or hail. But gradually the fine intervals increase in length. The squalls become less frequent until at last they cease altogether and the sky is a clear and brilliant blue throughout and the wind dies down.

If the observer is situated north of the path of the centre it is the left-hand back quadrant that passes over him ; and in this case there will not be the same sudden change in the character of the weather. The rain and cloud extend for some considerable distance behind the trough. Gradually the rain ceases and the clouds disperse, but there is no rapid clearing of the sky as in the southern half of the cyclone.

Temperature. The changes of temperature in a cyclone vary somewhat according to the time of year and even the time of day. In the winter the front is warm for the season, the back is cold and frosty. In the summer the kind of change that will be felt depends very much upon the weather of the preceding days. If for instance the cyclone follows a spell of hot and sunny weather the mere covering of the sky with clouds will cause the temperature to fall. But in spite of this the heat remains oppressive. The front, in fact, may at all seasons be described as "close" or "muggy." The back, with its clear sky, will naturally have a bright sun, which in the summer will be hot, but the air itself will even then be refreshingly cool, and the back is best described as "fresh."

Tropical cyclones. Tropical cyclones differ from those of temperate latitudes in several respects. Their area is usually less and their rate of movement slower. The gradient is generally steeper and the winds are therefore more violent and the rainfall heavier.

These are differences merely of degree, but there are other differences also. A tropical cyclone is more symmetrical, both in shape and in the distribution of rainfall and temperature. The isobars are nearly circular, and the centre of low pressure is nearly in the middle. There is no marked difference between the front and the back except in the direction of the wind. From whatever side the centre is approached the clouds and wind and rain increase from the margin inwards. But the increase does not continue to the centre. The centre itself is an area of calm, known as "the eye of the storm," where the winds and rain cease and even the clouds may partially disperse. This calm area may be from ten to thirty miles

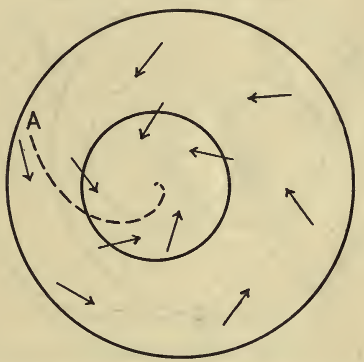


FIG. 11. Motion of the air in a stationary cyclone.

across. Although it is calm in the sense that the air is still, yet at sea it is no haven of rest. It is the meeting place of the waves produced by the winds around, and it is a chaos of disturbed and tumultuous waters.

Seen from above a tropical cyclone would appear as a broad ring of dense cloud with a small clear space in the middle. In a cyclone of temperate regions the cloud would be roughly semicircular with a ragged diametral edge.

Causes of the characteristic weather of cyclones. In Fig. 11 it will be seen that the winds everywhere blow inwards, not, indeed, directly to the centre but nevertheless always inwards. If the cyclone did not move, any particle of air such as *A* would therefore approach the centre in a spiral course as shown by the broken line. The air from all round would do the same and the motion would be that of a stationary eddy.

The cyclone, however, is not still. In our latitudes it moves in an easterly direction. By the time the particle *A* (Fig. 12) has reached the point *B*, the centre of the cyclone is no longer at *O* but at *O′*. It may easily happen, as in the figure, that the centre has moved forward faster than the particle has approached it; and in that case although the particle *A* is moving towards the centre all the time, it is actually getting further behind it.

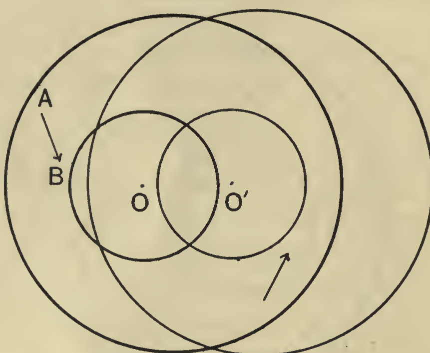


FIG. 12. Motion of the air in a moving cyclone.

Without going more fully into the matter it is easily seen that the particles which have the best chance of reaching the centre are those which are aimed, as it were, towards a point in front of the centre, that is to say, the particles carried by the southerly winds in the south-eastern quadrant of the cyclone. By tracing the course of the winds from hour to hour it has been shown that this is the case. The air that reaches the centre approaches it from the southern side of its path. Sometimes, in a slow-moving cyclone, it may originally have come from the northern side, round by the back to the south. The motion is not that of a stationary eddy, but it is not unlike that in the moving eddies of a rapid stream; but in such eddies the whirling motion is usually more rapid in proportion to the forward movement than it is in an ordinary cyclone.

Since more of the surrounding air reaches the middle than blows out, the pressure there would necessarily increase and the cyclone would gradually die away, unless this air found some escape. But cyclones often exist for days without any diminution in intensity. It is evident therefore that the air which reaches the centre does not remain but somehow finds its way out again, and its only possible outlet is upwards. It is not sufficient, however, that it should simply go vertically upwards, for in that case its pressure would not be

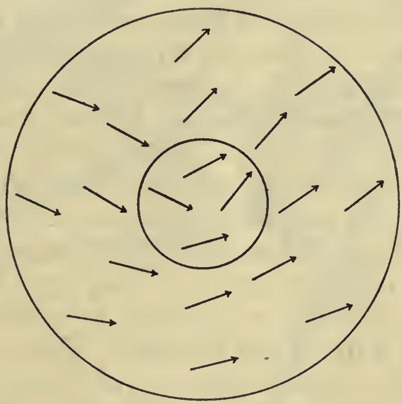


FIG. 13. Upper currents in a cyclone at a height of about 10,000 feet. (After Bigelow.)

removed. Sooner or later it must escape laterally, or the low pressure will not be maintained. Observations of the upper clouds in a cyclone have shown that at a high level the air actually does move outwards, the outward movement being chiefly in the forward direction The upper currents, as deduced from the movement of the high clouds, are shown in Fig. 13.

Towards the middle of a cyclone the air is therefore rising. As it rises, the pressure upon it decreases. It expands and its temperature falls, and the water vapour in it condenses as cloud or rain.

In a climate such as ours the winds are variable, but nevertheless, throughout the North Temperate Zone, the prevailing direction is from the west or south-west. The variability is due chiefly to the passage of cyclones and other atmospheric disturbances. There is in fact a constant flow of air in an easterly or north-easterly direction, and a cyclone is an eddy in the flow and moving like an eddy in a stream of water. Therefore cyclones in these latitudes usually move towards the east.

Owing to the absence of friction with the ground the upper air moves more rapidly than the air at the surface. The rising air of the cyclone is accordingly carried chiefly forward and tends to overspread the front rather than the back. This is probably one of the reasons, but not the only one, why the rain of a cyclone falls mostly in the front half.

The difference in temperature between the front and the back of a cyclone appears to be due chiefly to the direction of the surface winds. In the front the winds come from a southerly quarter and are therefore warm. They are moving to a colder latitude and are accordingly less able to hold the water vapour that they contain. Thus they tend to become moist and the sensation of closeness to which they give rise is due to their warmth and moisture. This is the case more especially in the southern part of the front. In the north-eastern quadrant, where the winds become more easterly, they may in winter be cold.

In the back of the cyclone the winds are generally northerly and are therefore cool. They are moving to a warmer region and are accordingly becoming drier. It is their coolness and dryness that make the back of the cyclone "fresh."

Cyclone paths in north-western Europe. Although a cyclone is a disturbance in a more or less permanent stream of air and its movement is determined chiefly by the flow of the stream, its path is modified to some extent by the nature of the surface over which it moves. A high mountain range forms an obstacle which it can hardly surmount and even low hills appear to interfere with its progress. Consequently the cyclones which approach our shores from the Atlantic tend to follow certain routes in preference to others. The paths indeed are very vaguely defined and many cyclones stray far from any regular track. But, on the whole, cyclones seem to travel more easily over the sea, and if they cross the land they usually choose the lower-lying ground.

The principal paths are shown in Fig. 14. The greater number of cyclones in our area skirt our western shores and

FIG. 14. Cyclone paths in Western Europe. (After Van Bebber.)
The width of the paths is approximately in proportion to the number of cyclones which follow them.

travel north-eastward to the Norwegian coast. Some pass from the Atlantic into the North Sea, between Scotland and Scandinavia. Some, which have a more southerly course, traverse the low lying ground from the Bristol Channel to the Wash, or from the Bristol Channel to the mouth of the

Thames; or they may come up the English Channel or bend southwards into France.

Occasionally a cyclone appears to become embayed, and it may remain in almost the same position for two or three days. This happens, now and then, in the English Channel, causing a succession of very rainy days in south-eastern England.

The path is influenced not only by the surface over which the cyclone moves, but also, and perhaps even to a greater extent, by the neighbouring high pressure areas. It is probably on account of the variability thus introduced, that the routes are so inconstant as they are.

ANTICYCLONES

General description. An anticyclone is an area of high pressure with the pressure gradually decreasing outwards.

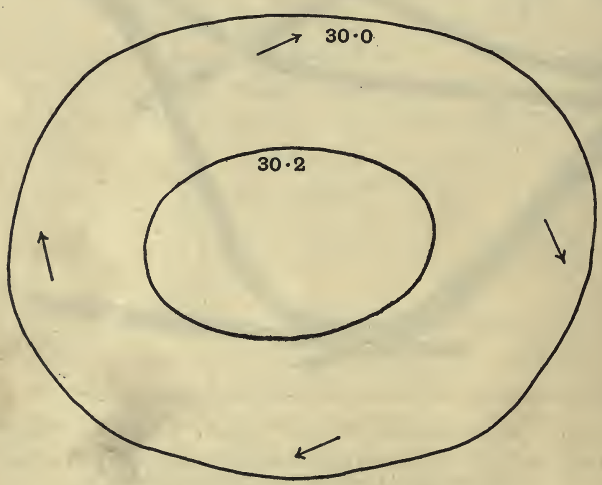


FIG. 15. Isobars and winds in an anticyclone.

The isobars are therefore more or less oval or circular, the highest isobars being inside. So far as the distribution of pressure is concerned an anticyclone is the opposite of a cyclone, and there is a corresponding contrast in the character of the weather associated with it (Fig. 15).

An anticyclone as a rule does not travel in any well-defined path. It may wander undecidedly in almost any direction; it may spread outwards on one side or another and may afterwards retreat; it may remain stationary for days with very little change. It disappears sometimes by drifting slowly away, sometimes by gradually diminishing in intensity. But whatever changes take place, they are usually slow and never violent.

Winds. The isobars in an anticyclone are far apart, especially towards the middle, and the winds are therefore light. Near the centre there is a calm, or very light and variable airs. It is this stillness of the air that forms the primary characteristic of an anticyclone. It is in fact an almost motionless mass of air taking little part in the movements around it. It is analogous in some respects to the patches of smooth water which drift down-stream amongst the swirls and eddies of a river in flood.

On account of the general stillness of the air local influences make themselves felt. Differences of temperature due to unequal heating of land and water or of mountain and plain produce local winds. Land and sea breezes, mountain and valley winds often become noticeable; in a cyclone, on the other hand, they are completely overpowered by the strong winds due to the cyclone itself.

On the margins the winds are steadier and more definite in direction but are still usually light. When a cyclone approaches they increase in force and may become strong, but this is the effect of the cyclone rather than of the anticyclone.

Apart from the modifications due to local influences the general direction of the winds is naturally outwards, away from the high pressure. In the northern hemisphere, they are deflected to the right by the earth's rotation, and consequently on the eastern side of the anticyclone the winds are northerly, on the western side southerly, on the northern side westerly and on the southern side easterly.

Weather. It has already been pointed out that on account of the general stillness of the air in an anticyclone local winds may be produced by local causes, and for similar reasons there may be local rains. Indeed, according to Shaw and Lempfert, "Local changes of many kinds may take place within them, and almost any kind of weather, except those which represent violent atmospheric changes, may be associated with their central regions."

Nevertheless the weather in an anticyclone is usually fair or fine. In England the interior of an anticyclone may have an overcast sky and there may even be rain, though the amount is usually small. Both cloud and rain are probably connected in some way with the insular position of our island, and possibly with the small size of the anticyclones that visit us, for in the larger anticyclones of the globe the interior is usually almost free from cloud. Even in England this is often, if not generally, the case.

In the middle of such an anticyclone, without cloud and without wind, the weather experienced will depend largely upon the season. Although there is but little actual cloud the atmosphere is not clear. The sky above is blue, but the horizon has the characteristically hazy appearance so often seen in settled fine weather. The absence of cloud permits the sun to shine brightly during the day, but at night there is free radiation and consequent loss of heat. In summer therefore the day is hot, but when the sun has set the temperature begins to fall. Even in the middle of summer it usually falls sufficiently to allow of the formation of mist and dew in the early morning, but this disappears soon after sunrise.

In the late summer or early autumn the mists and dews are more pronounced and may last for several hours after the sun has risen.

In winter the sun is never high and it is above the horizon for so short a time that it produces little warmth. The nights are longer, the loss of heat is greater, and often there is frost and fog. The fog may be so thick as to last throughout the following day.

In spite of the exceptions already noticed, due probably to local causes, the weather just described is that which is characteristic of the middle of an anticyclone. Towards the margins it varies. Although the winds are slight there is a general drift of the air, and the temperature depends largely on the source from which it comes. Thus on the eastern margin the winds are northerly and the temperature is therefore low for the time of year. In winter it is likely to be frosty with perhaps some snow.

On the southern margin the winds are easterly. In England they come from the continent, which is hot in the summer and cold in the winter. Therefore with us the temperature will be high in the summer ; in winter it will be cold and often frosty. In the winter the sky is frequently overcast and black.

The winds of the western margin are southerly, and this side of the anticyclone is usually warm.

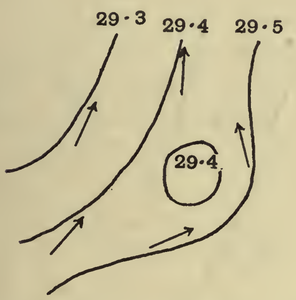


FIG. 16. Secondary cyclone with a centre of low pressure.

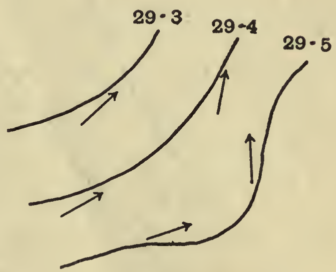


FIG. 17. Secondary cyclone without any definite low-pressure centre.

On the northern margin the winds are westerly, coming to England from the Atlantic, which is relatively cool in summer and warm in winter. The temperature will be comparatively cool in summer, in winter it will be warm.

OTHER ISOBAR SHAPES

Secondary cyclone. A secondary cyclone appears on a chart as a mere bend in the isobars, usually, but not always, on the margin of a cyclone. Sometimes there is a definite centre of low pressure in the middle (Fig. 16), but more commonly the pressure decreases gradually towards the primary cyclone (Fig. 17). The bend may cover a considerable area or it may be so small that the observing stations are not close enough to reveal its presence. But, unfortunately for the forecaster, the effect upon the weather is out of all proportion to the magnitude of the bend.

? The winds obey the usual law. In the interior the isobars are far apart and the winds are light. On the convex side of the bend the gradient is steeper and the winds are strong and gusty.

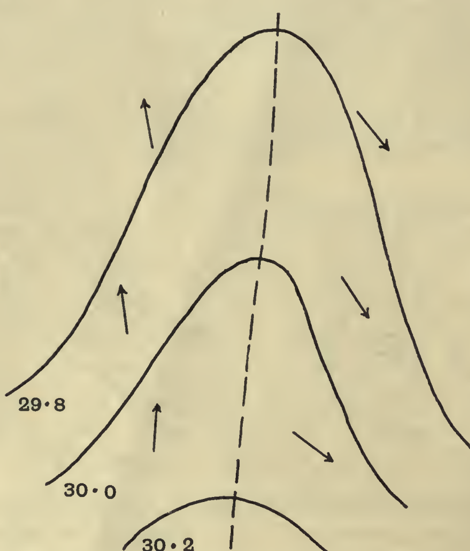


FIG. 18. Isobars and winds in a wedge.

A secondary cyclone usually travels in an easterly direction. It is ushered in by gusts of wind and heavy squally showers of rain, and in the middle is a steady downpour of rain with little or no wind. Thunderstorms are a frequent accompaniment of a secondary cyclone, especially in summer.

On account of the small size of secondary cyclones, the rain is often limited to a very small area and within a few miles the weather may be fine throughout the day.

Wedge. A wedge is a triangular area of high pressure, with the pressure highest at the middle of the base and decreasing towards the point and sides. It is the region of high pressure between two cyclones (Fig. 18).

Most cyclones in north-western Europe pass to the north of England and the point of the wedge is accordingly directed northwards. But if the path of the cyclones lies to the south the wedge will point southwards.

The wedge necessarily moves with the cyclones between which it lies. The line of highest pressures from the point to the middle of the base is called the crest, and the direction of movement is approximately at right angles to the crest.

The winds (in the northern hemisphere) will be as shown in the diagram. In the usual case, with the wedge pointing northwards, they will be northerly in the front half, southerly in the back half. The isobars are generally rather far apart and the winds are therefore light.

The front half of a wedge is characterised by a remarkable clearness of the atmosphere. The sky is cloudless and brilliantly blue. There is no haze and the horizon is sharp and clearly defined. The air is fresh and the barometer rising.

Shortly before the approach of the crest, however, the western sky grows milky, and through the thin white clouds which produce this appearance the sun and moon may form halos. After the crest has passed, the barometer begins to fall, the sky becomes overcast and the drizzling rain of the following cyclone begins.

V-depression. A V or V-depression is the opposite of a wedge. It is a triangular area of low pressure, with the pressure lowest at the middle of the base and increasing towards the point and sides. The line of lowest pressures from the point to the middle of the base is called the trough, and the movement is usually at right angles to the trough in an easterly direction. The V, however, may point in almost any direction (Fig. 19).

In the case of a V-depression with its point directed southwards, the winds in the front half are southerly and in the back half northerly. The isobars are close together and the winds are strong. The trough is a line of violent squalls, often with thunderstorms, and as it passes there is a sudden shift in the direction of the wind, which is often very dangerous to shipping.

The distribution of cloud and rain in a V-depression varies; but usually the weather is not unlike that in a cyclone, with most of the characteristics intensified. There is a milky sky in front, forming halos round the sun and moon. The front half is wet, with heavy and squally rain. At the trough the squalls attain their greatest violence and the wind shifts. Then the clouds begin to break, and after a series of heavy clearing showers they finally disappear, leaving a bright sky and a cold fresh wind as at the back of a cyclone.

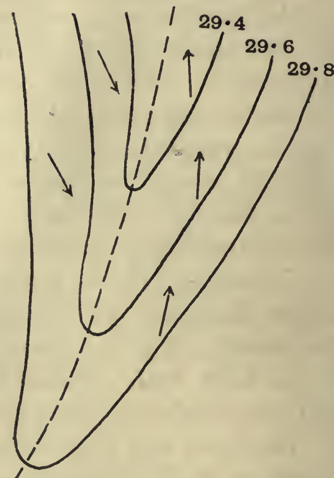


FIG. 19. Isobars and winds in a V-depression.

CHAPTER IV

DISTRIBUTION OF PRESSURE AND CIRCULATION OF THE ATMOSPHERE

General distribution of pressure. As the weather for the day depends upon the distribution of pressure for the day, so the climates of the globe depend upon the general distribution of pressure over the globe. In some parts of the world the pressure is permanently high and rain is seldom seen. In other parts the pressure is always low and rain is nearly constant. In others the pressure changes from day to day and the weather varies with the pressure. The distribution of pressure over the globe, taking the average pressures for the whole year, is shown in Map I.

There is a permanent belt of low pressure about the equator, passing completely round the globe. On each side of this is a belt of high pressure with its central line about 30° or 35° away from the equator. The temperate regions are both areas of low pressure, but towards the poles the pressure again increases. The high-pressure belts of the tropics show a tendency to spread over the great continental masses of the northern hemisphere, but apart from this the general distribution may be represented diagrammatically as in Fig. 20.

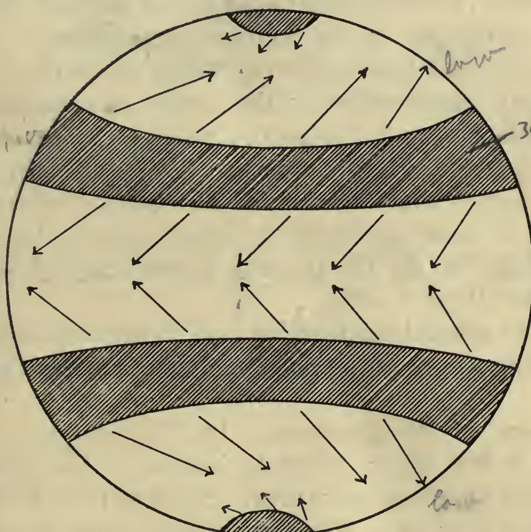


FIG. 20. Diagram of the general distribution of pressure and winds of the globe.
The high-pressure areas are shaded.

From this distribution of pressure it is easy to deduce the direction of the prevalent winds, remembering always that they blow from high pressure to low pressure but are deflected to the right in the northern hemisphere, to the left in the southern hemisphere.

From the tropical high pressure belts the air flows to the equatorial low pressure, giving north-easterly winds in the northern hemisphere, south-easterly winds in the southern hemisphere. These are the "trade-winds" and they are remarkably steady and constant in direction.

Winds also blow from the tropical high pressure to the low pressures of the temperate regions. In the northern hemisphere they are south-westerly, in the southern hemisphere north-westerly. These are the "westerlies." They are not so steady as the trade-winds, especially in the northern hemisphere, for in these regions cyclonic and other disturbances are common.

Finally, from the high pressures of the poles the winds blow outwards, from the north-east in the northern hemisphere from the south-east in the southern hemisphere. The polar high pressure and the outflowing winds are better defined in the Antarctic regions than in the Arctic, but even there they do not extend very far from the pole.

A comparison of the diagram (Fig. 20) with the map (Map 1) shows that the actual distribution of pressure is influenced to some extent by the distribution of land and sea. It is also affected by the position of the sun and varies with the seasons (Maps 2 and 3).

Both the general distribution and the modifications are the effect of differences of temperature combined with the earth's rotation.

Influence of temperature. In the absence of other influences a hot region tends to become a low-pressure and a cold region a high-pressure area. Suppose that *AB* (Fig. 21 *a*) represents a level plain where the temperature is uniform and the air is still. In order that the air may be still the pressure at all points on the same level must be the same, for otherwise a wind will blow from the higher to the lower pressure.

In the diagram the pressure on the ground is 30 inches. At a certain height above the ground it is 29 inches. At a higher level 28 inches and so on. These surfaces of equal pressure are called isobaric surfaces, and as the air is still they must be level.

Suppose that a part *CD* of the plain (Fig. 21 *b*) is heated and becomes hotter than the rest. The air in contact with it will be warmed and will expand, lifting the air above it. Consequently the isobaric surfaces above *CD* will be raised as in Fig. 21 *b*. The air however cannot remain in this condition.

The pressure at *E* is now greater than at *F* and *G* on the same level, and accordingly the air will flow outwards from *E* to *F* and *G*.

When this happens some of the air above *CD* is removed from there and is added to that above *A* and *B*. Consequently the pressure *on the ground* is no longer uniform. It becomes less than 30 inches on *CD* and more than 30 inches at *A* and *B*. The isobaric surface of 30 inches ceases to coincide with the ground but runs as in Fig. 21 *c*. Near the ground the isobaric surfaces will now be concave upwards while higher in the air they are concave downwards.

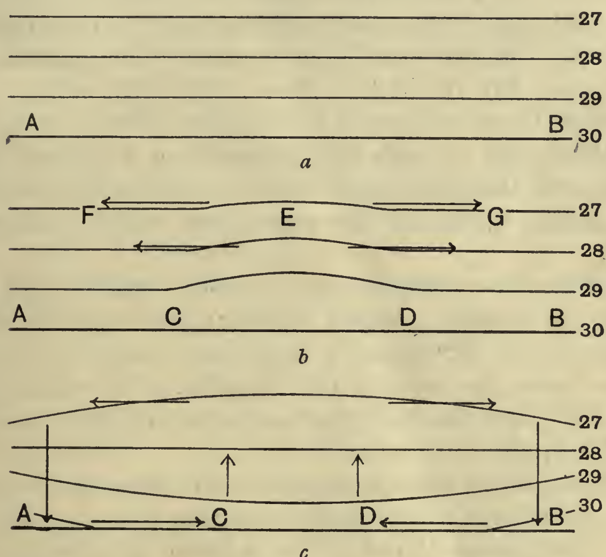


FIG. 21. Effect of heat upon pressure and winds.

Because the pressure on the ground at *A* and *B* is now greater than at *CD*, there will be a surface wind from *A* and *B* inwards to *CD*.

The effect of heating *CD* is therefore to cause an outflowing current above and an inflowing current below; and on the ground-level to decrease the pressure on the heated part and increase the pressure on the cool part.

Moreover, the air over *A* and *B* is pressed downwards by the addition of air above and therefore sinks. The warmer and lighter air over *CD* is lifted upwards by the influx of the colder and denser air below and therefore rises. The upper currents will be cooled as they flow away from the source of heat, the lower currents will be warmed as they flow towards it ; and consequently as long as the difference of temperature on the ground is maintained the circulation will be kept up.

It must not be supposed, however, that this circulation extends to the top of the atmosphere. When first *CD* is heated, only the layer of air in contact with it is warmed and expanded. But because the air is elastic this expansion does not at once lift the layers above. Its only effect for the moment is to compress the next layer. Thus it takes time to establish the circulation described. In a general way it may be said that the longer the difference of temperature is kept up, the higher in the atmosphere will the circulation extend.

According to the principle explained above the lowest pressure, at sea-level, should be about the equator and the highest pressure at the poles and there should be a gradual increase from the equator to the poles. The surface winds should blow from the poles to the equator and the upper winds from the equator to the poles.

Influence of the earth's rotation. But owing to the rotation of the earth these winds cannot maintain their original directions. They are deflected to the right in the northern hemisphere, to the left in the southern hemisphere.

If a large body of men is moving to a central point from all around, and those in front reach the centre about the same time, there will be a block. If those behind still move forward a very high pressure will soon be produced at the centre. This illustration represents fairly well the conditions of the upper currents on a non-rotating earth. They will produce a high pressure at the poles, and the effect of this high pressure is to cause under currents from the pole to the equator.

But suppose that before they reach the centre the front ranks turn to the right (not necessarily at right angles), the hinder ranks will catch them up, and the block and high pressure will not be produced at the centre but in a ring around the centre and at some distance from it. The same thing will happen if all ranks turn to the right but those in front turn more than those behind.

The conditions now are somewhat similar to those of the upper currents on a rotating earth, flowing from the equator to the north pole. As they move northwards they are deflected to the right and instead of adding to the pressure at the pole, they produce a ring of high pressure between the equator and the pole. On the surface this high pressure will cause under currents flowing both to the equator and to the pole.

This illustration will perhaps serve to show why, on a rotating earth, the principal high pressure is not at the pole. But the whole problem is one of great complexity and the mathematical theory of the general circulation of the atmosphere is still very imperfect.

Seasonal changes. If the sun were always vertical at the equator the position of the high- and low-pressure belts would remain unchanged throughout the year. But in our summer the sun is above the Tropic of Cancer, in our winter above the Tropic of Capricorn. The pressure belts move with the apparent movement of the sun, but not to the same extent. In the northern summer the equatorial low-pressure belt lies a few degrees north of the equator, in the southern summer a few degrees south.

But there is a greater change than this. In the northern winter the northern land-masses are far colder than the neighbouring seas and they therefore become areas of high pressure. The tropical high-pressure belt accordingly spreads northwards over the continents of America and Asia.

At the same time it is the southern summer, and in the southern hemisphere the land-masses are much hotter than the sea. They become areas of low pressure and the southern tropical high-pressure belt is broken over South America, Africa and Australia.

In the northern summer, which will be the southern winter, the conditions are reversed. The high-pressure belt is interrupted over the northern continents but spreads outwards over the southern continents. Owing to the small extent of the latter the effect is much less marked than in the northern hemisphere.

These changes in the position and extent of the pressure belts are shown diagrammatically in Figs. 22 and 23, in which

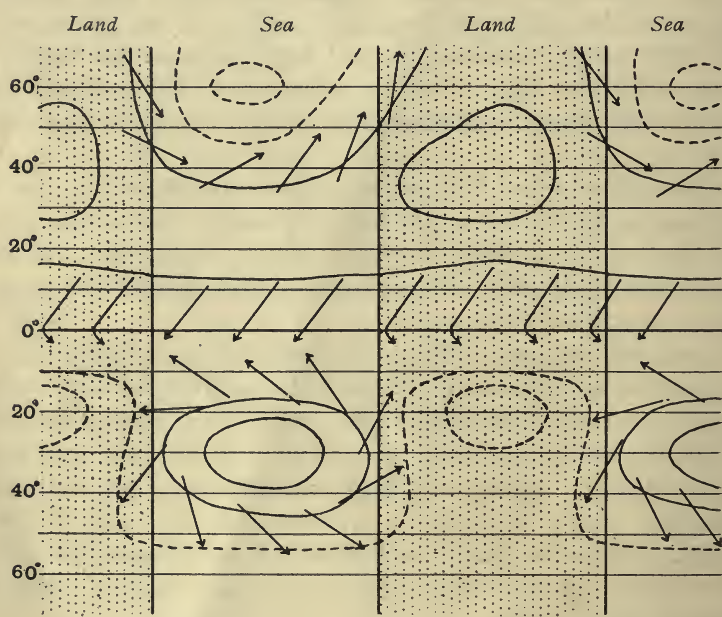


FIG. 22. Diagram of the distribution of pressure over land and sea in January.

Isobars over 30 inches shown as continuous lines; isobars under 30 inches as broken lines.

the land and sea are supposed to stretch from north to south in alternate bands.

Since the continental masses alter the distribution of pressure to so great an extent they will have a corresponding effect upon the winds. The winds will always blow away from the high pressures but with the deflection proper to each hemisphere, and their directions are shown in the diagrams.

Besides the change due to the migration of the belts from north to south and back again, the most notable seasonal variation is that upon the eastern margins of the continents. Here the winter winds are directed towards the equator, and the summer winds towards the pole. Their precise direction and their extent in latitude will depend upon the form of the continental mass. The change is actually most noticeable in south-eastern Asia, where it gives rise to the "monsoons";

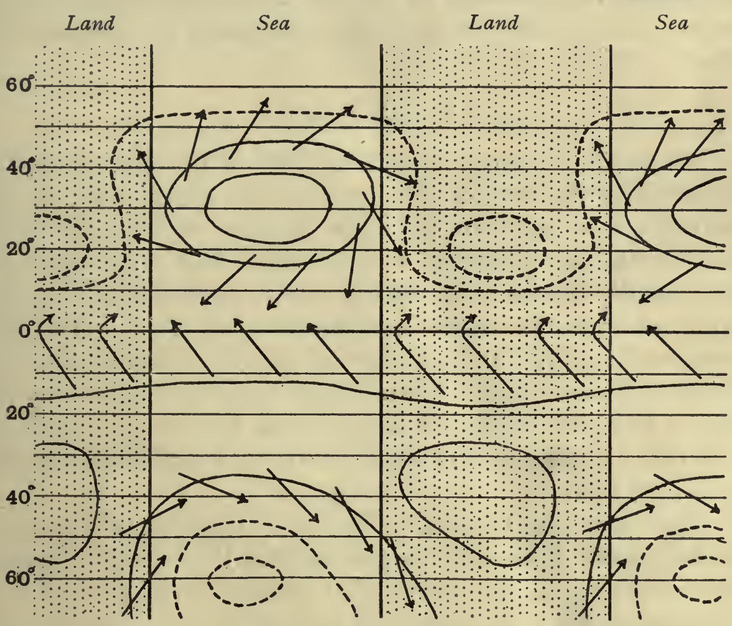


FIG. 23. Diagram of the distribution of pressure over land and sea in July. Isobars over 30 inches shown as continuous lines ; isobars under 30 inches as broken lines.

but a similar effect is observable in the south-eastern part of the United States and elsewhere.

It will be seen from the diagrams that apart from the effect of the continents there would be somewhat similar seasonal variations of the wind near the equator, because the line of lowest pressure lies south of the equator in our winter, north of the equator in our summer. In our winter therefore the north-east trades are carried across the equator and are then deflected to the left. In our summer the south-east trades are carried north of the equator and are then deflected to the right. But if the globe were covered with water the effect would be limited to a narrow belt extending only a short distance on each side of the equator; and, since the deflecting force near to the equator is very small, the trade-winds when they cross the line would deviate but little from their original directions. This is seen to be the case in the middle of the Pacific, where there are no large land-masses to intensify the effect.

Local Winds

The winds of the general circulation of the atmosphere, as described in the preceding pages, are the consequence of the decrease of temperature from the equator to the poles combined with the rotation of the earth. But there are also local winds due to local differences of temperature. In the strict sense of the term the monsoons are local winds, for they are caused by the interference of the continental masses with the general fall of temperature from the equator to the poles. But they are on so large a scale and cover so wide an area that in the ordinary sense of the word they can hardly be called local.

Land and sea breezes. The most familiar of the local winds are the land and sea breezes which at certain seasons and in certain regions blow day by day with the greatest regularity. Even in so variable a climate as that of England they may be felt during settled weather.

They are the result of the unequal heating of land and water. It is well known that when the sun's rays fall with equal intensity upon land and sea, the temperature of the land rises more quickly than that of the sea. But, on the other hand, after the sun has set the land cools more quickly than the sea.

Consequently during the day, while the sun is shining the land becomes a hot area compared with the sea. The circulation described on p. 39 is set up; an upper current of air flows from land to sea and on the surface a lower current from sea to land. ~~The latter~~ is ~~the sea breeze~~. It does not begin to blow as soon as the sun has risen, for at night the land has become cooler than the sea. It is not till the sun has had time to raise the temperature of the land above that of the sea, that these currents start. Moreover it has already been shown that the establishment of the circulation between cold and hot areas takes time. It begins on a very small scale and extends gradually, if the heating continues, both horizontally and vertically. On the coast the sea breeze often blows in the forenoon, inland it may not be felt till a much later hour.

In the afternoon when the sun has sunk so low that the heat received from it is less than the heat lost by radiation, the sea breeze begins to die away, but it does not cease entirely until the temperature of the land has fallen to that of the sea.

The land still continues to cool more rapidly than the sea, and consequently the sea becomes a hot area compared with the land. A reverse circulation is then set up; the upper currents flow from the sea towards the land, the lower currents from the land towards the sea. It is the latter that are felt on the surface of the ground, and they form the land breeze, which is experienced in the evening and at night.

The length of time during which either the land or the sea remains hotter than the other is only a few hours. Consequently the height to which the circulation of air extends is not great. Observations with pilot-balloons at Coney Island near New York showed that the sea breeze was felt only to an altitude of 500 or 600 feet.

Land and sea breezes can only be experienced when the sky is fairly free from cloud, for otherwise the sun will not warm the land to any great extent by day nor will much heat be lost by radiation at night. They are strongest when the diurnal range of temperature is greatest. In England they are seldom noticed excepting in the summer, for in the winter the sun has very little power, even at noon. Their strength is never very great and consequently they are easily overpowered by the winds due to other causes.

The conditions favourable to the production of land and sea breezes are therefore a clear sky and the absence of other winds. These are the characteristics of anticyclones, and land and sea breezes are accordingly most marked in anticyclonic areas, whether these are temporary or permanent.

Lake breezes. In the neighbourhood of large lakes, off-shore and on-shore breezes are sometimes produced in precisely the same way. They do not differ in any essential respect from land and sea breezes but, in correspondence with the comparatively small area of the water surface, they are usually weaker.

Mountain and valley winds. In mountainous regions winds are often caused by the unequal temperature of the free air and of the air in contact with the mountain sides. But the way in which the difference is brought about will be more readily understood after the vertical distribution of temperature has been explained, and these winds will therefore be dealt with in the chapter on that subject.

CHAPTER V

THE HORIZONTAL DISTRIBUTION OF TEMPERATURE

Measurement of temperature. The temperature is measured by means of the thermometer. On the continent the Centigrade thermometer is usually employed, but in English-speaking countries the Fahrenheit thermometer is still used for most meteorological observations.

The temperature of a place is the temperature of the air within a few feet of the ground at that place. The temperature of the ground itself may be distinctly different.

It seems an easy matter to take the temperature of the air; but it is not so easy as it seems. If a thermometer is hung against the wall of a house it will be affected by the temperature of the wall as well as by that of the air. If it is hung from some support by a string, so that it is completely surrounded by free air, away from any buildings or trees, it will be heated by the direct rays of the sun, and even when the sun is not shining it will be influenced by radiation from the earth.

There are two ways in which the true temperature of the air may be obtained with considerable accuracy. The thermometer may be protected from all external radiation, by a screen of some kind which allows free access to the air. The form of screen which is most commonly employed in England is known as Stevenson's. It is a kind of wooden box, raised above the ground on posts. To protect the thermometer from the rays of the sun the roof is double, with an air-space between the two boards which form it. Radiation from the earth is intercepted by the floor, which may be double also. The sides are louvred, like Venetian shutters, thus keeping out the sun's rays while allowing the air to enter freely. The thermometer is supported on a frame inside, so that its bulb is near the middle, as far as possible from the roof and floor and sides.

Such a screen is not very portable and explorers generally use a "sling thermometer." This is nothing more than a simple unmounted thermometer, like an ordinary chemical thermometer, with a piece of string tied to the ring at the upper end. By means of the string it is swung round and round at a moderate rate in the open air. In a minute or two it will attain the temperature of the air, even if the sun is shining. While swinging, it is brought continually into contact with fresh particles of air, and its temperature is therefore kept down very nearly to that of the air, in spite of the heating effect of the sun's rays. If the sun is very strong, the thermometer may be shaded with an umbrella while it is being swung and read.

Isotherms. The distribution of temperature is shown on maps by means of "isotherms." An isotherm is a line along which the temperature is everywhere the same. The method of drawing them is the same as the method of drawing isobars.

Temperature as well as pressure varies with the altitude, though not so regularly. In a hilly district the isotherms often run nearly along the contour lines. If therefore we wish to consider the effects of other causes than altitude upon the distribution of temperature, the thermometer readings must be "corrected for altitude" just as the barometric readings are corrected in isobaric maps. Unfortunately the allowance to be made for altitude is less certain in the case of temperature and is undoubtedly influenced by the form of the ground. The average fall of temperature with increase of altitude is about 1° F. for 320 feet. If, then, the thermometer at a place 640 feet above sea-level stands at 52°, the temperature "reduced to sea-level" would be 54°.

In almost all isothermal maps the temperatures have been corrected for altitude; and in such maps an isotherm is a line along which the temperature, reduced to sea-level, is everywhere the same.

Sources of heat. There are two possible sources from which the warmth of the earth's surface may be derived. The first of these is the sun, the effects of which are evident to everyone. The second is the interior of the earth.

Seasonal variations of temperature are not felt below the ground to a greater depth than 60–80 feet, and beneath that limit mine-shafts and well-borings show that the temperature increases downwards. The rate of increase varies greatly in different places, and even at different depths in the same shaft; but a large series of observations gave an average rise of about 1° F. for each 64 feet of descent. If this rate is maintained the interior of the earth must be excessively hot; and in any case, since it is certainly hotter than the exterior, it must warm the surface to some extent. It is impossible to determine how great its influence may be, but since it must affect the poles as much as the equator it is evidently very small compared with that of the sun, and for practical purposes it may be neglected.

A certain minute amount of heat must also be received from the stars and, by reflection of the sun's heat, from the moon, but the quantity is so small that it is barely appreciable even by the most delicate instruments.

Consequently the only source of heat that need be considered is the sun.

Insolation. It is convenient to have a term to express the amount of heat received from the sun and the term employed is insolation. The insolation for a day is the total amount of heat received from the sun during that day. The expression "intensity of insolation" has nearly the same meaning as the common phrase "strength of the sun" or "strength of the sun's rays"; but it is used more definitely. If we say that at a certain time the intensity of insolation at one place is twice that at another, we mean that during equal very short periods of time the former place receives twice as much heat (per square foot or per square inch) as the latter.

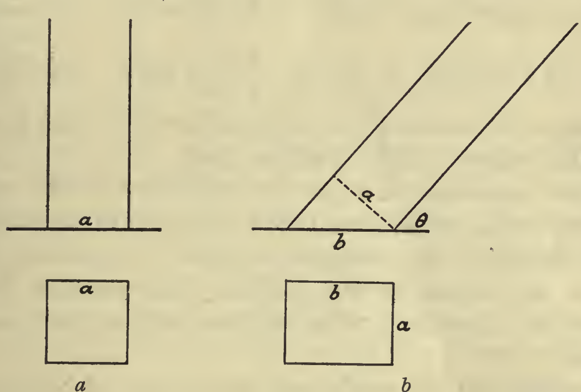


FIG. 24. Areas illuminated by vertical and inclined beams.

Insolation on an airless and waterless globe. If the earth were a solid globe of uniform composition, without water and without an atmosphere, it would be a comparatively simple matter to calculate the effect of the sun's heat upon its surface.

If a beam of light in section one foot square falls vertically upon a plane horizontal surface, its light and heat will cover an area of one square foot (Fig. 24 *a*). But if the same beam falls obliquely its light and heat will be spread over a larger area (Fig. 24 *b*). For simplicity we may suppose that in both figures the sides of the beam are respectively parallel and perpendicular to the paper. In the former case the area illuminated will be a square, with each side one foot in length; in the latter it will be an oblong with a breadth a of one foot but with a length b greater than one foot, and its area will be b square feet. As the angle of altitude of the sun, θ, decreases the length of b increases and the area of the oblong increases in the same proportion.

But since the total amount of heat received from the beam remains the same, the heat per square foot is evidently inversely proportional to the area illuminated, that is, the intensity of insolation is inversely proportional to b, or directly proportional to $\frac{1}{b}$.

In the upper part of Fig. 24 b, $\frac{a}{b} = \sin\theta$, and since the width a of the beam is 1 foot, $\frac{a}{b} = \frac{1}{b} = \sin\theta$. The intensity of insolation is therefore proportional to $\sin\theta$, that is, to the sine of the angle of altitude of the sun.

At the equinoxes the sun is vertical over the equator; elsewhere its altitude at noon is the complement of the latitude, and the intensity of insolation at noon is therefore proportional to the cosine of the latitude. Throughout the day, the altitude is greatest at the equator and decreases towards the poles.

Accordingly at the equinoxes the amount of heat received from the sun will be greatest at the equator and will decrease gradually to the poles.

At the summer solstice the sun is vertical over the Tropic of Cancer and it may be thought that the insolation will be greatest there. But a new factor is now introduced. At the equinoxes the day is 12 hours long and the night is 12 hours long all over the earth; but at the solstices the length of the day varies with the latitude, as shown in the following table:

Lat.	0°	10°	20°	30°	40°	50°	60°	66½°
	h. m.	h. m.	h. m.	h. m.	h. m.	h. m.	h. m.	h. m.
Longest day	12 0	12 35	13 13	13 56	14 51	16 9	18 30	24 0
Shortest day	12 0	11 25	10 47	10 4	9 9	7 51	5 30	0 0

Consequently, although at our summer solstice the sun will be hotter at the Tropic of Cancer than in lat. 60° N., it will not be above the horizon for so long a time. The length of the day more than compensates for the obliquity of the sun's rays and the amount of heat received during the day is greater in lat. 60° N. than at the Tropic.

The calculation of the insolation at different latitudes in these circumstances is rather difficult, even on the assumption that the atmosphere has no influence; and only the results need be given here. Taking the insolation at the equator for March 21 as 1000, the insolation at various latitudes on June 21 is shown in the following table:

Latitude N.	0°	10°	20°	30°	40°	50°	60°	70°	80°	90°
Insolation on June 21	881	975	1045	1088	1107	1105	1093	1130	1184	1202

According to this table the insolation on June 21 is greatest at the North Pole, and in England it is greater than in the Sahara. Such results appear to be totally at variance with actual observations; but there are two points which should be borne in mind. In the first place the table gives the total amount of heat received during the day, which is quite a different thing from the temperature attained. Secondly, the table takes no account of the absorption of the sun's rays by the atmosphere. It applies to a globe without an atmosphere, or, in the case of the real earth, to the external surface of the atmosphere and not to the surface of the ground.

Influence of the atmosphere on insolation. Although the air seems so transparent, it does not allow the rays of the sun to pass through it without loss. Some of the light and heat is absorbed and does not reach the earth. The absorption appears to be due chiefly to the water-vapour and carbon dioxide in the air.

It is evident at once from Fig. 25 that where the rays are oblique the length of their journey through the atmosphere is greater than where they are vertical. Consequently the loss of heat is greater and their power at the earth's surface is less. The atmosphere in fact greatly increases the effect of obliquity, and the intensity of insolation does not vary with the sine of the altitude of the sun but much more rapidly.

It is difficult to determine how much of the heat is lost in this way, but in the case of vertical rays it is believed to amount to one-third of the total.

Since the atmosphere increases the effect of obliquity it increases also the effect of latitude; and taking the average temperature for the whole year, the poles are colder compared with the equator than they would be if there were no atmosphere.

Influence of land and water. Taking the whole year round, the amount of heat received from the sun is greatest at the equator and decreases gradually towards the poles.

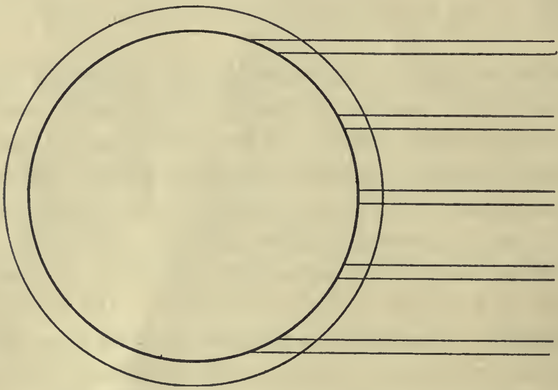


FIG. 25. Influence of the atmosphere on insolation.

But the equator is not the hottest part of the globe. The "thermal equator," or line of maximum temperature, lies for the most part north of the real equator.

The principal reason for this and other irregularities is that heat affects land and water differently, and therefore the distribution of temperature is greatly influenced by the distribution of the continents and oceans.

In general, the land-masses heat more quickly in the summer and cool more quickly in the winter than the seas; and there are several reasons for this difference between them.

(1) The specific heat of water is much greater than that of the solid earth. It takes nearly five times as much heat to raise the temperature of a pound of water 1°, as it does to raise the temperature of a pound of sand 1°. On the other hand, while its temperature is falling 1°, the pound of water gives out nearly five times as much heat as the pound of sand.

The amount of the sun's heat received by a land-mass or a sea is affected by its area and not by its weight, and therefore it is the specific heat by volume rather than by weight that concerns us; and the difference in this is not quite so great. It requires about twice as much heat to raise the temperature of a cubic foot of water 1° as it does to raise the temperature of a cubic foot of sand 1°.

(2) The rays of the sun penetrate more deeply into water than into earth. The daily variations of temperature are not perceptible below a depth of about 3 feet in earth; in water they may sometimes be observed at 60 feet beneath the surface. The seasonal variations disappear at a depth of 60 to 80 feet in earth, of 300 to 600 feet in water.

Thus the sun's rays falling on water are engaged in heating a much larger volume of material than when they fall on land. The temperature reached is therefore not so high; but when the sun sets there is a larger quantity to cool and the cooling is slower.

(3) Water is mobile and land is not. When the sun's rays fall on land their effect is practically limited to the area on which they fall; in water, convection currents are set up and the heat is partly carried away. Consequently when the sun's rays fall on water they warm not only the area on which they fall but also the surrounding parts.

(4) When the sun shines on water a considerable proportion of the heat is used in evaporating the water and not in raising its temperature. On dry land there is no such loss. In marshes and swamps a certain amount of heat will disappear in the same way.

(5) Much of the heat that falls upon the water is reflected and does not raise its temperature. Land surfaces are poor reflectors and but little heat is lost in this way.

(6) On the whole the sky is cloudier over the ocean than over the land. The clouds obstruct the rays of the sun, but they also hinder loss of heat by radiation. Their effect is therefore to retard both the heating and the cooling of the water.

All these causes tend to make the water heat more slowly than the land, and all excepting the fourth and the fifth make it cool more slowly.

Distribution of temperature in the British Isles. The difference between land and sea is very clearly shown by the distribution of temperature in the British Isles in summer and in winter (Fig. 26).

In the July map it will be seen that the general direction of the isotherms is from east to west, but they bend northwards over the land, southwards over the sea. The highest temperature is in the neighbourhood of London. In the January map the general direction of the isotherms is roughly north and south, with the highest temperatures in the west; they bend southwards over the land, northwards over the sea.

In July the sun is high for a large part of the day, and the nights are short. More heat is received from the sun during the day than is lost at night by radiation, and the temperature is therefore rising, both on land and sea. It rises most rapidly in the south, where the sun is most powerful, and consequently the general direction of the isotherms is from east to west. But the temperature of the land rises more quickly than that of the sea and therefore the land becomes hotter than the sea in the same latitude. The highest temperature is not in the extreme south, because the coast is cooled by the waters of the English Channel; it is a little inland, about the neighbourhood of London. Moreover, although the general direction of the isotherms is from east to west, they bend towards the pole over the land, towards the equator over the sea.

In January the sun is always low, even in the middle of the day, and its rays accordingly have little power. The days are short and the nights are long. More heat is lost at night than is received by day, and both land and water are therefore cooling. But the water cools more slowly than the land and by the month of January the temperature of the land is

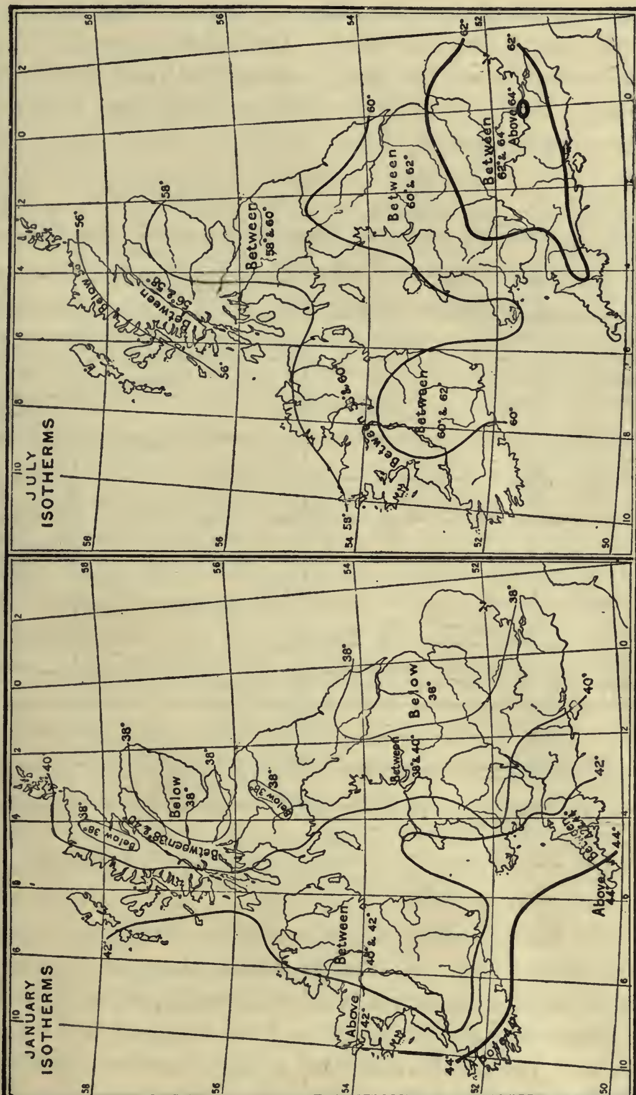


Fig. 26. Isotherms of the British Isles.

considerably less than that of the sea. So great is the difference and so weak are the sun's rays at this season that the Atlantic is now a more important source of heat than the sun. Consequently the isotherms in January run roughly parallel to the Atlantic coast, and the warmest part of the British Isles is not the south, but the west. The Irish Sea and North Sea are also warmer than the neighbouring land and the isotherms accordingly bend towards the pole over the sea, towards the equator over the land.

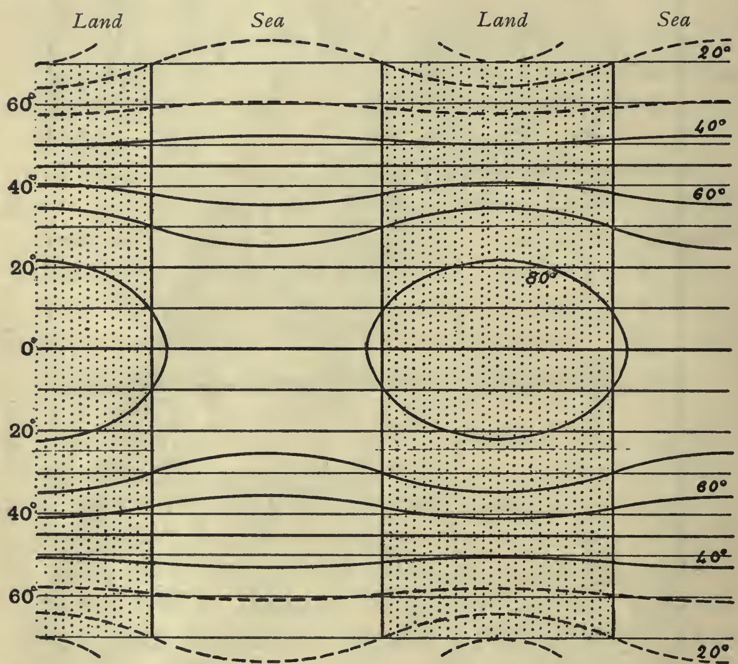


FIG. 27. Influence of continents and oceans on the general distribution of temperature.

In the British Isles, therefore, when the sun at noon is high up in the sky, the land is hotter than the sea and the isotherms bend equatorwards over the sea, polewards over the land; when the noonday sun is low, the land is cooler than the sea and the isotherms bend equatorwards over the land, polewards over the sea.

General distribution of temperature over the globe. This is a general principle and applies to the whole earth. Towards the equator, where the sun at noon is always high, the conditions are those of the English summer; towards the poles, where the sun is always low, the conditions are those of the English winter. Near the equator therefore the isotherms bend polewards over the land, equatorwards over the sea; towards the poles the bends are in the opposite directions. At some intermediate latitude, land and sea will be at about the same temperature and the isotherms will be nearly straight The general distribution of temperature may accordingly be represented as in Fig. 27, in which land and sea stretch north and south in alternate bands. The temperatures in the middle of the ocean are those of the middle of the South Pacific in the same latitudes; the temperatures in the middle of the land strips are taken in a similar way from those of the great continents of the eastern hemisphere.

On comparing this diagram with a map of the world showing the actual annual isotherms (Map 4), it will be seen that there is a general correspondence between the two. If, for example, we trace the isotherm of 30° in the northern hemisphere from west to east, we find that it bends polewards over the Pacific, equatorwards over North America, polewards over the Atlantic and again equatorwards over the Eurasian continent. In the southern hemisphere the bends are much less marked, because the land-masses are narrow.

Near the equator on the other hand the isotherm of 80° in the northern hemisphere bends towards the equator over the Atlantic, towards the poles over America and Africa.

About lat. 45° N. the isotherm of 50° is comparatively little affected by the distribution of land and sea, and does not deviate far from the parallel of latitude, indicating that the difference of temperature between land and sea is small.

Effect of the prevalent winds. Thus the diagram represents very fairly the actual distribution of temperature over land and sea. But even if we allow for the difference in the form of the continents, it is not perfect. In the diagram the isotherms of the north temperate zone are farthest from the equator in the middle of the ocean and nearest in the middle of the land; in the map they are farthest on the western coasts of America and Europe, they are nearest close to the eastern coasts of America and Asia.

This is the effect of the prevalent winds. There are two ways in which the distribution of temperature is influenced by the winds. A wind is a movement of the air from one place to another and it therefore carries, as it were, the temperature of one place to another. Neither the interchange of air nor the transference of temperature is complete; but it is sufficient to be of great importance.

The winds also blow the surface waters of the ocean in the direction in which they themselves are travelling. Without winds the coldest and densest water would be at the bottom, the warmest and lightest at the top, and the isothermal surfaces in the water would be horizontal as in Fig. 28. When

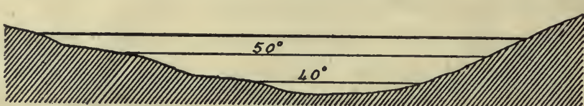


FIG. 28. Isotherms in a lake of still water.

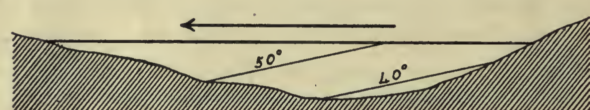


FIG. 29. Isotherms in a lake when a wind is blowing.

a wind blows in the direction of the arrow in Fig. 29 the warm surface water is carried to the windward shore, colder water comes to the top on the leeward shore, and the isotherms are displaced as shown. The effect is to raise the temperature of the shore towards which the wind is blowing and to lower the temperature of the opposite shore.

North of lat. 40° N. the prevalent winds are from the south-west. They therefore blow the surface waters of the ocean towards the east, warming the western coasts of the continents and cooling their eastern coasts. Moreover, on the western coasts the south-west winds are coming from the warmer sea; on the eastern coasts they are coming from the colder interior of the land. Therefore in this way also they raise the temperature of the western coasts and lower the temperature of the eastern coasts.

If, then, we wish to alter the diagram (Fig. 27) so as to show the effect of the winds, we must place the isotherm of 40° nearer to the equator on the eastern shore of the continent and farther away from it on the western shore; and similarly with all the other isotherms within the region of the south-westerly winds. Moreover, since the western side of the

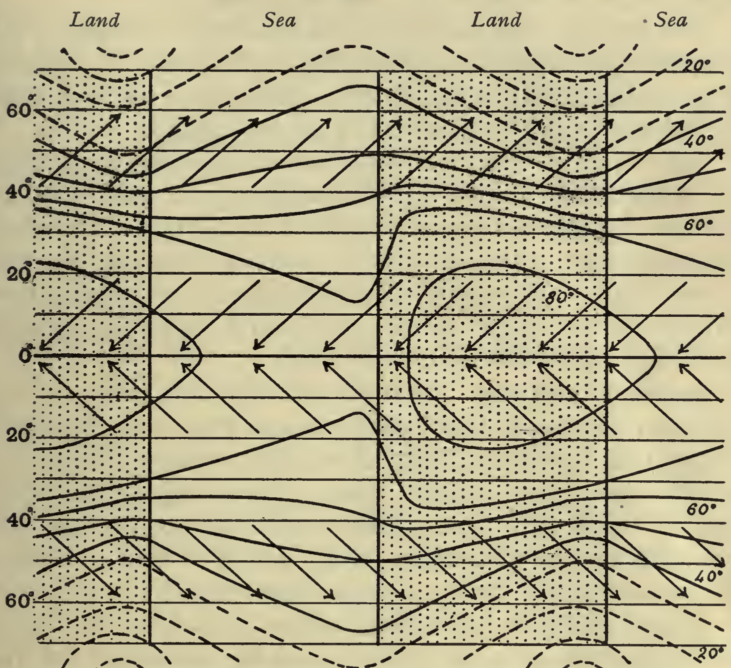


FIG. 30. Influence of the winds on the course of the isotherms.

continent is warmer than the east, the minimum temperature will be displaced eastward. The general result of these modifications is shown in Fig. 30.

Between the tropics the prevalent winds are the trade-winds. There is not the great difference of temperature between land and sea that there is in temperate latitudes and consequently the direct effect of the winds is slight. But they blow the surface waters towards the west, warming the eastern shores of the continents, while the western shores are cooled by the upwelling of the colder water from below. Within the tropics generally the eastern shores of the continents are therefore warmer than they would be without the winds, the western shores are cooler. To make the proper alteration in the diagram (Fig. 27) we must place the isotherms farther from the equator on the eastern coasts of the land-masses, nearer to the equator on the western coasts.

But because the prevalent winds on the western coasts blow from the land, the cooling effect is limited to a comparatively narrow strip close to the coast, and the isotherms accordingly take the form indicated in Fig. 30.

This figure is now complete and shows diagrammatically the general distribution of temperature over land and sea when we take into account the effect of the prevalent winds. Comparison with the map of the actual isotherms discloses a very close agreement. The differences are due almost entirely to the form of the continental masses, and their effect upon the direction of the winds.

Seasonal variations. Both in the diagram (Fig. 30) and the map (Map 4) the isotherms are what are known as "annual isotherms." They show the average temperature for the whole year, of the places through which they pass. But isotherms may also be drawn showing the average temperatures for a month or any other period of time, and these will usually differ from the annual isotherms.

In the northern hemisphere generally, July is the hottest month and January the coldest, and the isotherms for July may therefore be taken as showing the general distribution of temperature in summer and the isotherms for January as showing the general distribution in winter (Maps 5 and 6).

It will be seen at once that the seasons bring great changes not only in the actual temperature, but also in the distribution of temperature. The course of the July isotherms is very different from that of the January isotherms or of the annual isotherms.

The general nature of the changes, and the reason for the changes, will be most easily understood if we take the diagram of annual isotherms in Fig. 27, and consider how it is likely to be modified in July.

In this diagram the part of the continent north of lat. 45° N. is colder than the sea. But in July it will not be so. It is then the northern summer, and the northern continents will be warmer than the seas. The bends of the isotherms will therefore be reversed and will be directed southwards over

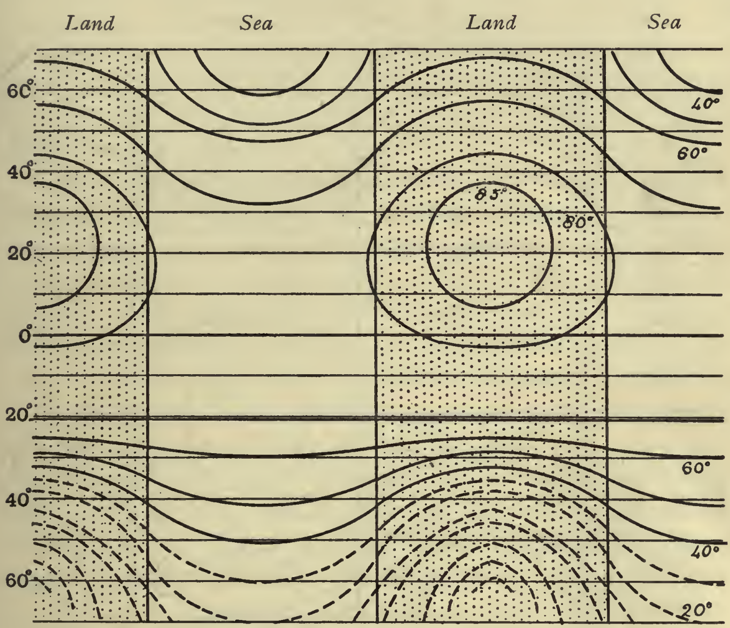


FIG. 31. Diagram of general distribution of temperature in July.

the sea, northwards over the land The lowest temperature will be found on the ocean and not on the land.

Within the tropics there will be comparatively little change. There the land is always warmer than the sea. In July this difference will be increased north of the equator and somewhat decreased south of the equator. The maximum temperature will be a little north of its position in the annual diagram.

In the southern hemisphere July is the middle of winter. Beyond the tropics the land will be colder than the sea. Even if we take the average temperatures for the year it is colder, as is shown in the diagram (Fig. 27) and map (Map 4) of the annual isotherms. But in winter the difference is intensified and accordingly the bends in the isotherms will be in the same direction as in the annual diagram but will be more pronounced.

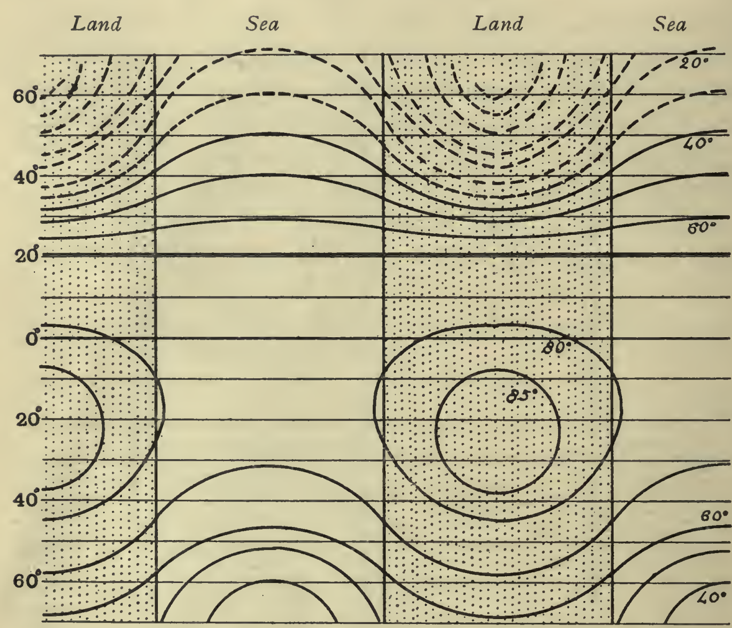


FIG. 32. Diagram of general distribution of temperature in January.

The effect of all these modifications is shown in Fig. 31 which is a diagram of the isotherms for July, just as Fig. 27 is a diagram of the annual isotherms.

It is hardly necessary to consider in detail the diagram for January (Fig. 32). It is the northern winter and the southern summer. The conditions of the two hemispheres are interchanged and the diagram is like that for July reversed. In fact, if a mirror stands along the northern (or southern) edge of the July diagram, the reflection in the mirror will be exactly like the diagram for January.

Range of temperature. The difference between the summer and the winter temperatures of any locality is known as the annual range of temperature at that locality. But the phrase will have different meanings according to the definition of the terms summer and winter temperatures.

Strictly speaking, the range is the difference between the highest and lowest temperatures ever experienced at the place in question. This is the absolute range of temperature. But the maximum and minimum will not be the same every year; and if we take their averages for a series of years, the difference may be called the mean annual extreme range, or simply the annual extreme range.

The maximum and minimum temperatures are exceptional and endure for a short time only. Perhaps a better idea of the difference between summer and winter is given by comparing the average temperatures of the hottest and the coldest months. The difference between these is the range of monthly mean temperatures, and it is this difference that is most commonly called the annual range or the mean annual range.

In the British Isles July is in most places the hottest month and January the coldest, so that the difference between the temperatures of these months is the annual range. In July the temperature is highest towards the south and decreases northwards; in January it is highest towards the west and decreases eastwards. It follows from this that the annual range is not everywhere the same. It is least towards the north-west, where the January temperature is high and the July temperature low; it is greatest towards the south-east where the January temperature is low and the July temperature high.

The increase of range towards the east is continued right across the continent of Europe and almost to the eastern shores of Asia. The following table gives the longitude, the mean annual temperature, the January and July temperatures, and the annual range of a number of places, all of which are in nearly the same latitude.

	Lat. N.	Long. E.	Mean Ann. Temp.	Jan.	July	Range
Cambridge ..	52° 13′	0° 6′	48·6°	37·6°	61·5°	23·9°
Utrecht ..	52° 6′	5° 11′	48·0°	34·2°	62·6°	28·4°
Hanover ..	52° 22′	9° 45′	47·1°	32·7°	63·1°	30·4°
Berlin ..	52° 30′	13° 23′	47·3°	31·3°	64·6°	33·3°
Posen ..	52° 25′	16° 56′	46·6°	29·3°	65·5°	36·2°
Warsaw ..	52° 13′	21° 0′	45·1°	25·9°	65·8°	39·9°
Tambov ..	52° 44′	41° 28′	40·8°	11·3°	68·9°	57·6°
Irkutsk ..	52° 16′	104° 19′	31·3°	−5·4°	65·1°	70·5°

It should be noted that Irkutsk stands at a considerably higher level than any of the other towns in the list.

A similar increase in range is found in all continental masses in the temperate regions. Not only are the western coasts warmer (on the average for the year) than the eastern, but they also have a much more uniform temperature. The western coasts are influenced by the winds from the oceans and accordingly they share in the uniformity of temperature characteristic of large masses of water.

It is in north-eastern Siberia that the greatest ranges of temperature have been observed. The range of monthly means at Yakutsk amounts to 112·1°; the annual extreme range to 158·0°, and the absolute range (in 32 years) to 185·8°. Ranges almost as great as these have also been recorded in the north-eastern parts of Canada. In the British Isles the greatest range of monthly means is only about 26°; in the western part of Ireland it is about 15°.

In equatorial regions the range is very small. At Jaluit in the Marshall Islands the difference between the hottest and the coldest months is only 0·8°. Here the climate is influenced by the sea, but even at Equatorville on the Congo, in the middle of Africa, the difference is no more than 2·2°.

Towards the tropics, where the air is dry and the sky is clear, the range increases, but it is small compared with the ranges in Eastern Siberia and Canada. It is in fact in the so-called Temperate Zone that the greatest extremes of temperature are experienced.

Construction of range-maps. In general, maps showing the range of temperature are drawn by plotting the observed ranges at different points and proceeding as in the case of isotherms. But if, for any area, January is everywhere the coldest month and July the hottest, it is possible from the isothermal maps for these months to construct a map of annual ranges. In Fig. 33 the isotherms for July are shown as thin continuous lines, the isotherms for January as thin broken lines. Where the July isotherm of 60° crosses the January isotherm of 38°, the range is the difference between these temperatures, namely 22°. Similarly at every intersection the range can be at once determined. By drawing lines joining points where the range is the same, we obtain lines of equal range, or "range-lines" as they may be called. But these connecting lines are not necessarily straight, and there are certain rules which must be obeyed in drawing them.

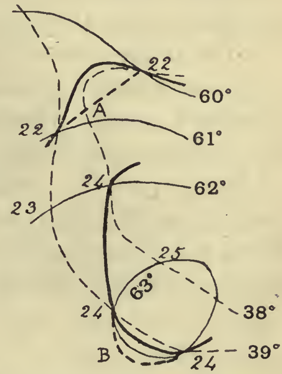


FIG. 33. Construction of range-maps.

The range-lines must not cross an isotherm except at the intersections. For example, the range-line for 22° must be drawn as shown by the heavy continuous line and not as indicated by the heavy broken line. It cannot cross the isotherm of 38° at *A*, for at that point the winter temperature is 38° and the summer temperature is more than 60°; the range is therefore more than 22°. The range-line for 22° can only cross the January isotherm of 38° at its intersection with the July isotherm of 60°.

The range-lines must cross *both* isotherms at the intersections. The range-line for 24° must not be continued in the direction shown by the heavy broken line at *B*, which crosses the January isotherm of 39° but only touches the July isotherm of 63°. It cannot lie at *B*, for there the January temperature is more than 39°, and the July temperature is less than 63°; the range is therefore less than 24°.

The map is divided into a number of spaces by the crossing of the January and July isotherms. In drawing the range-lines in the manner described, two range-lines of different value must never lie in the same space. It is perhaps unnecessary to give the proof of this rule. It is useful sometimes towards the edge of the map, where there may be a doubt as to which way the range-line is to cross an intersection.

The range-lines of the British Isles obtained in this way from the isotherms for January and July agree very fairly with the observed annual ranges, because in most places January is the coldest month and July the hottest. Even in the case of large areas the method gives a good idea of the difference between winter and summer; but unless all parts of the area have the same month as their hottest and the same month as their coldest, it will not give the real annual range.

CHAPTER VI

VERTICAL DISTRIBUTION OF TEMPERATURE

The vertical gradient of temperature. On the top of a mountain it is almost always colder than at the foot. Similarly, in balloon ascents it is usually found that the higher the balloon rises the lower the temperature falls. It is only now and then that there is any exception to this rule, even for a short distance.

Yet on the top of a mountain the rays of the sun have passed through a smaller thickness of atmosphere than they have when they reach the foot; and consequently we might expect them to be more powerful and to produce a higher temperature. They really are more powerful. High up in the mountains, when the sun is shining, the face becomes sunburnt more quickly than on the plains; and in taking photographs the exposure required is distinctly less. But in spite of this the temperature of the air is lower. The only possible conclusion is that it is not the rays of the sun that warm the air.

The rate of fall of temperature with increase of height is called the vertical gradient of temperature, or simply the temperature gradient. It varies from time to time and from place to place, but by taking the mean of many series of observations it is found that on mountain slopes the average vertical gradient is about 1° F. for each 320 feet of ascent. In the free air the average rate of decrease near the earth's surface is probably less than this.

Occasionally it happens that there is an increase of temperature as we go upwards. The vertical gradient is then said to be reversed. But such reversals do not often extend very far.

The air is heated chiefly from below. Air is not readily warmed by radiant heat, that is, by the heat which radiates outwards from a hot body. A fire may feel hot to the face, while the intervening air is still quite cold. The radiant heat from the fire warms the face, but has comparatively little effect upon the air through which it passes. In the same way the rays of the sun may heat the earth, while they hardly affect the air through which they travel.

It is true that in passing through the entire thickness of the atmosphere a considerable proportion of the light and heat from the sun is absorbed and is in fact spent in raising the temperature of the atmosphere. The total amount of heat absorbed is large, but it is spread through so great a mass of air that the increase of temperature is small.

The air is not equally transparent to all radiant heat. It is less transparent to the ultra-red rays than to those of the visible part of the spectrum. The former give heat without light, and are the only rays emitted by a body that is warm but not incandescent, such as the earth itself. Thus the air absorbs a larger proportion of the heat radiated out from the earth than of the heat which passes through it from the sun. For this reason the air is warmed by the earth rather than by the rays of the sun.

But there is another reason, and a more important one, why the air receives its heat principally from the earth. Although the air is comparatively little affected by radiant heat passing through it, it is warmed at once by actual contact with anything hotter than itself, and cooled by contact with anything colder.

Thus the air is but little affected by the direct rays of the sun. It is warmed to some extent by radiant heat from the earth and still more by contact with the earth. Therefore it is warmed from below and it is natural that the temperature should decrease upwards.

Experimental illustration. A very simple experiment which may be made on any still and sunny day, will show how little direct effect the rays of the sun have upon the temperature of the air. Hang a thermometer in the open air, freely exposed to the sun. The mercury rises and the thermometer soon shows a decidedly high temperature. Now swing the thermometer round and round, still in the open air and still in the sun. Its temperature at once falls, although it is just as much exposed to the sun's rays as before. The reason for the difference is that although the sun warms the thermometer it does not warm the air. The air therefore is colder than the thermometer, and when the thermometer is swung rapidly, so as to bring it continually into contact with fresh air, its temperature falls to that of the air. When, on the other hand, the thermometer is hanging quietly, the sun warms the thermometer, and the thermometer warms the air immediately in contact with it.

If the day is windy, the experiment cannot be performed unless the thermometer is sheltered in some way from the wind, for the wind produces the same effect as the swinging of the thermometer. Even a gentle breeze is sufficient to cause a considerable fall in the temperature of the hanging thermometer. In order to protect it from the wind the thermometer may be hung in a flask, but in that case the experiment is not quite so satisfactory, for the hanging thermometer and the swinging thermometer are not then exposed in quite the same way to the rays of the sun.

The upward movement of changes of temperature. When the sun shines, the earth is heated and warms the air in contact with it. The heat spreads slowly upwards, partly by conduction from one layer of air to the next, partly by convection, the heated air rising and its place being taken by colder air from above. The earth being the principal source from which the air derives its heat, the temperature will decrease upwards.

If the sky remains clear after the sun has set, the earth rapidly loses heat by radiation and its surface becomes colder than the air. It cools the air in contact with it. The lowest layer of air then cools the next, and in turn each layer loses heat to the one below it and the cold spreads upwards. In this case the temperature upon the ground is low, and there is a partial reversal of the temperature gradient.

That this is the way in which the air is warmed and cooled is shown by the observations made on the Eiffel Tower. The Tower is 300 m. high (rather less than 1000 feet), and there are observing stations at the bottom, at the top, and at two or three intermediate heights. The following table shows the average temperature (in degrees Fahrenheit) at various hours of the day in the months of December and July, at the lowest and the highest stations:

Mid night	2 a.m.	4 a.m.	6 a.m.	8 a.m.	10 a.m.	Noon	2 p.m.	4 p.m.	6 p.m.	8 p.m.	10 p.m.	Mean
					DECEMBER.	Lower Station						
33·8	33·4	33·1	32·9	32·7	35·2	37·9	39·0	37·6	36·0	35·1	34·2	35·1
						Upper Station						
34·5	34·3	33·8	33·8	33·4	34·0	35·2	36·0	35·6	35·2	35·1	34·9	34·7
					JULY.	Lower Station						
58·1	57·0	55·6	57·7	63·3	67·8	70·2	71·1	70·2	68·2	63·7	60·6	63·7
						Upper Station						
59·2	57·7	56·5	56·8	58·3	61·2	63·9	65·1	65·5	64·4	62·2	60·8	61·0

These observations are shown graphically in Fig. 34.

Both in December and in July the average temperature at the bottom of the Tower is higher than that at the top. The maximum is also higher at the bottom than at the top, but the minimum is lower. At an altitude of 1000 feet the temperature does not rise so high nor fall so low as on the ground; the range of temperature is less. In both months the temperature at the foot of the Tower is higher than that at the top as long as the sun is up and for some time afterwards; but in the early morning, when the ground has cooled, the temperature gradient is reversed.

In December the minimum is reached at the lower station at 7.25 a.m., at the upper station at 7.55 a.m.; the maximum at the lower station occurs at 1.50 p.m., at the upper station about 2.30 p.m. In this month the sun rises about eight o'clock and sets about four. As soon as it is up the fall of temperature ceases and a rise begins. The temperature of the ground continues to increase until noon; but because the sun is low in the sky it rapidly loses its power, and in the afternoon

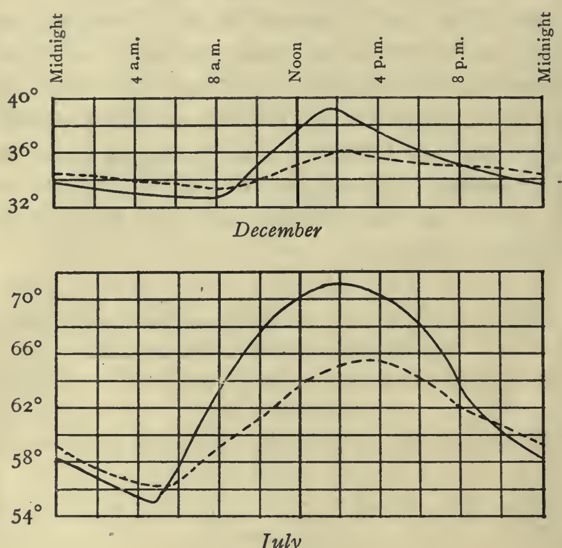


FIG. 34. Temperature changes at the Eiffel Tower.
The continuous line shows the temperature at the foot of the Tower; the broken line shows the temperature at the top of the Tower

the earth is losing more heat by radiation than it gains from the sun's rays. Therefore the ground itself is at its maximum shortly after midday. The effect of this is felt at once by the air at the lower station, which accordingly has its maximum about the same time. But the air at the upper station is warmed by conduction and convection from below, a process which takes time, and the maximum is therefore 40 minutes later.

In July the minimum at the lower station occurs at 5.0 a.m., at the upper station at 5.55 a.m. The maximum at the lower station is reached about 2.0 p.m., and at the upper station at 3.0 p.m. In this month the sun remains high in the sky far into the afternoon. It is not till two o'clock or even later that the heat received from the sun's rays becomes less than the amount lost by radiation. The maximum of the ground itself and of the air at the lower station therefore occurs about this time. At the upper station the effect is not felt until an hour later.

These observations show quite clearly that the air is warmed from below upwards, and that the cooling also, after the sun has lost its power, begins at the bottom. Hence we may conclude that the sun's rays do not themselves warm the air to any great extent, but that they warm the earth and the earth warms the air.

It must not be supposed, however, that the rays of the sun have no influence upon the temperature of the air through which they pass. Their effect, indeed, upon pure air is small, but it is greatly increased if there are suspended particles of dust or drops of water. These are warmed by the sun's rays and communicate their heat to the surrounding air. In this way the sun may warm the air within a cloud; or if the cloud be dense, its surface, heated by the sun, may warm the air above it.

Effects of expansion and compression. It is not only because the air is warmed from below that its temperature decreases upwards. There is another reason, depending upon the diminution of pressure in the same direction.

When air is compressed, without any heat being added to it, its temperature is raised. This is the chief reason why a bicycle-pump becomes warm when it is in use. No doubt the heating of the pump is due in part to the friction of the piston within the barrel; but if that were all it would be almost equally great when we move the piston backwards and forwards with the nozzle open to the air. Moreover, the heating of the pump when it is inflating a tyre is greatest near the nozzle. The friction is very little greater there, but the compression of the air is at its maximum.

When air is allowed to expand by reducing the pressure upon it, without any addition or subtraction of heat, its temperature falls.

At the earth's surface the pressure is greater than it is above. If the air is forced in any way to rise, it moves into a region of lower pressure; and it therefore expands and becomes cooler. If on the other hand the air above is forced to descend, it moves into a region of higher pressure; and it is therefore compressed and becomes hotter.

There are two principal ways in which the air may be forced to rise. It may be heated by contact with the warm earth. The heat will cause it to expand and become lighter. If it becomes lighter than the surrounding air it will rise like oil in water.

Or a wind may blow against a mountain side. In that case, even if the air is cold and heavy, it will be forced up the slope.

Descent of the air is caused most frequently by cooling. If a mass of air is cooled so much as to be heavier than the surrounding air, it will sink.

In these ways there is a constant tendency for the lower layers of the atmosphere to become mixed with one another. Any air that falls is warmed by compression, any air that rises is cooled by expansion, and thus the average temperature increases downwards, and decreases upwards.

The vertical movements of the air are due chiefly, as we have seen, to the form of the ground and to the changes of temperature of the ground, and the influence of these does not extend indefinitely upwards. Above a certain altitude their effect becomes imperceptible, vertical movements cease, and beyond that limit the temperature no longer decreases upwards.

Stable and unstable equilibrium. When a wind blows against a mountain side it is mechanically forced to ascend. When it reaches the top it may continue to rise into the air or it may sink down the other side of the mountain. Which of these courses it takes, will depend upon the vertical gradient of temperature.

The effect of the vertical gradient will be most easily understood if we take a special case as an example and, for simplicity, round numbers will be used throughout. The exact figures would be slightly, but only slightly, different from those given. Moreover, it will be assumed that the air is free from water-vapour.

If we begin with dry air at a pressure of 30 inches and a temperature of 60°, and decrease the pressure upon it to 29 inches, then without any communication of heat to or from outside its temperature falls to 55°. If the pressure is further reduced to 28 inches, the temperature will fall to 50°. If we now increase the pressure to 29 inches, the temperature will rise again to 55°, and if to 30 inches the temperature will return to 60°.

Now suppose that a hill rises from sea-level to a height

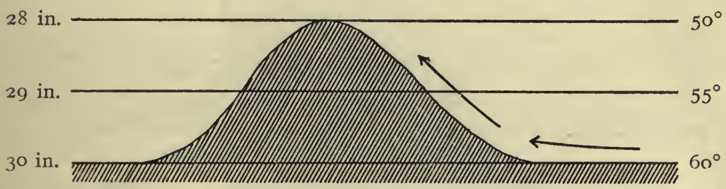


FIG. 35. Indifferent equilibrium (for dry air).

of 1800 feet, and that the pressure at the bottom is 30 inches and the temperature 60°. The pressure at a height of 900 feet will be 29 inches (using round numbers, as throughout the example) and at the top, or 1800 feet up, 28 inches. If the wind blows towards this hill, the air at the foot will be forced upwards, and when it reaches a height of 900 feet the pressure on it will be only 29 inches and its temperature will fall to 55°. When it reaches the top its pressure is reduced to 28 inches and its temperature to 50°.

These may or may not be the actual temperatures of the air at those heights. If they are, then the temperature of the air which has been forced upwards will always be the same as that of the air already at the same level and its density will therefore also be the same. If the wind ceases there will be no tendency for this air either to go on rising or to sink back. In such a case the air is said to be in a state of indifferent equilibrium. In this case the vertical gradient of temperature is 5° F. for 900 feet (Fig. 35).

But the temperature at 900 feet is not necessarily 55°. Suppose that the temperature at the foot is still 60°, at 900 feet 54°, and at 1800 feet 48° (Fig. 36). The vertical gradient is now 6° F. for 900 feet. The pressures will be scarcely altered. If the wind blows towards the hill, the air at the foot will be forced upwards. At 900 feet its pressure is reduced to 29 inches, its temperature to 55°; at 1800 feet its pressure is 28 inches, its temperature 50°. Under these circumstances it will be observed that the temperature of the risen air is higher than that of the air already at the same level, and the risen air is therefore lighter than the air surrounding it. Accordingly, even if the wind ceases, this risen air continues to rise;

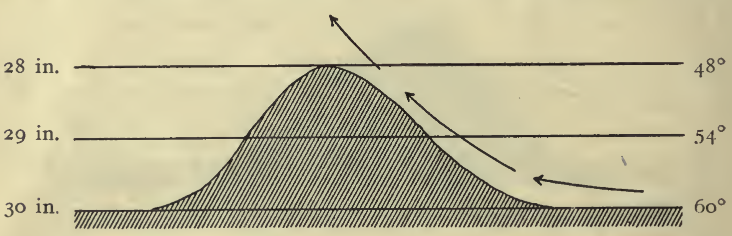


FIG. 36. Unstable equilibrium (for dry air).

and if the temperature gradient above the mountain top is the same, it will go on rising indefinitely. In this case the air is said to be in a state of unstable equilibrium. It may be left to the student to show that with this vertical gradient, if any mass of air is forced to descend, it will continue to descend even after the force has ceased to act.

Lastly, suppose that the temperature at the foot is 60°, at 900 feet 56°, and at 1800 feet 52° (Fig. 37). The vertical gradient is 4° F. per 900 feet. When the wind blows towards the hill the air at the foot is forced upwards as before. At a height of 900 feet its pressure becomes 29 inches and its temperature 55°, at a height of 1800 feet its pressure is reduced to 28 inches and its temperature to 50°. In this case the air, after it has begun to rise, is always colder and therefore heavier than the air already at the same level. Therefore if the wind ceases, it drops back again. If the wind continues to blow, it is carried to the top of the hill and falls down on the other side. The equilibrium is said to be stable. In such circumstances it may be shown in a similar way that if any mass of air is forced downwards it becomes hotter and lighter than the surrounding air, and therefore when the force ceases to act, it rises again.

In a condition of indifferent equilibrium if any mass of air is forced either upwards or downwards it assumes the temperature of the surrounding air; and when the force ceases to act, it remains in the position to which it has been moved. In unstable equilibrium if any mass of air is forced

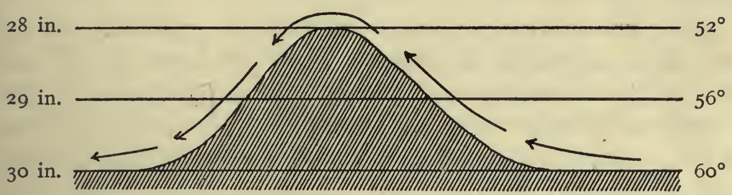


FIG. 37. Stable equilibrium (for dry air).

upwards it becomes hotter than the surrounding air, and if the force ceases, it still continues to rise; if it is forced downwards it becomes colder than the surrounding air, and when the force ceases, it continues to fall. In stable equilibrium if any mass of air is forced upwards it becomes colder than the surrounding air, and when the force no longer acts, it falls back to its original level; if it is forced downwards it becomes hotter than the surrounding air, and when the force ceases to act, it rises again to the level from which it started.

The temperature gradient for indifferent equilibrium is called the normal gradient. For unstable equilibrium the gradient is greater than the normal; for stable equilibrium it is less.

Effect of water-vapour. In the preceding example it was assumed that the air was free from water-vapour, and the gradient for indifferent equilibrium is in that case the normal gradient for *dry* air. If water-vapour be present, as it always is in fact, the conditions become more complicated. As long as the water remains in the form of vapour, the cooling caused by expansion is nearly the same as with dry air; but when the temperature falls below a certain point, condensation of the vapour begins. In condensation latent heat is liberated, and accordingly the temperature does not fall so rapidly as if the air were dry. If, for instance, we take dry air at a pressure of 30 inches and a temperature of 60° and reduce the pressure to 29 inches the temperature falls to 55°; but if the air contains so much water-vapour that condensation begins while the pressure is being reduced to 29 inches, then when that pressure is reached the temperature will be more than 55°—on account of the heat set free by the condensing vapour.

Thus the normal vertical gradient, or gradient for indifferent equilibrium, is less for moist air than for dry air; and it varies according to the amount of water-vapour present.

Vertical gradient at great altitudes. The vertical gradient near the earth's surface is far from constant and varies greatly even in the same locality, but it rarely, if ever, approaches the normal gradient for dry air.

As the height above the ground increases, the vertical gradient usually increases also, *i.e.* the temperature falls more rapidly for each 100 feet of ascent. Under ordinary conditions the gradient continues to increase up to an altitude which varies greatly, but may be put roughly at 10,000 feet. Above this it is nearly equal to the normal gradient for dry air, and it remains fairly constant to a height of about 7 miles (in England). Beyond that limit the temperature ceases to fall and is almost constant up to the greatest heights that have been reached by pilot-balloons. There are even indications of a slight increase of temperature upwards.

Thus the part of the atmosphere accessible to observation may be divided into three layers according to the vertical gradient:

(1) A lower layer, from the ground-level to about 10,000 feet, in which the vertical gradient is considerably less than the normal gradient for dry air.

(2) A middle layer, from about 10,000 feet to 7 miles, in which the vertical gradient is nearly equal to the normal gradient for dry air.

(3) An upper layer, from 7 miles upwards, in which there is either no gradient, or a very small reversed gradient.

The upper layer is commonly known as the isothermal layer. It is sometimes called the stratosphere or the advective region of the atmosphere, while the air below it is called the troposphere or the convective region.

In the lower layer there is always water-vapour present and the vertical gradient must therefore always be less than the normal gradient for dry air. At greater altitudes, the temperature is lower, and, if any vapour rises from the earth, the greater part of it is condensed and forms clouds and rain. The heavier types of cloud lie for the most part between 3000 and 10,000 feet, though rising currents may carry their summits to much greater altitudes. Above 10,000 feet the amount of water-vapour present is very small, the air is nearly dry, and the vertical gradient is nearly the normal gradient for dry air.

We have already seen that wherever there are vertical movements of the air, the temperature must decrease upwards; and therefore the fact that there is no decrease in the isothermal layer shows that vertical movements cease where the isothermal layer begins. The lower part of the atmosphere is disturbed by the differences of temperature on the surface of the earth. Ascending and descending currents are produced, but apparently the influence of the surface is not felt beyond a certain altitude. Just as the disturbance which produces land and sea breezes is very shallow, so we may expect that even the largest disturbances will have an upward limit.

The height at which the isothermal layer begins is not everywhere the same and even at the same place it is not constant. But it is only within the last few years that it has been systematically investigated and the number of observations is not yet sufficient to define its limits. It is clear however that towards the equator its altitude is greater than in more northern latitudes. In Europe it begins at a height of about 7 miles, near the equator at a height of 10 or 11 miles.

The base of the isothermal layer is sometimes very sharply defined, and the change of gradient is sudden. But at other times the high gradient of the air below diminishes more gradually and there is no definite boundary to the isothermal layer. We may conclude that although the isothermal layer is little affected by the disturbances below, yet at times their influence extends into its base.

Temperature on mountains. Over a plain, when the sun is shining, the temperature decreases upwards, because, as we have seen, the air receives most of its warmth from the earth. If the air is still, the isothermal surfaces[1] will be horizontal,

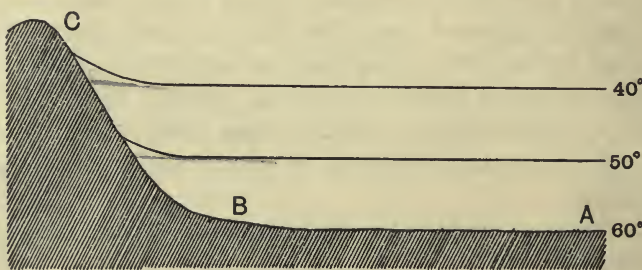


FIG. 38. Isothermal surfaces against a mountain-side heated by the sun.

for otherwise there will be hotter and lighter air lying amidst heavier and denser air. The distribution of temperature will be similar to that shown in the right-hand part of Fig. 38, where *AB* represents the surface of the plain, with a temperature of 60°.

But if a mountain, *BC*, rises from the plain, the mountain is warmed by the sun in the same way as the plain and becomes a source of heat to the air into which it penetrates. The air on the mountain side is warmed by contact with the mountain

[1] "Isothermal surfaces" must not be confounded with the "isothermal layer." The latter is a particular layer of the atmosphere, of unknown thickness, in which the temperature is known to be nearly uniform. An isothermal surface is any surface (in the geometrical sense of the word) on which the temperature is everywhere the same; it has no thickness.

and becomes hotter than the air at the same level above the plain. The isothermal surfaces therefore rise towards the mountain, as shown in the diagram.

Since the air in contact with the mountain is warmer than the air at the same level outside, it is also lighter. Therefore it rises and the cooler air from outside flows towards the mountain. In this way, even though the mountain itself may be as hot as the plain, the air on the side and top of the mountain is kept cooler than the air that rests on the plain.

At night if radiation is unchecked, both the plain and the mountain grow cold and cool the air in contact with them. The isotherms will now bend down towards the mountain as shown in Fig. 39. The air on the mountain-side becomes

FIG. 39. Isothermal surfaces against a mountain-side cooled by radiation.

colder and heavier than the free air at the same level above the plain. Accordingly it sinks downwards and the warmer air from outside flows towards the mountain.

Mountain and valley winds. It follows from this that, in still weather, when the sun is strong there will be a flow of air towards the mountain and up its slope; at night, if the sky is clear, there will be a flow of cold air down the mountain-side. The flow is naturally concentrated into the valleys running up into the mountain; and accordingly in many mountain districts a wind blows up the valleys by day and down them by night. Such winds are often felt in the Alps, and in the Himalayas they are frequently very strong. In cloudy climates they are scarcely noticeable; and, like land and sea breezes, they may be overpowered by winds due to other causes.

Inversions of temperature. In winter the nights are long, and, if the sky be clear and the air still, the conditions illustrated in Fig. 39 may be maintained for hours. Gradually the cold air collects in the hollows and valleys, and a very marked reversal of the temperature gradient, or "inversion of temperature," is brought about.

A pool of cold air lies in the valley, on the mountain-side above the pool the air is comparatively warm, and higher still the ordinary decrease of temperature with increase of altitude begins. Such inversions of temperature are common in the Alps and even amongst our own mountains they occur occasionally, when the nights are calm and clear.

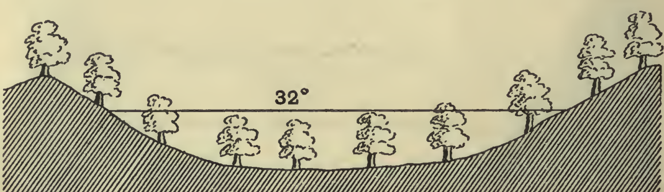


FIG. 40. Inversion of temperature.

Under such circumstances it may freeze in the valley while higher up there is no frost. An interesting example of this phenomenon has been recorded in the English Midlands. It happened towards the end of May in 1894, after the ash-trees were in leaf. The air was still and the sky was clear. At night cold currents flowed down the slopes in the manner described and the hollows were half filled with air below the freezing-point, while the warmer and lighter air floated quietly upon the denser air beneath. The trees on the floors of the valleys were frost-bitten to their tops. On the slopes the leaves of the lower branches were affected, but the upper branches were untouched. The trees on the higher ground escaped the frost altogether (Fig. 40).

The Föhn. Besides the mountain and valley winds already described there is a wind of a very different type which blows from the mountains to the plains. In Switzerland it is called the Föhn, on the eastern side of the Rocky Mountains it is known as the Chinook, and similar winds with various local names are met with in other parts of the world. The special characteristic of these winds is their dryness and their high temperature.

If the barometric pressure on one side of a mountain chain is higher than on the other a wind will blow towards the chain from the higher pressure, and if the atmosphere is in a state of stable equilibrium it will descend the other side.

The wind will always contain a certain amount of water-vapour and if the mountain chain is high enough, some of the vapour will be condensed. Accordingly the decrease of temperature as it ascends the mountain is less than if the air were dry, on account of the liberation of latent heat. On the other side as the air descends the pressure upon it increases and its temperature rises. The rise of temperature is the same as for dry air and the gain of temperature in descending is therefore greater than the loss in ascending.

An example similar to those already given will serve to make the matter clear. Suppose that the chain is 1800 feet high and that the wind starts towards it with a temperature of 60°. We have already seen that if it were dry air, its temperature when it reached the top would have fallen to 50°, owing to expansion. But if there is any condensation of vapour, the fall is less, and the temperature actually reached will be higher, say, for example 51°. In descending the other side the temperature increases. There is no further condensation, and the increase is practically the same as for dry air. When it reaches the level from which it started, its temperature will have risen 10°, and will be about 61°. It is hotter as well as drier than when it started.

If the air is saturated to begin with, so that condensation goes on all the way up the windward slope, the temperature at the top will be much higher than in the example given and the difference of temperature on the two sides will be far greater. In such a case the increase of temperature in descending the leeward slope is approximately twice the decrease in ascending the windward slope.

Thus the Föhn is a very hot and a very dry wind, and if it comes in spring, it rapidly melts the snow in the valleys and on the slopes of the mountains.

Temperature of plateaux. A plateau differs from a mountain or a mountain chain in the fact that it is much broader. It is in fact a very much larger mass projecting into the atmosphere and it produces a greater effect upon the temperature of the air.

In Fig. 41 *AB* represents the surface of a plain and *CD* the surface of a plateau rising above it. If both of these are equally exposed to the rays of the sun, the plateau will become as hot as the plain—possibly hotter, because the rays have

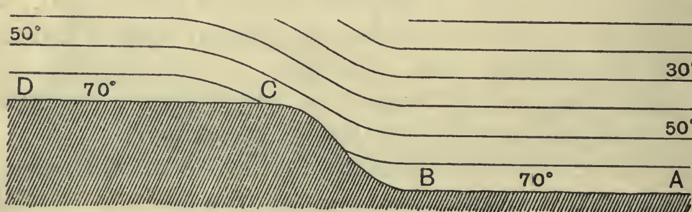


FIG. 41. Isothermal surfaces over plateau and plain.

a smaller thickness of the atmosphere to penetrate. Supposing that each of them is heated to a temperature of 70°, then each will tend to produce a vertical distribution of temperature such as is shown on the right-hand and the left-hand sides of the diagram respectively. On the slope *BC* the isothermal surfaces will bend upwards, as on the sides of a mountain in similar circumstances.

With such a distribution the air cannot be at rest, for the air at 60° above the plateau is on a level with the air at 30° above the plain; and in general the air above the plateau is hotter and lighter than the air at the same level above the plain. The effect is much the same as if the two were covered with liquids of different density: the plain, for example, with water, the plateau with oil. The denser fluid flows beneath the lighter and the lighter fluid flows over the denser.

Accordingly the cooler air from outside flows under the warmer air of the plateau, displacing it and forcing it to rise and at some higher level to overflow.

In consequence of the influx of cold air, the air on the plateau is kept cooler than it would otherwise be, especially near the edge. But the inflowing air is continually being warmed by the heat of the plateau itself, and the temperature still remains far higher than that of the air at the same level above the plains.

The general form of the isothermal surfaces will therefore be as shown in Fig. 41. But it is probable that at higher altitudes the bend above the edge of the plateau will gradually diminish.

In this diagram the plateau is comparatively low. The air on its surface has a temperature of 70°, the air at the same level above the plain has a temperature of 40°. The difference of density is the same as if both rested on a plain one part of which was heated to 70° and the other part to 40°. If the plateau is high the difference will be far greater than can ever be caused by unequal heating of a plain; and accordingly the winds that flow inwards to a heated plateau are much stronger than those that flow towards a heated plain.

It is because the central part of Asia is an elevated table-land that the Asiatic monsoons are so strongly marked. If it were a low lying plain there would still be monsoons but they would be comparatively weak.

CHAPTER VII

HUMIDITY OF THE ATMOSPHERE

Water-vapour in the air. Wherever air and water are in contact, there is always an interchange of particles between them. Particles of air pass into the water, particles of water pass into the air in the form of vapour Therefore there is always a certain amount of water-vapour present in the air and a certain amount of air present in the water. Some of the air in the water passes back again into the air and some of the vapour in the air passes back into the water.

If the amount of vapour present in the air is small, more particles of water will pass from the water into the air than from the air into the water, and the water gradually "dries up" or "evaporates." But if the amount of vapour already in the air is large, then as many particles of vapour may pass from the air into the water as from the water into the air, and the water will not evaporate. In such a case the air is said to be "saturated" or in ordinary language "to hold as much vapour as it can contain."

Even from the surface of snow or ice evaporation may take place. Sometimes during a long frost the snow that has previously fallen, gradually disappears without melting.

The vapour in the air above the oceans is carried by the winds into the heart of the continents, and even in the driest parts of the globe the air is never absolutely dry. But the amount of vapour present varies greatly according to the place and the season.

It is a matter of common observation that in general evaporation is most rapid at a high temperature. Therefore we may expect to find the greatest amount of vapour over the oceans near the equator and the smallest amount over the land in a cold region such as north-eastern Asia in winter.

The amount of water that can exist as vapour in a cubic foot of air has a definite limit, depending solely upon the temperature. When the limit is reached the air is said to be saturated, and if more vapour of water is forced into the space, the excess above the limit will condense into the liquid form. It is important to notice that the pressure of the air has no effect ; more dry air could be pumped into the same space without producing any condensation.

The following table gives the number of grains of water in a cubic foot of saturated air at various temperatures :

Temp. Fahr.	30°	32°	40°	50°	60°	70°
Grains per cu. ft.	2·21	2·37	3·09	4·28	5·87	8·00

At 50° a cubic foot of saturated air contains 4·28 grains; it can take up no more. If its temperature be raised to 60° it will be able to contain 5·87 grains, that is, it will be able to take up 1·59 grains more than it actually holds. If on the other hand its temperature falls to 40°, it can only hold 3·09 grains, and therefore 1·19 grains will condense as water. For these figures to be exactly correct the air must be confined in a vessel, so that the change of temperature has no effect upon its volume.

Evidently therefore the feeling of wetness or dryness in the air does not depend only on the actual amount of vapour present. In the case described the same air, with the same amount of water in it, would feel dry at 60° and wet at 40°. At 60° a damp cloth hung in it would become drier, at 40° it would become wetter.

So far as the sensations or the formation of cloud and rain are concerned, it is not the actual amount of vapour present which is of importance, but the proportion of the actual amount to the maximum amount possible at the particular temperature. The air of the Sahara holds more vapour per cubic foot than the air of England on a wet day in winter, but because of the high temperature it is able to take up still more.

Absolute and relative humidity. In any particular case the actual amount of vapour in a cubic foot of air (or a litre, if metric measures are being used) is called the absolute humidity of the air in question; the proportion of this amount to the amount possible at the actual temperature of the air is called its relative humidity. If for example the temperature of the air is 50° and the amount of vapour is 2·14 grains per cubic foot, then the relative humidity is $\frac{2 \cdot 14}{4 \cdot 28}$ or 50%, 4·28 grains per cubic foot being the possible amount at this temperature.

Absolute humidity is often expressed in terms of the pressure exerted by the aqueous vapour present. The relative humidity is then the proportion of the actual vapour pressure to the vapour pressure in saturated air of the same temperature. This gives practically the same result for the relative humidity as when the weights of vapour per cubic foot are compared.

Dew-point. The quantity of water-vapour in the air can be determined chemically by passing a known volume of air through tubes containing calcium chloride or some other drying reagent, and noting the increase of weight. But the process is rather long, and some quicker method must be employed if daily observations are to be made. One of the simplest is the determination of the dew-point.

If the air is gradually cooled, it will at length reach the temperature at which it is saturated, and if it is cooled any further, condensation will take place. The temperature at which condensation begins, that is, the temperature of saturation, is called the dew-point.

Having determined the dew-point, reference to a table such as that on page 84, but more complete, will give at once the amount of vapour present, or the absolute humidity. If, for example, the dew-point is 40°, then the air which has been tested contains 3·09 grains per cubic foot. It should perhaps be mentioned that in the original preparation of these tables, the amount of vapour present in saturated air at different temperatures was determined chemically.

The instruments employed for determining the dew-point are known as hygrometers. Various forms are described in elementary text-books of physics and as they are seldom used in meteorological observations, nothing further need be said about them here.

Wet and dry bulb thermometers. The humidity of the air is usually determined by the use of a pair of thermometers, the bulb of one of which is kept continually moist. Such a pair of thermometers is sometimes called a psychrometer.

Both thermometers are mounted on a stand, with their bulbs projecting beyond the mount, so as to be freely exposed to the air. One of the bulbs is left in this condition. Round the other muslin or cotton is loosely wrapped, wetted with water and kept constantly moist by allowing a few threads to dip into a little cup of water. The whole instrument should be placed inside a screen such as that described on page 47.

The thermometer with the dry bulb gives the temperature of the air. If the air is completely saturated with vapour the reading of the wet bulb thermometer will be the same. But if the air is not saturated there will be evaporation from the moist cotton round the bulb; this evaporation will cool the bulb and the reading will be lower than that of the dry bulb thermometer. The more rapid the evaporation the greater the cooling will be; and consequently the drier the air, the greater is the difference between the two thermometers.

The relation between this difference and the relative humidity is somewhat complex. But tables have been constructed which give the relative humidity at once when the readings of the two thermometers are known.

It is, however, clear that the rate of evaporation from the cotton around the wet bulb will depend not only upon the relative humidity but also on the rapidity of the currents of air past the thermometer. If, for example, we blow the wet bulb with a pair of bellows, the mercury in the thermometer will fall.

Consequently this method of determining humidity is not entirely satisfactory, for even in a Stevenson screen the air currents will not always be the same.

A form of psychrometer has, however, been devised in which this difficulty is avoided. It is known as Assmann's psychrometer. The bulbs and the lower part of the stems of both thermometers are enclosed in wide metal tubes, which are open at the bottom. At the top the two tubes unite into a single tube leading into a kind of reservoir which has an outlet to the air and in which there is a small ventilating fan operated by clockwork. In taking an observation the clockwork is wound and set going. It works the fan at the same rate and for the same length of time at every observation, and thus it always draws the same amount of air at the same rapidity past the bulbs of the thermometers. All observations are accordingly made under the same conditions so far as the movement of the air is concerned, and the results are much more reliable than in the ordinary form of psychrometer.

Condensation. When a mass of air is cooled below the dew-point, some of the vapour, as we have seen, condenses into the liquid or the solid form. In nature this condensation may take place in several different ways. The water produced may collect on solid objects as dew or hoar-frost ; it may hang in the air in little drops, forming mist or fog or cloud ; or it may fall as rain, snow or hail.

Dew and hoar-frost. Dew and hoar-frost are the effect of condensation not in the air itself but upon the surface of solid bodies exposed to the air. If the temperature of the air is higher than the dew-point, and a cold object at a low enough temperature is brought into the air, then the cold object cools the air in contact with it below the dew-point and causes condensation, but there is no condensation in the rest of the air. The water formed settles upon the cold object and is known as dew. If the dew-point is below the freezing point, the vapour will condense as spicules of ice and not as drops of water, forming hoar-frost instead of dew.

For the natural formation of dew there must be water-vapour in the air and the temperature of the grass or of other objects on the ground must fall below the dew-point. The greater the amount of vapour present and the lower the temperature of the ground the more abundant will be the dew.

It is essential that the air should be still or nearly so, for if the wind is strong there is too much mixing of the air and none of it remains in contact with the ground long enough to be cooled to the dew-point—unless indeed the whole mass of air is already near that temperature.

A warm day favours the production of dew during the following night ; for a warm day assists evaporation and therefore tends to increase the amount of vapour present. At night the ground must be cold, but it must not owe its coolness to a cold wind, for it is indispensable that the ground or grass should be colder than the air. It must be cooled by radiation. Consequently the conditions most favourable to the formation of dew are calm weather and a clear sky, allowing the ready access of the sun's rays by day, and free radiation by night.

These are the conditions characteristic of anticyclones and of the fronts of wedges; and it is accordingly, when these types of pressure distribution prevail, that dew, and also hoar-frost, most frequently occur.

It was Dr Wells, in the year 1818, who first gave a satisfactory explanation of the formation of dew; and since his time until about 30 years ago, it was believed that the vapour which condenses into dew was already in the air at sundown. But in 1885 Dr John Aitken, a Scottish meteorologist, published a number of observations which show that this view is not quite correct.

He cut a turf out of his lawn and fitted it into an iron tray so that it could be removed or replaced at will. On an evening when dew was likely to form, the turf was taken out and weighed and then put back into its place in the lawn. After a heavy dew had been deposited and the turf was wet, it was again removed and weighed. In spite of the dew upon it, it weighed less than before, showing that the turf had lost more water to the air than it had gained by condensation.

He also made shallow boxes or trays of tin-plate, about three inches deep; and rested them upside down upon the ground, so that the bottom was about three inches above the surface. Dew was deposited profusely upon the inside of the trays, while on the outside there was little or none. Similarly it may be noticed that when a stone lies loose upon the ground there is often a far more copious deposit of dew upon its lower than upon its upper surface.

These experiments seem to prove that the vapour which forms the dew rises from the ground. In such a climate as ours the earth is always more or less moist, and this moisture is continually evaporating and passing upwards into the air. Grass and other plants assist the process considerably. They draw up the moisture through their roots and transpire it through their leaves. The warmer the earth is, the more rapid will be the stream of vapour from the earth into the air. As long as the air is warm and unsaturated and the surface of the ground is also warm, the vapour will still remain as vapour in the air. But loss of heat by radiation at night cools the blades of grass and also the actual surface of the ground (while the earth beneath still remains warm), and the vapour condenses upon them as dew.

When the earth beneath the surface is cold, evaporation takes place more slowly and the vapour rises much less freely. There is, accordingly, less vapour in the air immediately above the ground, and the grass must cool to a much lower temperature before the dew-point is reached. In such circumstances the dew-point is likely to be lower than 32° F., and hoar-frost will be formed instead of dew.

From what has been already said, it will be evident that dew is most likely to be abundant when the earth beneath the surface is warm, but the surface itself is cold. This is the reason for the heavy dews of early autumn. The earth is still hot from the summer's sun, but the nights are long enough for the grass and the surface of the ground to become very cold by radiation, if the sky is clear. In spring the earth for some distance below the surface is still cold after the winter's frosts, and gives off vapour much more slowly.

Mist or fog. In the formation of dew the air is cooled where it is in contact with something colder than itself, the rest of the air remaining comparatively warm ; and condensation takes place upon the surface of the cold object. But if the air is cooled throughout, condensation takes place within the air itself. Small drops of water are formed, which may float in the air for a considerable time, and a fog or mist is produced or, higher in the atmosphere, a cloud.

It might be expected that the drops of water would fall as soon as they are formed. They actually do fall, except when an upward current of air prevents them, but they are very small and owing to the resistance of the air their fall is very slow. The smaller the size of the drop the greater, compared with its weight, is the resistance of the air.

If the drops are spherical and we denote the radius by r, the weight of the drop is proportional to the volume, *i.e.* to $\frac{4}{3}\pi r^3$, and the resistance of the air is proportional to the area of the greatest cross-section, *i.e.* to πr^2. The weight increases with the cube of the radius, the resistance increases with the square of the radius. Therefore the larger the drop the greater is the weight compared with the resistance, and the smaller the drop the smaller is the weight compared with the resistance. A large drop therefore falls rapidly, a small drop slowly.

There are several ways in which air near the ground may be cooled throughout its mass.

It may be cooled by radiation. It is not so much the air itself that is cooled in this way as the dust-particles within the air. But the dust-particles cool the air in contact with them. Each particle becomes the nucleus of a drop of water, and the effect is the same as if the whole mass of air had been cooled. It is not often, however, that a fog is formed by radiation alone.

Air may be cooled by mixing it with colder air, but the amount of condensation brought about in this way cannot be great, for while the warm air is cooled, the cold air is warmed and is therefore able to hold more vapour.

From the table on page 84, it appears that:

1 cu. ft. of saturated air at 60° contains 5·87 grains of water-vapour.

1 cu. ft. of saturated air at 40° contains 3·09 grains of water-vapour.

The result of mixing, if there were no condensation, would be practically

2 cu. ft. of air at 50° containing 8·96 grains of water-vapour.

But on looking at the table it will be seen that 2 cu. ft. of air at 50° cannot contain more than 8·56 grains. The extra 0·40 grain will be condensed. The figures are approximate only, because in the process of condensation latent heat will be given out, and the final temperature of the mixture will be a little more than 50° and its volume a little more than 2 cu. ft. But the calculation shows that when two masses of saturated air at different temperatures are mixed, there is condensation of vapour, but the amount condensed is small.

It appears therefore that neither radiation nor mixing is likely to produce a really heavy fog.

There is another way in which a fog or mist may be formed. Instead of cooling the air we may add to it more water-vapour, and if we add more than enough to saturate the air, condensation of the surplus vapour will take place. This is what happens above a hot bath in a cold room. More vapour passes from the water to the air than the air can contain, and some of it condenses into little drops of water and becomes visible to us as steam.

The conditions are precisely similar when the air is cold and the earth is warm and moist or, at sea, the water is warm. The vapour rising from the warm earth or sea, like that from the hot bath, is partially condensed and forms a fog.

The ground-fogs of autumn are due, in part at least, to this cause. The earth is warm and is giving off vapour, but the actual surface may be cooled by radiation. The layer of air resting upon the ground is cooled, partly by contact with the surface, partly by the dust-particles scattered through it, which have also lost heat by radiation. Thus the surface is covered by a thin layer of cold air, which will collect especially upon the lower-lying ground. It will be unable to hold all the vapour that was already in it and that still continues to rise into it from the warmer earth beneath the surface, and condensation will take place upon the particles of dust within the air.

A fog of this kind is formed in much the same way as dew, and in fact it is generally accompanied by a heavy fall of dew. But in order to form fog as well as dew, not only must the tips of the grass blades grow cold, but the air above must also be cooled. The thickness of the fog is usually not more than a foot or two, excepting in the hollows, because only the lowest layers of air are, as a rule, sufficiently cooled.

When cold air drifts slowly and quietly into a region where the earth or sea is warm, a fog will be formed in a similar way, and as the thickness of the air-drift may be considerable the fog may extend to a height of some hundreds of feet. But there must be no strong wind, or the vapour is carried away.

From late autumn to spring a gentle drift of cold air will occur in England if we are situated on the southern or eastern margin of an anticyclone. It may produce fog at any time, but it is most likely to do so in late autumn or early winter, before the earth beneath the surface has experienced the full effect of the winter's cold.

The fogs of the Atlantic off Newfoundland are due to a conjunction of causes. The cold Labrador current and the warm Gulf Stream are here in close proximity. At certain seasons ice-bergs from the Arctic Seas are brought down in large numbers by the former and melt in the warmer waters of the latter. All the conditions favouring the production of fog are present. If the warm air over the Gulf Stream mixes with the cold air over the Labrador current there is likely to be condensation. If the cold air from the Labrador current flows over the Gulf Stream it will be unable to hold the large amount of vapour given off by the warm waters of the latter. The ice-bergs, too, assist in the formation of fog, by cooling the air in their immediate neighbourhood; and because they are often of considerable height, their effect is not confined to the lowest layer of the atmosphere.

It is said that during the prevalence of the fogs the air is usually warmer than the sea beneath. If this is really the case it would appear that the mixing of the warm and cold air and the cooling influence of the ice-bergs are the principal causes of the fogs.

Effect of dust-particles. It has already been explained that heating and cooling of the air are sometimes brought about by heating and cooling of the dust-particles within the air. They are solid, and are readily warmed by the rays of the sun, and easily lose heat by radiation at night, like the ground itself. But they also play another part in the condensation of vapour.

It was shown by Aitken that if the air in a flask is suddenly cooled throughout its mass the mode in which condensation takes place depends upon whether the air is dusty or free from dust. The sudden cooling is brought about by reducing the pressure in the flask by means of an air pump and thus causing expansion. If the air is full of particles of dust, condensation takes place upon those particles and little drops of water are formed throughout the air, and the flask is filled with fog. If the number of particles is less, fewer drops are formed, but they are larger. If the air is entirely free from dust, condensation takes place only on the sides of the flask. For some time it was believed that dust-particles were absolutely essential to the formation of mist, cloud or rain.

It has since been proved, however, by Mr. C. T. R. Wilson that under certain conditions condensation can take place upon the ions of the air itself; but the temperature must be lowered considerably below the normal saturation point before such condensation begins.

Clouds. In mountainous districts clouds often cover the tops of the mountains, and if we climb up till we reach the clouds we find that they are precisely similar to the mists of the lower ground. A cloud in fact is a mist formed high up in the atmosphere.

Like a mist it can only be produced by the cooling of the air *en masse*, and this cooling may be brought about in several ways.

Stratiform clouds. (Pls. I and II.) Possibly it may be due in some cases to loss of heat by radiation, and especially by radiation from the dust-particles in the air. This is not likely to be an effective method except when the air is still, and it appears probable that such a cloud would be in the form of a layer, for as it grows the air beneath will be more and more protected from further loss of heat.

A cloud may also be formed by the mixing of warm and cold air. There are often currents of air flowing in various directions at different heights in the atmosphere, and at the surface of contact between two currents there will be a certain amount of mixing. If the temperatures are not the same there may be condensation. A cloud formed in this way will usually be a thin sheet, with a current in one direction above it and a current in a different direction below it. The surface of contact may be thrown into waves, like the surface of

A. W. Clayden, photo.

Strato-cumulus.

a river when a wind is blowing. Owing to the low density of air the waves will probably be much higher than water-waves, and, on account of the lower temperature and pressure, condensation will be greatest at their crests. The cloud itself will therefore be influenced by the waves and may appear as long bars stretching across the sky. Or there may be two sets of waves crossing each other, and the sheet will be broken into rounded patches with rivulets of blue between.

Clouds like these, which are of small thickness but of wide extent, may be grouped together as stratiform.

Cumulus clouds. (Pl. III.) Neither radiation nor mixture can, however, produce the great masses of cloud which are often seen in an English sky. These are due for the most part to rising currents of air. As the air ascends the pressure upon it is reduced and it expands; and accordingly throughout its whole mass the temperature falls. At a certain height the dew-point is reached and condensation begins. Clouds formed in this way therefore have a nearly level base. But if the air continues to rise, condensation will still go on and the cloud may tower upwards to an altitude of many thousand feet. Usually it seems to be made of a number of globular masses merged together, and presents some resemblance to the steam from a locomotive. It is in fact an ascending column of vapour, the upper part of which is rendered visible by condensation.

Such clouds as these are called cumulus clouds. Their apparent shape will vary according to their position in the sky. When they are near the horizon both their flat bases and their rounded tops are visible. When they are near the zenith they will be seen from below; their tops will be hidden, and only their bases will be visible, and the outline may be quite irregular.

The strength of the winds usually increases upwards and therefore the top of a cumulus cloud often leans to one side or another, generally forwards, *i.e.* in the direction in which the cloud is travelling. If the upper currents are very strong it may be drawn out like the trail of steam from an express train at full speed.

If the top of the cumulus reaches so high that the temperature of the rising air is sufficiently reduced, spicules of ice may be formed instead of drops of water. The top loses its rounded outline and hair-like streamers spring out from it. When this happens rain or hail generally falls from the base of the cloud.

At other times the rising current to which the cloud is due seems suddenly to cease, or perhaps it meets a horizontal current blowing across it. The top of the cloud spreads outwards in a horizontal sheet, and while the lower part may be a typical cumulus the top is stratiform. Occasionally the rising current does not entirely cease where the outward spreading begins, and the top of the cumulus breaks through the horizontal sheet.

Owing to the effects of perspective a stratiform cloud which is at a comparatively low level, may seem to cross or rest upon the top of a more distant cumulus cloud, and the two together may present a deceptive resemblance to the form above described. But if they are watched for some time their relative positions will usually alter and it will become obvious that the two clouds are distinct from one another.

Cirrus clouds. (Pl. IV.) On account of the upward decrease of temperature the greater part of the water-vapour is condensed at comparatively low altitudes, and above 7000 feet the amount of vapour present is usually small, except where the rising current of a cumulus cloud carries the vapour to greater heights. But even at an altitude of six or seven miles the air is not absolutely dry.

If from any cause the small amount of vapour present at great altitudes is condensed, the cloud which is produced will not consist of water but of little crystals of ice. It will be thin, and the light of the sun or even of the moon will pass through it with very little loss.

Clouds of this kind are known as cirrus clouds. By day they are almost uniformly white, never showing more than the faintest trace of shade; but at sunset they may be brilliantly coloured. Often they consist of a small white clot with streamers hanging downwards. Sometimes the streamers

Plate III

A. W. Clayden, photo.

Cumulus.

(On the edge of a storm.)

are drawn out into long horizontal threads, owing to strong currents in the upper air: and not uncommonly the cloud takes the form of a feather with the web stripped from one side. Immediately in front of a cyclone the cirrus clouds become so abundant as to form a thin continuous sheet, giving the sky a somewhat milky look and causing halos round the sun and moon. A sheet of cirrus of this kind is known as cirro-stratus.

International nomenclature of clouds. The variety of form amongst clouds is so great that the names given above are not sufficient for descriptive purposes. There are, for instance, many kinds of cirrus differing greatly from one another in their general appearance, and there are also many clouds which are intermediate in character between the three principal types. Terms such as cirro-cumulus, cumulo-stratus, etc. have been proposed for these varieties; but unfortunately different writers have used these terms in different senses.

In order to introduce uniformity into the nomenclature the International Meteorological Committee held at Upsala in 1894 decided upon a classification of clouds for international use, and an "International Cloud Atlas" was issued with figures of the principal varieties. Some improvements were suggested at a Conference at Innsbruck in 1905 and these have been incorporated in the second edition of the Atlas.

In this international scheme the following types of cloud are distinguished:

1. *Cirrus.* Detached clouds of delicate and fibrous appearance, often showing a feather-like structure, generally of a whitish colour.

2. *Cirro-stratus.* A thin, whitish sheet of clouds, sometimes covering the sky completely and giving it only a milky appearance (it is then called cirro-nebula), at other times presenting, more or less distinctly, a formation like a tangled web.

3. *Cirro-cumulus* (mackerel sky). Small globular masses or white flakes without shadows, or showing very slight shadows, arranged in groups and often in lines.

4. *Alto-stratus.* A thick sheet of a grey or bluish colour.

5. *Alto-cumulus.* Largish globular masses, white or greyish, partially shaded, arranged in groups or lines, and often so packed that their edges appear confused.

6. *Strato-cumulus.* Large globular masses or rolls of dark clouds often covering the whole sky, especially in winter.

7. *Nimbus* (rain cloud). A thick layer of dark clouds, without shape and with ragged edges, from which steady rain or snow usually falls.

8. *Cumulus*. Thick clouds of which the upper surface is dome-shaped and exhibits protuberances, while the base is horizontal.

9. *Cumulo-nimbus* (thunder-cloud, shower-cloud). Heavy masses of cloud rising in the form of mountains, turrets or anvils, generally surmounted by a sheet or screen of fibrous appearance (false cirrus) and having at its base a mass of cloud similar to nimbus.

10. *Stratus*. A uniform layer of cloud resembling a fog but not resting on the ground.

CHAPTER VIII

PRECIPITATION

If the condensation in a cloud goes sufficiently far, the little drops of water may coalesce into larger ones, which fall as rain. If the dew-point is low enough, the vapour condenses into little crystals of ice instead of drops of water, and these may unite and form flakes of snow. An ordinary flake of snow, such as we usually see in England, consists of a large number of small ice-crystals clotted together, and such flakes are most readily formed when the temperature is not too low and the crystals of ice are more or less wet. In very cold weather the ice-crystals usually remain separate.

Another form in which the condensed vapour may fall is hail. The way in which hail-stones are produced is not clearly understood. They are not simply drops of water which have been frozen as they fall, for they frequently show a remarkable concentric and radial structure. Moreover, they may reach a size far larger than is ever attained by drops of rain. In some way or other hail appears to be connected with electrical disturbances, for a hailstorm is often also a thunderstorm.

Rain-gauge. The total amount of water that falls on any given area, whether in the form of rain or snow or hail, is known as the "precipitation," or more commonly (though not quite correctly) as the "rainfall." It is measured by means

Plate IV

A. W. Clayden, photo.

Cirrus.

of a rain-gauge, which in principle is simply a funnel leading into a vessel of some kind to contain the rain that falls upon the funnel. The area of the opening of the funnel is known and by measuring the amount of water collected in the vessel beneath, we know how much has fallen on that area.

There are, however, several precautions which must be taken in order to ensure an accurate result. If the gauge is placed on the ground, a certain amount of water splashes upwards or may be carried upwards by swirls of air, and may enter the funnel. Consequently a rain-gauge on the ground always collects more water than a similar gauge set at a height of three or four feet. The standard height adopted in England is one foot.

Some of the water that falls on the inside of the funnel splashes outwards and is lost. To prevent this loss as far as possible, the sides of the funnel should be vertical for two or three inches at the top.

When the precipitation takes the form of snow, the funnel may be choked and filled before the storm has ceased. To lessen this danger the upper vertical-sided part of the funnel should be made fairly deep. In order to melt the snow a measured quantity of warm water may be poured into the funnel when the observation is being made, and this quantity must be subtracted from the total in the collecting vessel.

In English-speaking countries the amount of rainfall is usually expressed in inches[1]. It may be said, for example, that at a certain place an inch of rain has fallen during the day. The meaning is that if all the rain that fell during that day had stayed where it fell, it would have covered the ground to a depth of an inch.

A rain-gauge is therefore usually provided with a measuring-glass specially graduated to suit the funnel. If the area of the opening of the funnel is 16 square inches, it needs 16 cubic inches to cover that area to a depth of one inch. Accordingly the measuring-glass will be graduated so that a volume of 16 cubic inches is marked as 1 inch. The graduations are

[1] The Meteorological Office has recently adopted the measurement of rainfall in millimetres (see Appendix).

not cubic inches but depths on an area equal to that of the funnel opening.

In the plains of England a rainfall of an inch in a day is exceptional, in the more mountainous parts it is not uncommon. But even in the plains much heavier falls are occasionally experienced. On August 26, 1912, more than 8 inches fell in 24 hours near Norwich.

In the east of England the average rainfall for the whole year amounts to about 25 inches; in the plains of the west it may be as much as 30 or 40; in the mountains it is very much more. At Sty Head Pass in Cumberland the average annual rainfall is 169 inches, and for long this ranked as the wettest place in the British Isles from which records were obtained. But in recent years more rain-gauges have been set up amongst the mountains and greater rainfalls have been observed. In 1908 a rainfall of 237·32 inches was recorded at Llyn Llydaw on the flanks of Snowdon.

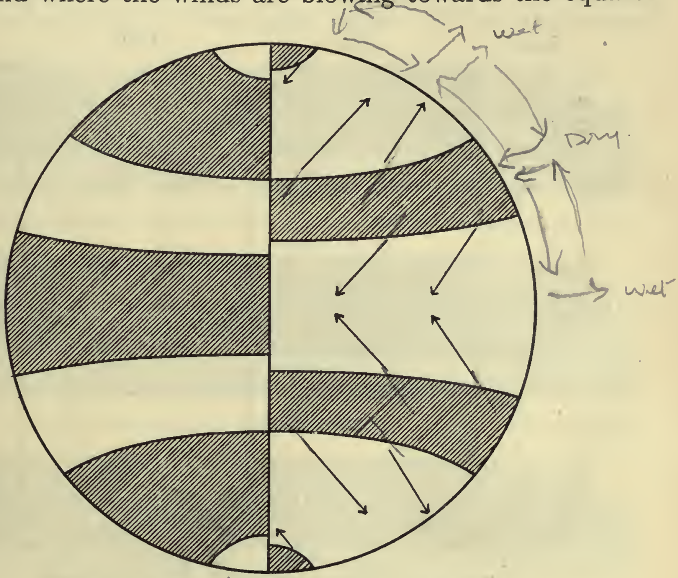
Outside the British Isles much greater extremes are met with. At Aden there may be no rain at all for two or three years. At Cherrapunji in Assam the annual rainfall is about 460 inches.

General distribution of rainfall. Rain is due to the cooling, *en masse*, of air which contains water-vapour. Apart from local and temporary conditions there are two principal methods by which this cooling may be brought about. The air may flow to colder latitudes or it may rise to greater altitudes where both temperature and pressure are less. In either case there will be a tendency to produce rain.

Where, on the other hand, the air is flowing to warmer latitudes or where it is descending, it will in general be able to take up more vapour and, in the ordinary sense of the word, it will be dry.

In the diagram of the general distribution of pressure on page 37 it will be seen that the earth is surrounded by a series of belts, of high pressure and low pressure alternately, parallel to the equator. The winds blow into each low pressure and out from each high pressure, and these winds are permanent. Since the low-pressure belts are also permanent it, is evident that the winds which blow into them must escape upwards; and since the high-pressure areas are permanent, the air which blows outwards must be replaced by air from above. Therefore towards the middle of the low-pressure areas the air must be rising, towards the middle of the high-pressure areas the air must be sinking.

Where the air is rising and where the winds are blowing towards the poles the climate will be wet; where the air is sinking and where the winds are blowing towards the equator



RAINFALL
The rainy areas are shaded

PRESSURE
The high-pressure areas are shaded

FIG. 42. Relation of rainfall to pressure.

the climate will be dry. Thus we may divide the globe into a series of alternate dry and wet belts corresponding with the general distribution of pressure and the permanent winds (Fig. 42). Such a division is necessarily diagrammatic. There is no sharp line in nature between the belts, and there are modifications due to the distribution of land and sea and to the seasons; but the diagram represents very nearly the probable distribution of rainfall on a globe covered with water.

It should perhaps be noted that in the diagram the dry belts are made wider than the high pressure belts, because on the equatorward margins of the high pressures the winds blow towards the equator and are therefore dry.

The greatest rainfall will be in the middle of the equatorial low pressure, for there the rising air is most heavily charged with vapour.

Influence of land and sea and of the winds. In general the rainfall will be heavier over the sea than over the land,

FIG. 43. Influence of land and sea on rainfall.
In each belt the depth of the shading indicates roughly the amount of rain.

because there the supply of vapour is greater; and consequently if land and sea stretched from north to south in alternate bands, the general distribution of rainfall might be a little more accurately represented as in Fig. 43.

In this diagram the rainfall is represented as greater over the sea than on the land. In the temperate rain-belt the rain is due mainly to cyclonic depressions and is considerably greater near the coast than in the interior of the continents. In the equatorial belt the rain is due chiefly to the ascending air of that region and there is no very marked difference between the rainfall of the coast and that of the interior. In all the belts the amount of rain decreases gradually towards the edges of the belt; but to avoid complication no attempt is made to show this in the diagram.

A very marked effect, however, is produced by the winds. In temperate latitudes the prevalent winds are westerly. They bring the moisture of the ocean over the western margin of the continents, and they take the dryness of the continent over the margin of the sea that washes its eastern shores.

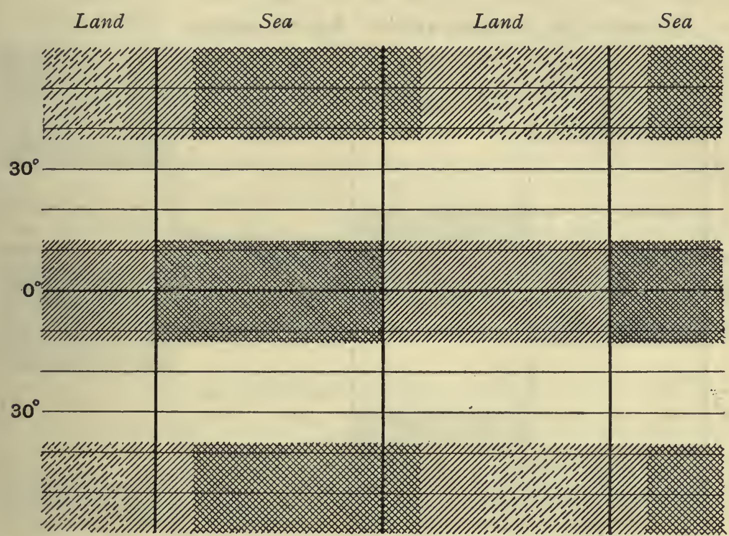


FIG. 44. Influence of winds on rainfall.

They shift, in fact, the whole band as shown in Fig. 43 a little to the east as in Fig. 44. Accordingly in these latitudes the western coasts of the continents have an oceanic rainfall, the interior and eastern coasts have a continental rainfall. The change from the oceanic to the continental rainfall will be gradual, since the winds will gradually lose their moisture as they proceed inland.

Within the tropics the trade-winds carry vapour from the oceans over the eastern margins of the continents. But because they are blowing towards the equator, as well as westward, they are becoming warmer and will have no tendency to drop their moisture unless they are forced upwards by rising ground. Their effect therefore will depend largely upon the configuration of the land. In favourable circumstances they may make the eastern margin of the continents within the tropics wetter than the western, but they will not necessarily do so.

The general distribution of rainfall over land and sea, allowing for the influence of the winds but neglecting that

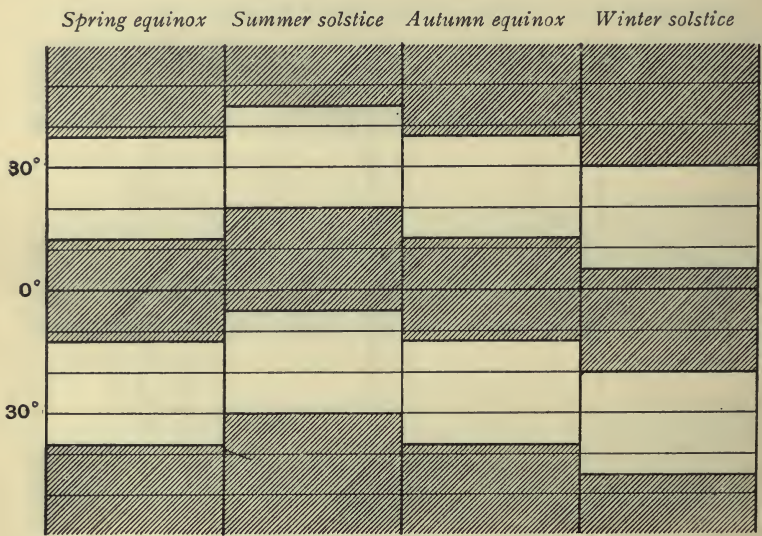


FIG. 45. Migration of rain-belts.

of altitude and of seasonal variations, will be as shown in Fig. 44.

Seasonal variations due to the migration of the rain-belts. As the year advances the position of the sun with respect to the earth is changed. There is a corresponding movement of the high- and low-pressure belts and along with these the wet and dry belts travel north and south. The extent of the migration varies in different parts of the world, but roughly it may be put at about 8° on each side of their mean positions.

Thus in the northern summer the middle of the equatorial rain-belt reaches 8° N., in the northern winter it has moved to 8° S. In consequence of these movements some parts of the world are in a rain-belt during one season and in a dry belt during another.

The approximate positions of the wet and dry belts over the oceans at the equinoxes and the solstices are shown diagrammatically in Fig. 45.

From these diagrams it is easily seen that any place between 30° and 40° north or south of the equator will be in the temperate rain-belt in the winter and in the tropical dry belt in the summer, and it will accordingly have a wet winter and a dry summer. This is the characteristic climate of a large part of the Mediterranean, and is accordingly often spoken of as the Mediterranean type of climate. For a reason which will be explained later it is not met with on the eastern coasts of the continental masses, but only upon their western sides. The Mediterranean itself, California, a part of Chili, the south-western corner of Africa, the south-western corner of Australia, and a small part of South Australia, all have their principal rains in the winter, while their summers are nearly dry.

Since there is no definite boundary between the wet and dry belts the limits of the Mediterranean climate are not sharply defined. There is a gradual transition from the more variable climate of the temperate rain-belt to the winter rains and summer droughts of the typical Mediterranean climate.

This is clearly brought out by the following table, in which the percentage of the total annual rainfall that falls during each season is given for several localities in the western half of the Spanish Peninsula.

	Latitude N.	Spring Mar.-May	Summer June-Aug.	Autumn Sept.-Nov.	Winter Dec.-Feb.
Oviedo ..	43° 28′	29·4	17·1	27·6	25·9
Guarda ..	40° 23′	31·2	8·0	30·3	30·5
Lisbon ..	38° 42′	30·3	3·4	28·0	38·3
San Fernando	36° 28′	29·0	2·3	29·0	39·7

Southwards the proportion of summer rain diminishes and the proportion of winter rain increases.

It is somewhat unfortunate that the term Mediterranean should have been applied to this type of climate; for in the Mediterranean itself the simple explanation given above is quite inadequate and the origin of the winter rains is much more complex. A glance at the map of the distribution of barometric pressure in January will show that the Mediterranean area lies on the southern instead of on the northern side of the high-pressure belt, and never comes within the region of the south-westerly winds. The prevalent winter winds in fact are northerly rather than southerly. In the western part of the basin these northerly winds may be looked upon as the south-westerly winds of the Atlantic deflected round the subsidiary anticyclone which covers Spain. But a large proportion of the Mediterranean rainfall appears to be due to low-pressure systems developed within the Mediterranean Sea itself, independently of the temperate belt of low pressure.

On the Atlantic coast of Spain and Portugal, however, the seasonal distribution of rainfall is due directly to the migration of the rain-belts in the manner described.

Places near the southern margin of the tropical dry belts, between latitudes 5° and 20°, north or south, will, on the other hand, be in the dry belt during the winter, while in the summer they will lie within the equatorial rain-belt. Therefore their winters will be dry and their summers wet.

In the middle of all the belts the seasonal changes will be slight. The equator is always in the rain-belt and hence the rain will be almost constant. In the greater part of the temperate regions there will also be rain at all seasons of the year. In the middle of the dry belts there will be practically constant drought.

Seasonal variations due to the distribution of land and sea. Hitherto we have imagined the wet and dry areas as continuous zones passing completely round the globe. But it has already been pointed out that, owing to the presence of the continental masses, the form of the high-pressure areas alters with the seasons; and there is a corresponding change in the distribution of rain.

Figs. 22 and 23 show diagrammatically the distribution of pressure in winter and in summer, and the directions of the winds due to such a distribution are indicated by arrows.

In our winter it will be observed that on the western coasts of the land, north of lat. 30° N. the winds are south-westerly. They blow from sea to land, and away from the equator. Therefore at this season the western coasts are wet, even down to the latitude of the Mediterranean. On the eastern coasts the winds are northerly and blow from land to sea; and therefore in winter these coasts are dry (Fig. 46 *a*). This is why the Mediterranean type of climate is not found on the eastern coasts.

In our summer the high pressure lies a little farther north, but is broken by the land-masses, which are now low-pressure

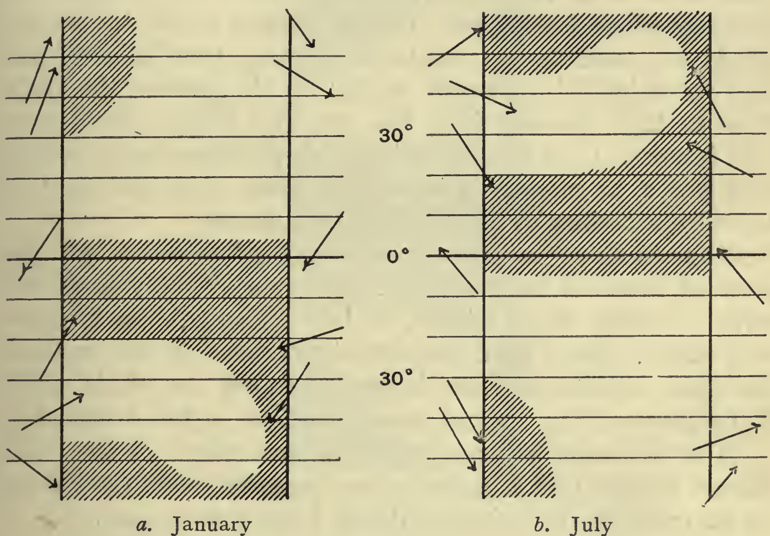


FIG. 46. Diagram of the seasonal distribution of rainfall on a continent.

areas. On the western coasts there are still south-westerly winds, but they do not extend so far south. About latitudes 30° to 40° N., the winds are more or less northerly, and though they come from the sea they are moving into a warmer area. Therefore they are dry and we get the summer drought of the Mediterranean type of climate. On the eastern coasts, the winds in the same latitudes are southerly and blow from sea to land. Therefore they bring rain (Fig. 46 *b*).

Thus on the western coasts between 30° and 40° north or south we have the Mediterranean type of climate with summer drought and winter rain; on the eastern coasts we have the Monsoon type with winter drought and summer rain. But the Monsoon type covers a wider extent in latitude than the Mediterranean type and towards the equator it merges into the similar type due to the migration of the equatorial belt of rain.

In the temperate zone outside latitude 40° N. the prevalent winds on the western coasts are from the south-west throughout the year and therefore there is rain at all seasons; but the maximum is in the winter, because then the gradient for westerly winds is steepest. On the eastern coasts the winter winds are more or less northerly, blowing from land to sea, and the rainfall is therefore small; in the summer the winds tend to blow inwards from the sea, and bring a little rain. The interior of the land-mass is a high-pressure area in winter and is dry, in summer it is a low-pressure area, and most of its rain, accordingly, falls during that season.

Thus in these latitudes, from about 40° to 60° or 70°, the western coasts of the continents have rain all the year round, with a maximum in winter, or late autumn; the interiors and eastern coasts have their heaviest rainfall in the summer, but even then the amount is small, because the winds which bring most of the vapour have travelled so far overland.

The following table, giving the percentage of the total annual rainfall that falls during each season, shows the change as we cross the Eurasian continent from west to east.

	Spring	Summer	Autumn	Winter
Ireland	21*	24	27	28†
East England.. ..	19*	28	30†	23
Central Germany ..	23	34†	23	20*
Central Russia ..	22	37†	25	16*
West Siberia	13*	42†	32	13*
East Siberia	12	58†	21	9*

* Minimum † Maximum

The tropical dry regions. Because of the seasonal changes in the distribution of rain, the dry regions of the tropics do not form a continuous belt around the globe or even across the land. On the western coasts they correspond fairly closely with the normal position of the dry belt as shown in Fig. 43. But the eastern coasts in the same latitude belong to the monsoon area and have heavy summer rains. Therefore on the western side of the continents the deserts extend to the sea, but on the eastern side they end before the coasts are reached. Towards the east, however, they tend to spread away from the equator, because in temperate latitudes the rainfall becomes very small in the interior of the continents.

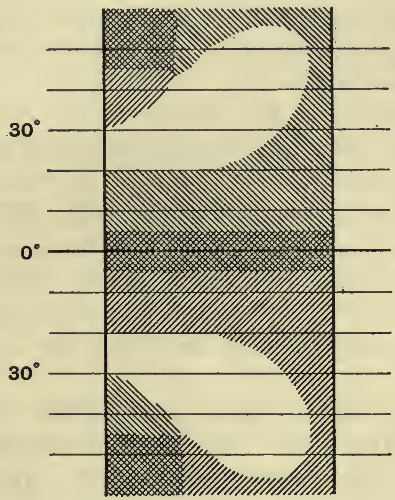


FIG. 47. Diagram of wet and dry areas on a continent. January rainfall indicated by shading inclined downwards from right to left; July rainfall by shading inclined downwards from left to right.

The general form which the dry regions tend to assume will be understood from the diagrams, Figs. 46 and 47. In Fig. 46 *a* the area of the January rainfall in a continent extending from north to south is shown diagrammatically. Considering only the northern hemisphere it will be seen that the equatorial rain-belt extends to 5° N. On the western coast the rainfall due to the westerly winds reaches southwards to 30° N. The interior and eastern coasts, as already explained, are dry.

In July (Fig. 46 *b*) the equatorial rain-belt extends to lat. 20° N. The temperate rain-belt reaches southwards only to 45° N., but it now stretches right across the continent, decreasing however in amount towards the east. On the eastern coast are the monsoon rains due to the winds blowing inwards from the sea. These rains reach northwards as far as the temperate rain-belt and thus the whole of the eastern coast is wet.

In Fig. 47 both the January and the July rainfall are shown, and it will be seen that the area which is dry in both months touches the western coast between latitudes 20° and 30° N. It expands inwards and northwards but nowhere reaches the eastern coast.

Owing to the irregular outline of the continents and to the influence of mountain chains the actual dry areas of the globe are not precisely of this form. But the map of annual rainfall (Map 7) shows that in the great land-masses of the northern hemisphere and also in Australia a region with a rainfall of less than 10 inches touches the western coasts about latitude 20° to 30°, expands inwards, especially on the side away from the equator, and does not reach the eastern coasts.

Influence of altitude upon precipitation. Half the water-vapour in the atmosphere lies below an altitude of 6500 feet, three-quarters lies below 13,000 feet. Therefore a high mountain chain is a very effective barrier to the passage of vapour, and accordingly we find that all the great ranges of the globe separate regions which differ from each other in climate. On the west of the Canadian Rockies is British Columbia with its heavy rains and equable temperature, on the east are the plains of Manitoba with a low rainfall and great extremes of temperature. On the south of the Alps is the plain of Lombardy, on the north the plain of the Danube, the one with a climate approaching the true Mediterranean type, the other with the climate characteristic of the temperate rain-belt.

But the influence of a mountain chain on rainfall is not due entirely to the fact that it rises like a ridge above a sea of vapour. Not only does it obstruct the easy passage of vapour from one side to the other, but it tends to cause the condensation of vapour on each side. For whenever a wind blows to the mountain chain it is forced upwards. The air expands and is cooled, and if it reaches a sufficient height, some of the vapour in it will condense, forming clouds and rain. Hence almost every mountain or mountain range has a heavier rainfall than the plains from which it rises. Even in the midst of the Sahara the ranges of Asben and Tibesti have regular seasonal rains.

If the mountain range is high the greater part of the vapour will be condensed on the windward side, and on the leeward side the rainfall will be small. It is chiefly for this reason that the west side of the British Isles is so much wetter than the east. In the southern half of France, which is also under the influence of the south-west winds, the heaviest rainfall is not on the west, but towards the east where the land is higher.

If the mountain range is low the rainfall on its leeward slope may be almost as great as on the windward side.

When the mountain range is very high, condensation will not necessarily continue to its summits. As the air ascends and is cooled, condensation will begin at a certain level, depending to a large extent upon the amount of vapour originally present. But as the air goes on rising, the quantity of vapour in it decreases. Sooner or later, as the air continues to rise, the amount condensed will begin to decrease, and finally condensation may entirely cease before the top of the range is reached. On a high mountain chain therefore there is a zone of maximum condensation, and this is usually far below the summits. Its altitude varies, chiefly according to the temperature and the amount of vapour present in the prevalent winds. On the Himalayas it is about 4000 feet above sea-level, but alters with the seasons. On our own mountains the rainfall is so variable that no definite statement is possible.

A curious instance of the influence of this zone of maximum condensation is given by Hann. In the Thian Shan, which rises from the arid plateau of Central Asia, the winter snows fall between the altitudes of 8000 and 10,000 feet. Below 8000 feet the range is dry and bare. Above 10,000 feet there is little condensation in winter, but in summer there are abundant rains. The zone of winter snows is also a belt of trees; above it, in the region of the summer rains, there is grass but no trees. The inhabitants move upwards in the winter, taking their flocks above the snows into the grassy belt (Fig. 48).

The snow-line. When the temperature is low the condensed vapour may fall not as rain but as snow, and the snow stays where it falls instead of running off like water. In polar

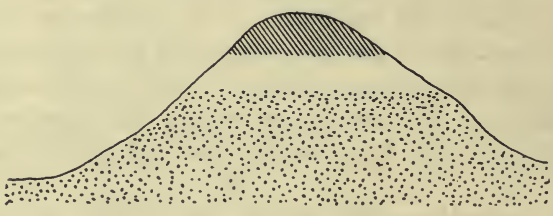


FIG. 48. Diagram of climate and vegetation zones in the Thian Shan. Obliquely shaded area :—summer rains, grass. Unshaded area :—winter snows, trees. Dotted area :—dry, barren.

regions the precipitation is in the form of snow throughout the greater part of the year. From the poles towards the equator the season during which the temperature is cold enough for snow decreases in length, and in general the annual fall of snow decreases, until about 40° from the equator it practically ceases at sea-level. At greater altitudes, however, snow continues to fall even to the equator itself.

Snow which has fallen on the ground may be removed in several ways. It may be blown away by the wind, it may slip down a mountain side as a snow-slide or an avalanche; it may be carried away slowly in the form of a glacier; or it may melt. But if it is not removed in some way it will stay where it fell, and every year will add to the thickness of the deposit.

Where the annual fall of snow equals or exceeds the amount removed in the year, the ground will be covered by a permanent layer of snow; where it is less the ground will be free from snow for a part of the year. In the polar regions the covering of snow is permanent even at sea-level, in lower latitudes it is only on the mountains that snow lasts throughout the year.

The lower limit of permanent snow upon the mountains is called the snow-line. It is not a sharp and definite line, for towards its lower edge the carpet of snow is ragged and irregular. Holes and rents appear, through which the bare ground shows. The covering becomes patchy, and downwards the patches grow smaller and more scattered until at length they disappear entirely.

The snow-line is at sea-level about the Arctic Circle and from there it rises, somewhat irregularly, to the equator.

The altitude of the snow-line is necessarily influenced by the temperature, and in general the higher the temperature at sea-level the higher also will be the temperature of the air above and the greater the altitude of the snow-line. Therefore there is a general rise of the snow-line towards the equator.

But the height of the snow-line does not depend entirely upon the temperature. It is influenced by the amount of snow that falls, and also by the form of the ground, for upon the latter will depend, in part, the amount of snow that stays. On a steep mountain slope much of the snow slides down in avalanches; on a gentle slope most of it must lie until it melts. In the latter case the snow-line will be lower than in the former.

The snow-line is the level where the amount of snow melted in the year is just equal to the amount which would otherwise collect. In a dry region therefore it will be higher than in a wet one, if the temperature, slope of the hill sides etc., are the same in both cases. Even if the dry region is the cooler of the two, its snow-line may still be higher. The Himalayas form a striking and well-known example. The snow-line on the northern slopes is at an altitude of about 18,000 feet, on the southern slopes at about 16,000 feet, although the southern slopes are naturally the hotter. But the difference in temperature is more than compensated by the difference in the annual snow-fall. The vapour brought by the south-west monsoon is mostly condensed on the southern side of the range, and comparatively little passes to the north.

The Himalayas are not a single chain, but a series of parallel chains, and although it is true of the range as a whole that the snow-line is higher on the north than on the south, yet with the individual chains the reverse is the case. This is shown diagrammatically in Fig. 49.

On each chain the precipitation is approximately equal on both sides, the sun is stronger on the southern side and therefore the snow-line slopes down from south to north. But the precipitation on the southernmost chain is greater than that on the next and accordingly its snow-line is lower.

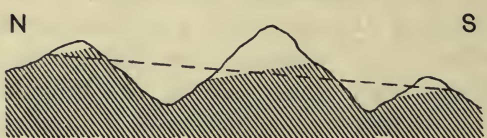


FIG. 49. Snow-line on the Himalayas.

In the same way the snow-line of each chain is lower than that of the chain north of it. Thus, although in each chain the snow-line slants down from south to north, yet taking the range as a whole the snow-line is higher on the northern side than on the southern, and the general slope, indicated in the figure by a broken line, is from north to south.

CHAPTER IX

THE OCEANS

General distribution of land and sea. If we look at a globe, with our eyes half shut so as to obscure all details, we shall see in the Western Hemisphere a large triangle of land, with its base in the north, against the Arctic Ocean, and its point in the south. In the Eastern Hemisphere the land is also widest towards the north, but instead of tapering southwards into a single point, it ends in two, the Cape of Good Hope and the island of Tasmania. It is like two triangles which have become united at their bases; and, as in the Western Hemisphere, the bases lie against the Arctic Ocean and the points are directed southwards.

Besides these three triangles, stretching from north to south, there is a fourth land-mass, of smaller size, with its centre about the South Pole.

Between the three triangular masses of land lie three still larger triangles of sea. But, necessarily, these are turned the other way. They are wide towards the south, where they unite with one another, and they taper towards the north In the Western Hemisphere is the Pacific, ending northwards at the Behring Straits. In the Eastern Hemisphere are the Atlantic and Indian Oceans. The former is fairly wide even at the Arctic Circle, the latter is completely closed towards the north by the union of the land-masses between which it lies.

There is still another area of sea, the Arctic Ocean, round about the North Pole.

The triangles, both of land and sea, are very irregular, but nevertheless it will be seen that there is a certain symmetry in the arrangement of the continents and oceans.

The tetrahedral theory. An explanation of this symmetry was put forward in 1875 by Lowthian Green. If a solid tetrahedron is placed in the midst of a globe of water, not sufficient to cover it (Fig. 50), the points and parts of the sides and edges will project above the water. They will form triangular masses of land, and between them, on the faces of the tetrahedron, will lie areas of water[1]. If one of the points of the tetrahedron is placed at the south pole and the other three symmetrically around the north pole, the arrangement of land and sea will be somewhat similar to that on the earth. There will be three triangles of land with their bases in the north and their points towards the south, and also a fourth

[1] It should be observed that if the tetrahedron has plane faces the water areas will be circles, more or less complete according to the extent of the projecting land-masses.

mass at the south pole. There will be three areas of water which will unite with one another towards the south and narrow towards the north, and also a fourth area at the north pole. The exact shapes of the land and the sea will depend upon the relative sizes of the globe of water and the solid tetrahedron.

Lowthian Green accordingly suggested that the solid earth is more or less tetrahedral in form; not that it is a regular tetrahedron with flat faces, but rather an irregular one with very convex faces.

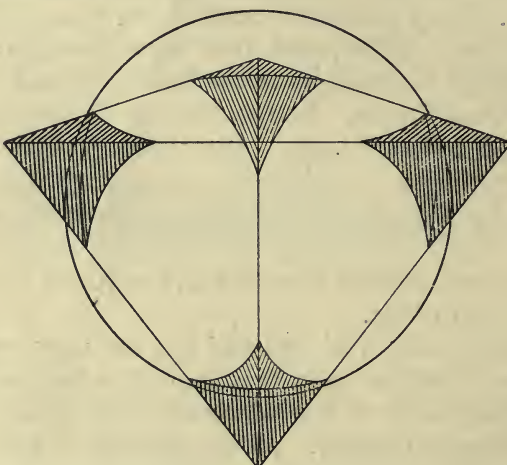


FIG. 50. A tetrahedron placed symmetrically in a globe.

To account for this form it is assumed that owing to loss of heat the earth is contracting. The interior is hotter than the exterior and is losing heat more rapidly and therefore contracts more quickly. Consequently in time there would be a space between the core and the outer crust, if it were not that the latter is not strong enough to support its own weight; it crushes inwards, like an arch that is too weak, in order to fit itself to the contracting core.

Fairbairn, the engineer, showed experimentally that when a cylindrical tube of iron is crushed by a pressure applied uniformly all round, it often assumes a triangular section; and this being so, it might be expected, from considerations of symmetry, that a hollow sphere would crush into a tetrahedral form.

It has since been proved that the shape which a cylindrical tube will assume when crushed inwards by a uniform pressure is influenced by the length of the tube; but the tetrahedron is still a probable form for a hollow sphere to take.

In the present state of our knowledge, however, the whole question of the causes which have determined the general form of the continents and oceans is highly speculative. The tetrahedral theory of Lowthian Green is only one of several, and there are serious objections to all. No general theory yet put forward shows any very close agreement with the history of the evolution of continents as deduced from a study of their geology.

Area and depth of the ocean. A glance at a terrestrial globe is sufficient to show that the area of the oceans is far greater than that of the land. According to the earlier estimates of Wagner water covers 71·7 % of the globe, land occupies 28·3 %. More recently Krümmel has given the figures as, water 70·8 %, land 29·2 %.

The chief reason for the difference between these estimates is our imperfect knowledge of the polar regions, both in the north and in the south. But the difference is not great, and in round numbers we may take the percentages as, water 71; land 29.

Attempts have also been made to determine the average height of the land and the average depth of the ocean, and also the proportion of land of various altitudes and the proportion of ocean of various depths. These estimates are necessarily even more uncertain than those of the relative areas of land and water. Sir John Murray's figures are as follows:

Land:

Height	Area	Percentage of whole globe
over 12000 feet	2 million sq. m.	1
6000–12000 „	4 „ „	2
3000– 6000 „	10 „ „	5
600– 3000 „	26 „ „	13
0– 600 „	15 „ „	8
	57 „ „	29

Sea :

Depth	Area	Percentage of whole globe
0– 600 feet	10 million sq. m.	5
600– 3000 ,,	7 ,, ,,	3
3000– 6000 ,,	5 ,, ,,	2
6000–12000 ,,	27 ,, ,,	15
12000–18000 ,,	81 ,, ,,	41
over 18000 ,,	10 ,, ,,	5
	140 ,, ,,	71

The hypsographic curve. If we take the figures in this table, and draw a diagram in which areas are represented by lengths along a horizontal line, and heights and depths, measured from sea-level, are represented by vertical distances from this line we obtain a hypsographic curve, as shown in Fig. 51. Such a curve represents in some respects the average form of the surface of the solid globe if the surface of the sea be considered as level; but the heights and depths are very greatly exaggerated.

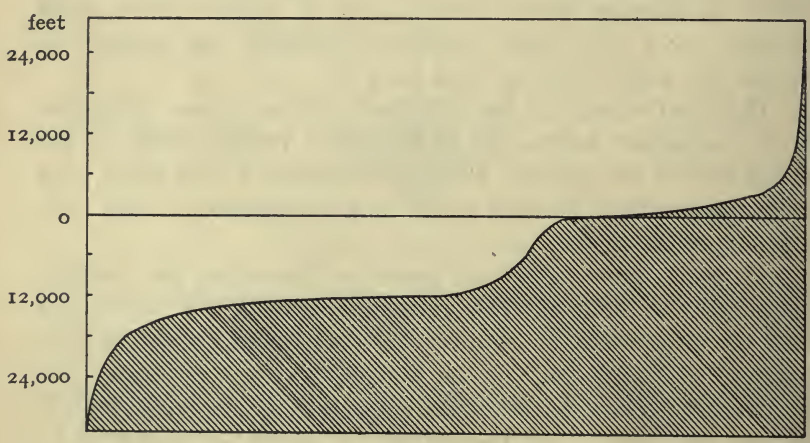


Fig. 51. Hypsographic curve.

From this diagram it will be seen that the ocean floor may be divided into four parts:

(1) The continental shelf; lying next to the land, and sloping very gently from the shore. It extends to a depth of about 100 fathoms.

(2) The continental slope; immediately outside the continental shelf, and sloping much more steeply. It must not be forgotten that owing to the exaggeration of the vertical heights in the diagram, the angle of slope is also greatly exaggerated. The continental slope extends from about 100 fathoms to about 2000 fathoms.

(3) The deep-sea plain; a broad and nearly level area forming by far the greater part of the ocean-floor. Its depth varies from about 2000 to 3000 fathoms but its slopes are always very gentle.

(4) The deeps; the deepest parts of the ocean, forming depressions in the floor, relatively small in area, and with comparatively steep sides.

In all the oceans these different parts can be readily recognised, although they may differ in proportions. In the Atlantic, for example, the continental shelf is generally wider than in the Pacific. In all oceans the greater part of the floor is formed by the deep-sea plain; and at various places in the plain are the deeper depressions known as Deeps. They are not as a rule in the middle of the oceans but rather towards the margins.

The continental shelf and slope. The continental shelf is the gently sloping part of the floor which forms a ledge around the continents. With it we may also include that part of a sea-beach which lies between tide-marks.

The continental shelf usually extends from sea-level to a depth of 100 fathoms. Its width varies greatly. Off the coast of Ireland it stretches westward for a distance of 50 miles or more; but it may be only a few miles wide, or it may even be absent altogether. The angle of slope is also variable and is usually least where the shelf is widest. West of Ireland it is less than one degree, and it is seldom more than two or three.

Seawards the continental shelf ends abruptly, and there is a rapid increase in depth beyond its outer margin. It is here that the continental slope begins, and the slope is really the edge of the shelf.

The angle of the continental slope varies far more than that of the shelf. Off Ireland, it is only about 5°; but off the coast of Spain it is much steeper and near Cape Toriñana it is in one place as much as 36°, a steep angle even for a mountain side.

The origin of the continental shelf is still uncertain. In the hypsographic diagram it appears as a direct continuation of the surface of the land beneath the level of the sea; there is no change of angle where the water begins. Many writers, accordingly, especially in America, hold that the continents really end at the outer margin of the shelf; that at some former time the sea reached only to the top of the continental slope and for some reason which is still unknown, it has since overflowed the edges of the continents.

In support of this view there is the undoubted fact that many river-valleys are continued beneath the sea, across the continental shelf, and open on the continental slope.

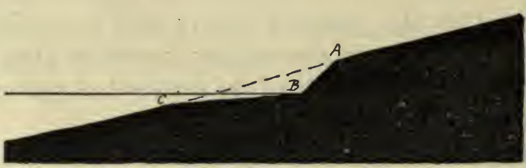


FIG. 52. Shelf cut by the sea.

According to this view the continental shelf was formed either by a rise in the level of the sea or a fall in the level of the land. Since the shelf is almost continuous around the shores of the Atlantic the movement must have affected half the globe, and it must have been remarkably uniform in amount. Unless there has been an increase in the quantity of water in the ocean it is difficult to understand how a change of level so widespread and regular could be brought about.

There are, however, other ways in which a ledge may be formed upon the margin of a continent. If the relative level of land and sea remains unaltered, the waves and currents will gradually wear away the edge of the land, cutting a notch in the original profile, as shown in Fig. 52. The side *AB* of the notch will form a cliff, and the floor *BC* will be a gentle slope, reaching from high-water mark to the depth where waves and currents cease to erode. The breadth of the platform will depend upon the resistance of the rocks, the strength of the waves and currents, and the length of time during which the level of land and sea remains unchanged.

Shelves of this kind exist round Iceland and the Färö Islands. On the coast of Norway a similar wave-cut platform has been lifted above the sea. But a shelf produced by erosion alone can hardly be more than a narrow fringe. The motion of the water due to a passing wave decreases very rapidly downwards and is scarcely felt at a depth of a hundred feet: and even currents seem to have little or no erosive action beyond 600 feet. Moreover, as the width of the platform increases, the force of the waves at its shoreward margin will diminish and at length they will cease to wear away the cliff.

FIG. 53. Shelf formed by deposition.

A shelf may also be formed by deposition (Fig. 53). The surface of the land is worn by rain and rivers and its edge by the waves of the sea. The broken material thus produced is laid down beneath the water, but always near the land. Off the mouth of the Amazon it is said that the sea is sometimes discoloured by mud at a distance of 300 miles. But even the largest rivers must deposit most of their burden near the shore. Waves and currents may bear it a little further, but their action is slight except in shallow water. Consequently the material derived from the land accumulates as a submarine terrace upon the margins of the continents. The edge of the terrace is the limit beyond which waves and currents cease to be effective, and within that limit the material collects and is distributed along the shore.

This limit, however, is not permanent, for as the terrace is built up and the water becomes shallower, the transporting action of the currents becomes more effective and they are able to carry the material farther than before. The terrace therefore grows gradually outwards by addition at its edge, in the same fashion as a railway embankment or the tip-heaps of a quarry.

According to this explanation the continental shelf and slope are formed by deposits brought from the land. The shelf is the surface of the deposit, smoothed and redistributed by waves and currents; the slope is the edge of the deposit, too deep to be thus affected.

The action of currents in general seems to cease at about 600 feet beneath the surface and therefore the edge of the shelf is usually found at about this depth.

The valleys which cross the shelf are not, according to this view, submerged valleys of erosion, but have been built up beneath the sea by the deposits brought down by rivers. When a rapid river enters the sea, its current continues outwards for some distance from the land. The current will be swiftest in the middle, while at the sides it will be retarded by friction with the water of the sea. Consequently the material which it carries will be laid down chiefly at the sides, and in the middle there will be but little deposition. Thus the river will form a channel on the shelf, not by excavating its bed but by building up the sides; and if the process continues long enough it may carry the channel right across the shelf. It is, however, evident that if the currents which sweep along the shore are stronger than the river, no such channel can be found. The current of the river must be strong enough to keep the middle of its channel fairly free from deposition.

The great width of the continental shelf in the northern part of the Atlantic may be due to deposition around the margins of the great ice-sheets which, in geologically recent times, covered the north of Europe and the north of America.

It should perhaps be added that a marginal shelf will seldom be due either to erosion alone or to deposition alone.

Both processes go on together and the shelf will be the result of their combined action.

The deep-sea plain. By far the greater part of the floor of the ocean is formed by the deep-sea plain. It is not actually a level surface and its depth accordingly is variable; but the slopes are so slight that they would be quite imperceptible to the eye.

The deep-sea plain is for the most part entirely free from the sediments brought down by rivers, but it is not formed of solid rock. A large part is covered by a kind of fine mud or "ooze" consisting of the shells and tests of minute animals and plants which during life float on the surface of the water. In the deeper parts there is a red clay, which appears to have been formed chiefly from the fine ash thrown out by volcanoes and carried out to sea by the wind.

The deeps. The deeps are depressions in the ocean floor still deeper than the deep-sea plain. Their sides are fairly steep and their area is generally small. They do not usually lie in the middle of the oceans, but towards the margins. Most of them are found near coasts where volcanoes are still active and earthquakes common. There are many, for example, around the shores of the Pacific, and there is one close to the West Indies.

The Atlantic Ocean. The Atlantic Ocean is rather irregular in outline. It is very wide towards the south where it opens into the Antarctic Ocean, but narrows towards the Equator between Africa and South America; it widens out between Africa and North America, and again narrows rapidly northwards. But it is not closed towards the north and communicates with the Arctic Ocean by a fairly wide opening.

On the borders of the Atlantic there are a number of seas which are more or less completely separated from the main ocean. In the north are Baffin's Bay and Hudson's Bay on the western side and the North Sea and the Baltic on the eastern side. All of these are shallow. Nearer to the equator are the Gulf of Mexico and the Caribbean Sea on the west and the Mediterranean on the east. All of these are deep.

The average depth of the Atlantic Ocean is rather more than two miles. The greatest depth which has been satisfactorily measured is 4561 fathoms in the Blake Deep, north of Porto Rico.

The continental shelf is generally well marked and is particularly wide in the North Atlantic, both on the eastern and the western coasts.

The deep-sea plain presents the usual characters; but it is not uniform in depth. Both from east and west it rises gradually towards the middle, forming a long and gentle undulation which divides the ocean longitudinally into two deeper basins. Except at the equator the crest of the rise is generally less than 2000 fathoms deep, and upon it stand the Azores, Ascension, Tristan da Cunha and one or two other islands. It is known in the North Atlantic as the Dolphin Rise, in the South Atlantic as the Challenger Rise; and it is often called the Atlantic Rise. It lies about half way between the eastern

FIG. 54. Diagrammatic section of the Atlantic Ocean.

and the western coasts, following the curves of the coasts with considerable fidelity.

Throughout the greater part of its extent the slopes on both sides of the Atlantic Rise are very gentle. But close to the equator the western side falls steeply to the Romanche Deep where a sounding of 4030 fathoms has been obtained. The only other deep with soundings of more than 4000 fathoms is the Blake Deep already mentioned.

The special characteristics of the Atlantic Ocean are the well-marked continental shelf, the rise in the middle, and the general absence of deeps. Diagrammatically therefore the general form of the section is as shown in Fig. 54, but it should be clearly understood that the figure is only a diagram, vertical heights are greatly exaggerated, and accordingly all slopes are very much steeper than in nature.

The Pacific Ocean. The Pacific Ocean is more regular in outline than the Atlantic and is roughly triangular in shape.

At the base of the triangle it is open to the Antarctic Ocean, at the apex it is almost closed, communicating with the Arctic Ocean only by the narrow Behring Strait. But the sides are not straight and the greatest width is not at the base but about the equator.

As in the case of the Atlantic there are a number of seas upon its margins which are more or less completely cut off from the main body of water. Most of them are upon the western side of the ocean. In the north there are the Sea of Okhotsk, the Sea of Japan and the Yellow Sea, all of which are comparatively shallow. Nearer to the equator are the China Sea, the Celebes Sea and a number of others amongst the islands of the Malay Archipelago. Most of these in their deepest parts are as deep as the ocean itself.

The average depth of the Pacific is about two and a half miles. Until recently the greatest depth actually measured

FIG. 55. Diagrammatic section of the Pacific Ocean.

was 5268 fathoms, near the Ladrone Islands; but a sounding of 5348 fathoms has now been recorded about 40 nautical miles east of Mindanao in the Philippines.

If we exclude the marginal seas on the Asiatic coast the continental shelf throughout the Pacific is either narrow or completely absent, and the edges fall almost uninterruptedly to the level of the deep-sea plain.

There is no continuous dividing ridge as in the Atlantic Ocean, but the surface of the plain rises up into a number of isolated plateaux, on which stand the groups of islands which are so characteristic a feature of the Pacific, especially of its southern portion.

Most of the deeps form narrow trough-shaped depressions along the margins of the Ocean. Amongst them are the Tuscarora Deep off Japan, the Atacama Deep along the coast of South America and many others. There are also deeps along the edges of some of the plateaux already mentioned. One of these stretches from the Tonga Islands nearly to New Zealand.

Thus the characteristic features of the Pacific Ocean are the feeble development of the continental shelf and the absence of any continuous dividing rise, the presence of numerous deeps along the margins and of isolated plateaux rising from the middle of the deep-sea plain. These characters are shown diagrammatically in Fig. 55, in which, as in the section of the Atlantic, vertical heights are greatly exaggerated. In this diagram one of the marginal seas of the North Asiatic coast is also shown. It will be seen that these seas occupy the position of the continental shelf, but they probably owe their origin to a different cause.

CHAPTER X

SALINITY OF THE SEA

Composition of sea-water. Sea-water is always salt, but the degree of saltness is not everywhere the same. In the North Sea, for example, the percentage of salt is less than in the midst of the Atlantic; in the Baltic it is very much less. In the Mediterranean, on the other hand, the proportion of salt is considerably greater than in any part of the open ocean.

The saltness is due mainly to chloride of sodium; but other salts are also present. On the average in 1000 grams of sea-water there are 35 grams of dissolved solids; and according to Dittmar these solids are constituted as follows;

Sodium chloride ($NaCl$)	27·213
Magnesium chloride ($MgCl_2$)	3·807
Magnesium sulphate ($MgSO_4$)	1·658
Calcium sulphate ($CaSO_4$)	1·260
Potassium sulphate (K_2SO_4)	0·863
Calcium carbonate ($CaCO_3$)	0·123
Magnesium bromide ($MgBr_2$)	0·076
	35·000

It is known now, however, that in a solution such salts will to a large extent be dissociated into their separate constituents. But the analyses imply that if a mixture of the salts in the proportions indicated were dissolved in the proper amount of water, the solution would be similar in composition to sea-water.

Besides the salts mentioned in the table many other substances are also found, but in such minute quantities that for most purposes they may be disregarded.

Although the total amount of dissolved solids in 1000 grams of sea-water varies in different parts of the ocean, the proportion of the different constituents to one another remains almost unaltered. If, for instance, the amount of sodium chloride is less than 27·213 then all the other constituents will be reduced in the same proportion.

The degree of salinity is usually expressed in parts per thousand. Thus the average salinity as given above is said to be 35 or, as it is often written, 35 ‰.

Specific gravity. The density or specific gravity of sea-water depends partly on its salinity and partly on its temperature. At freezing point water of the average composition given above, with a salinity of 35, has a specific gravity of 1·028. But as the temperature rises the water expands and the density decreases; and it is quite possible for water of low salinity and low temperature to be denser than water of high salinity and high temperature.

Origin of salt in the sea. The salt in the sea is no doubt derived in part from rivers. River-water always holds a certain amount of material in solution and most of this is carried into the sea. Evaporation removes water from the ocean but not the dissolved solids, and presumably therefore the sea is gradually growing more and more saline.

If this were the way in which the sea originally became salt, sea-water would be only a concentrated river-water. But it is by no means clear that the whole of the salt has come from rivers. There is a considerable difference between the salts in sea-water and those in an average river-water. In the former by far the greater part consists of chlorides,

FIG. 56. Salinity of the surface of the Atlantic Ocean.

especially chloride of sodium; in the latter the most abundant compounds are carbonates, especially carbonate of lime (calcium). Considering the very subordinate part played in sea-water by carbonate of lime, it appears at first sight improbable that sea-water could be produced by any concentration of the water of existing rivers. But this objection is not decisive. A very large number of animals which live in the sea, such as molluscs, corals, etc., form their shells or skeletons of carbonate of lime which they abstract from the water. Therefore the carbonate of lime brought into the sea is continually being used up, while the sodium chloride is left behind.

Possibly the difference in the relative abundance of carbonates and chlorides in sea- and river-water may be accounted for in this way; but there are other differences also. In the water of rivers, to judge from actual analyses, the sulphates exceed the chlorides and the proportion of potassium to sodium is very much greater than in the water of the sea.

The composition of river-water, however, varies greatly; and most of the rivers of the world have not yet been chemically examined. The analyses which have been made certainly suggest either that the salts of the sea have not been derived entirely from rivers or that in past times the composition of river-water must have been very different. But until a much larger series of analyses is available, from rivers in all parts of the globe, it is unsafe to assume that the average composition of river-water is even approximately known.

Distribution of salinity. The variations in the salinity of the surface waters of the ocean are due mainly to two causes. They depend upon the supply of fresh water and upon the rapidity of evaporation. The influence of these two causes is clearly shown in the map of the Atlantic, Fig. 56, in which isohalines, or lines of equal salinity, have been drawn. The map shows the salinity of the surface waters only. At a depth of 1000 fathoms or even 100 fathoms the isohalines would be very different.

The greatest proportion of salt is found in two areas which lie about the tropics of Cancer and Capricorn. From these regions the salinity decreases both towards the equator and towards the poles.

The high salinity at the tropics is due to the rapidity of evaporation beneath the clear skies and brilliant sun of the tropical anticyclones and the trade-wind region. The vapour which is formed is carried away by the winds and condenses elsewhere.

At the equator the heavy equatorial rains dilute the surface waters and the cloudiness of the region somewhat retards evaporation. About the mouths of the Niger and the Congo the salinity is particularly low, on account of the volume of fresh water poured out upon the surface by these rivers. Since fresh water is lighter than salt water (of the same temperature) the water of rivers may float for some time upon the sea before complete mixture takes place.

Towards the poles at certain seasons water that is almost fresh is supplied by the melting of the ice, and accordingly the salinity of the surface is low. For reasons which will be explained later this less saline water tends to flow especially along the eastern margins of the continents. As it moves towards the equator it encounters water of increasing salinity but also of increasing temperature. At first it floats upon the surface, but farther on the difference of temperature more than compensates for the difference of salinity; the cold polar water finds itself amongst water which is more saline but, on account of its high temperature, less dense than itself, and accordingly it sinks beneath.

Partially enclosed seas. Seas like the Mediterranean and the Baltic, which communicate with the ocean only by narrow straits, show much greater differences in salinity.

In the Mediterranean at the Straits of Gibraltar the salinity is about 36·5, but it increases eastwards and on the Syrian coast it exceeds 39. At the southern end of the Red Sea the salinity is about 36·5, but it rises towards the Gulf of Suez, where it is more than 41. These are examples of seas with a higher salinity than any part of the open ocean.

In the Black Sea on the other hand the surface salinity over the greater part is only 18 or 18·5, and in the Sea of

Azov it is considerably less. In the Baltic the salinity at the entrance varies considerably and is influenced, among other things, by the direction of the wind. About the island of Rügen it is only 7 or 8, and it decreases northwards. At the heads of the Gulfs of Bothnia and Finland the salinity is often less than 2, and in spring the water is practically fresh.

The salinity of these seas, like that of the open ocean itself, depends upon the relation between the supply of fresh water and the loss by evaporation.

In the Mediterranean and the Red Sea the evaporation is great and the rainfall small. The Red Sea is practically without rivers. The Mediterranean receives the waters of the Rhone, the Po and other rivers, but the total amount is small compared with the area of the sea.

In the Black Sea evaporation is less than in the Mediterranean, and many large rivers such as the Danube, the Dniester, the Dnieper and the Don bring a much larger supply of fresh water, in proportion to the size of the sea.

In the Baltic also, which lies in a colder region, evaporation is comparatively slow. A large amount of fresh water is received from the numerous rivers of Sweden and northern Russia, especially during the melting of the snow in spring and early summer.

Inland seas and lakes. In a lake which has an outlet there is no tendency for the water to be more saline than that of the rivers which flow into it, for the salt which they bring is carried away by the river that flows out. Absence of an outlet implies that evaporation is at least equal to the supply, for otherwise the depression would fill up until the water overflowed. In such cases there is no escape for the salts brought in by the rivers which enter the lake, and the water must gradually grow more and more saline. The degree of salinity will depend in part upon the length of time that the lake has existed without an outlet.

Even in the same lake, however, there may be considerable variations. In the northern part of the Caspian the salinity is less than 14, while in the shallow Gulf of Karabugas, which is connected with the rest of the Caspian only by a narrow opening, the salinity reaches 170. In this gulf evaporation is rapid and there is a constant inward stream of water from the Caspian to make up for the loss.

The Dead Sea is the saltest of the larger lakes, the salinity reaching 237·5, but even this is exceeded by some of the smaller lakes in the dry regions of the globe.

In most of these inland seas and lakes the composition of the dissolved salts is not the same as in ordinary sea-water. Sodium chloride is usually, though not always, abundant; but other constituents play a more important part than in the oceans.

CHAPTER XI

TEMPERATURE OF THE OCEAN

Surface temperature. At the equator the average temperature of the surface water is about 80°, and it decreases, in general, towards the poles. At about 28° sea-water freezes, and the temperature of the open sea cannot therefore fall below this point, but the temperature of the covering of ice may sink far below zero.

In enclosed seas in the tropics the temperature rises considerably higher than 80°. In the Red Sea it reaches 90° or even more.

Although in general the temperature of the surface water decreases gradually from the equator to the pole, there are considerable irregularities. They are not so great as on the continents, but even on the oceans the isotherms are seldom strictly parallel to the equator. The western part of the North Atlantic, for example, is much colder than the eastern part in the same latitude. The causes of these irregularities have already been explained. The most important are the winds and currents and the distribution of land and sea.

Vertical distribution of temperature. Even at the equator it is only the surface water that is warm, the deeper water is always cold. Fig. 57 shows diagrammatically the results of the observations of the *Challenger* in 24° 20′ N. 24° 28′ W.

Depths are represented by distances along the vertical line and temperatures by horizontal distances to the right.

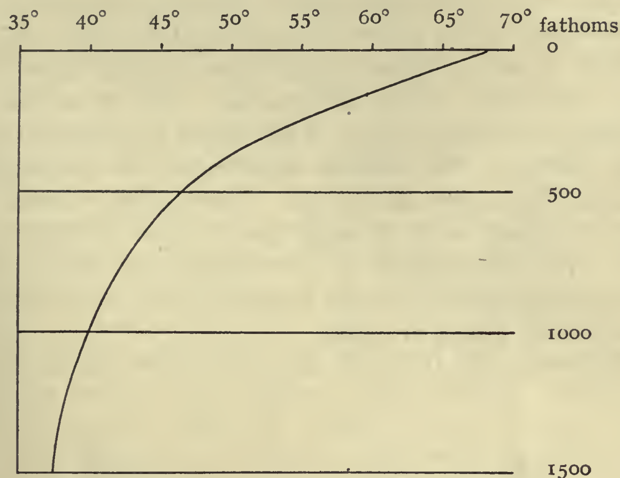


FIG. 57. Temperature curve, 24° 20′ N., 24° 28′ W.

Fig. 58 shows in a similar fashion the average temperature at different depths for the whole ocean, as calculated from the whole series of observations made by the *Challenger*.

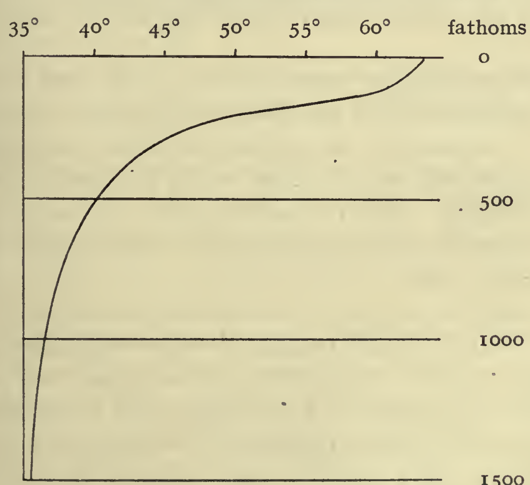


FIG. 58. Average temperature curve.

It will be seen that for the first few hundred feet the decrease of temperature is very rapid, but it then becomes very slow. Beyond 1000 fathoms there is scarcely any change and the whole of the water below this depth is about 35° or 36°.

Where there are warm or cold currents, and near the polar regions, the form of the temperature curve is modified, but throughout the greater part of the ocean in tropical and temperate latitudes the vertical distribution of temperature is similar to that shown in Fig. 58. There is everywhere a thin layer of warmer water floating upon a great thickness of cold water. The temperature of the surface layer varies with the place of observation, but the temperature of the deeper water is everywhere nearly the same.

The ocean, like the land, receives its heat mainly from the

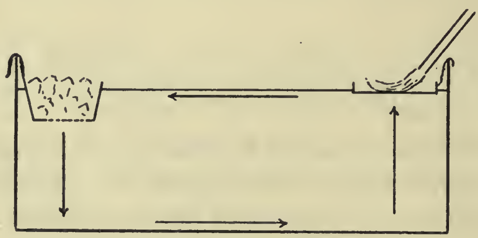


FIG. 59. Currents produced by differences of temperature.

sun. It is accordingly warmed from above, and since water is a very bad conductor the downward passage of the heat is very slow. If, however, the sun were constantly shining and the water were perfectly still the heat would gradually penetrate to the lowest depths. But towards the poles the sun's rays have very little power even in summer, and accordingly the water there is always cold.

The conditions are easily illustrated by a simple experiment (Fig. 59). Take a small trough of water and allow it to rest till the water is still. If the temperature is uniform the surface of the water will be level. At one end of the trough place some ice, which may be kept in position by a strainer or a net of wire-gauze. At the other end fix a plate of metal just touching the surface of the water—a convenient method is to float the lid of a small tin box upon the water, mooring it by means of a wire passed over the edge of the trough. Direct the flame of a bunsen burner against the upturned surface of the lid. The water in contact with the lid is warmed by the bunsen flame and accordingly expands; at the other end the water is cooled by the ice and contracts. At the heated end the surface will be slightly raised by the expansion, at the cooled end it will be slightly lowered by contraction, and therefore a surface current of warm water will flow from the bunsen flame to the ice. But because some of the water is now removed from the warm end to the cold end, the pressure at the bottom of the tank will be increased at the cold end and decreased at the warm end. Therefore, an undercurrent of cold water will flow from the cold end to the warm end. At the cold end the water will slowly sink owing to the addition of water above, at the warm end the water will slowly rise owing to the addition of water below; and as long as the difference of temperature is kept up the circulation will go on. The presence of the lower current may be shown by dropping a crystal of permanganate of potash into the water beneath the ice, the presence of the upper current by throwing some light powder upon the surface. The difference of temperature may be observed by placing a thermometer with its bulb just below the surface and another with its bulb at the bottom of the tank.

The heated end of the tank represents the equator, where the sun warms the surface; the cooled end represents the pole, where the water loses its heat.

It is because of this continual flow of the warm water on the surface towards the pole and of the cold water beneath towards the equator that the deeper water of the ocean is always cold. The water does not stay long enough at the equator to be heated to any great depth.

Temperature in enclosed seas. That this is the true explanation of the coldness of the deeper water in the open ocean is clearly shown by the vertical distribution of temperature in enclosed seas such as the Mediterranean and the Red Sea.

In the Mediterranean the average temperature of the surface water near the Strait of Gibraltar is about 65°, which is about the same as that of the surface water of the Atlantic outside the Strait. For the first 190 fathoms the decrease is nearly the same as in the open ocean, and the temperature falls to 55° F. But beyond this depth there is no further change and even in the deepest parts of the sea the temperature is still 55° F. (Fig. 60).

The Mediterranean is divided from the Atlantic by a ridge at the Strait of Gibraltar, and over this ridge the greatest depth is 190 fathoms. Therefore, down to 190 fathoms there can be interchange of water between the Atlantic and the Mediterranean, and the temperatures are nearly the same. But the deeper and colder water of the Atlantic cannot enter the Mediterranean; and the temperature of the bottom of the Mediterranean is the same as that of the coldest waters which can reach it from outside.

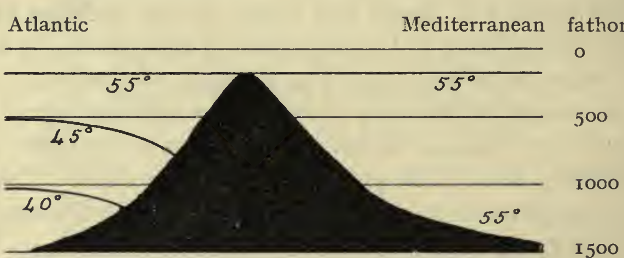


FIG. 60. Temperatures of the Atlantic and Mediterranean.

It is the same with the Red Sea. The average temperature of the surface water near the Strait of Bab el Mandeb is about 85°, while in the Indian Ocean it is rather less. In the Red Sea, as in the Indian Ocean, the temperature falls rapidly to a depth of 200 fathoms, where it is 70°. In the Indian Ocean it continues to decrease beyond that depth, but in the Red Sea the whole of the water below 200 fathoms is at 70°. The depth at the entrance from the Indian Ocean to the Red Sea is about 200 fathoms, and as in the case of the Mediterranean the temperature of the deep water of the Red Sea is the same as that of the coldest waters which can enter the sea.

If the Red Sea or the Mediterranean were completely shut off from the ocean, the temperature at the bottom would no doubt gradually rise until it was equal or nearly equal to that at the surface.

Effect of the winds on the distribution of temperature. It has been shown on p. 134 that the expansion of water by heat tends to raise the level of the ocean at the equator above its level at the pole, and consequently to produce a movement of the surface waters from the equator to the pole. But since it is only the upper layers that are warmed by the sun, the difference of level is small and the movement due to this cause must be very slow and practically imperceptible.

Indirectly, however, the difference of temperature produces much greater effects, for it is the cause of the winds, and the movement of the surface waters appears to be governed mainly by the direction of the winds.

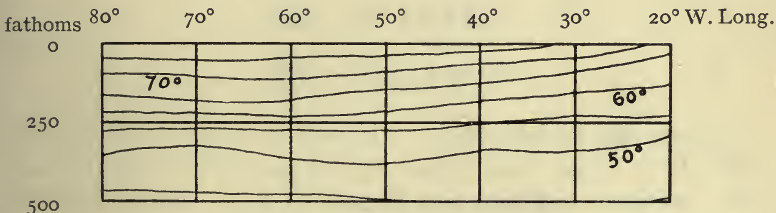


Fig. 61. Isotherms of the Atlantic in 20° N. Lat.

Fig. 61 is a section of the water of the Atlantic along the parallel of latitude 20° N. down to the depth of 500 fathoms. The temperature is shown by means of isotherms and most of these, it will be observed, slope downwards from east to west. Everywhere the temperature decreases downwards but not everywhere at the same rate. The warmest water is at the surface on the American side of the ocean, while on the African side the surface water is comparatively cold.

In this latitude the prevalent wind is the north-east trade. The warm surface water is therefore continually blown towards the American coast, while on the African side its place is taken by cooler water rising up from below (cf. p. 58).

Fig. 62 is a similar section along the parallel of 50°, where the ocean is much narrower. The isotherms in this diagram slant downwards from west to east. The warmest water is on the European coast, while on the American side even the surface is cold. The prevalent winds are from the south-west, and blow the surface waters towards the east, while on the American coast the cold water wells up from below.

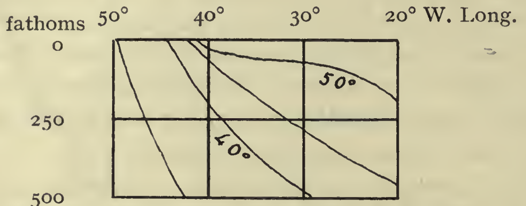


FIG. 62. Isotherms of the Atlantic in 50° N. Lat.

CHAPTER XII

WAVES AND TIDES

Movements of the ocean. The water of the ocean is never still. It is blown into waves by the wind, it rises and falls with the tides, and in many places there are definite currents either permanently in one direction or changing with the tide or with the season.

Waves. When water is thrown into waves, its surface takes the form shown, in section, in Fig. 63[1]. The waves travel in some definite direction, but a cork thrown into the water does not travel with the waves. It moves up and down, to and fro, but unless it is blown by the wind or carried by a current it returns to the same position with each wave and does not permanently leave its place. The cork must move with the water on which it floats, and thus it is clear that although the wave travels forward the particles of water do not.

The highest part of the wave is called the crest, the lowest the trough; the distance from crest to crest, or from trough

[1] The profile of an ordinary wave in water is approximately a trochoid. It is similar to the curve traced by a marked point on a carriage-wheel when the carriage travels in a straight line on a level road, but compared with this curve it is inverted in position.

to trough, is called the length of the wave and the vertical height of the crest above the trough is the height or amplitude.

In deep water the motion of the particles at the surface is nearly circular. At the crest the movement of the particle is forward, at the middle of the hinder slope it is downward, in the trough backward, and at the middle of the front slope upward.

The wave is felt beneath the surface, but the amount of movement diminishes rapidly downwards. The up and down movement decreases more quickly than the movement to and fro and thus beneath the surface the motion of the particles becomes elliptical. At a depth equal to the length of the wave the extent of the movement is only about $\frac{1}{500}$th of the extent at the surface. Consequently waves have very little

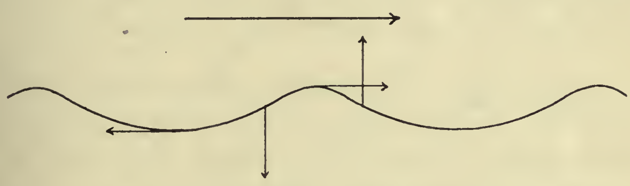


FIG. 63. Profile of ordinary waves.
The long arrow shows the direction of movement of the waves ; the short arrows show the direction of movement of particles at different points of the wave.

effect excepting near the surface, and even in the stormiest seas the disturbance is confined to a shallow layer of the water.

Our natural impressions of the height of waves are greatly exaggerated. When a wave dashes against a cliff the water may be thrown up a hundred or two hundred feet or even more ; but it is no longer part of the wave. On board ship an approaching wave seems higher than it is, because the ship is on the slope of the preceding wave and is heeling towards the one that is approaching. The highest wave actually measured by Scoresby was 43½ feet, and it is probable that in the open ocean the height of waves formed by the wind seldom exceeds 50 feet.

Speed of waves. The speed of waves depends partly upon their length and partly upon the depth of the water.

When the water is shallow relatively to the length of the wave, the velocity depends on the depth alone and is proportional to the square root of the depth.

When, on the other hand, the water is deep relatively to the length of the wave, the velocity depends on the length alone and is proportional to the square root of the length.

If the water is neither shallow nor deep, relatively to the length of the wave, the velocity is affected both by the depth and the length. Roughly speaking, we may say that if the depth of the water is greater than half the length of the wave the velocity depends chiefly on the length ; if the depth of the

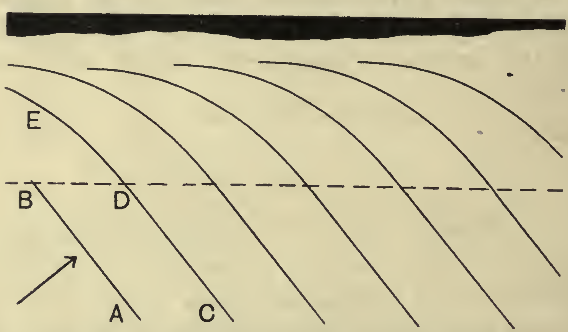


FIG. 64. Waves approaching a shelving shore.

The arrow indicates the direction of the wind. The broken line marks the position where the water becomes shallow enough to affect the speed of the waves.

water is less than half the length of the wave, the velocity depends chiefly on the depth.

Consequently, in the open ocean the speed of a wind-wave depends upon its length, but on a shelving shore it depends upon the depth of the water.

It is for this reason that on a sandy beach the crests of the waves are nearly parallel to the shore whatever the direction of the wind may be.

On such a shore the depth of the water gradually increases outwards. In the open sea the crests of the waves are at right angles to the wind, as shown by the lines *AB*, *CD* (Fig. 64). But as they approach the coast the end *E* of the wave

CDE will be in shallow water before the end *C*, and will accordingly travel more slowly. Consequently, the wave gradually turns until its crest is nearly parallel to the shore.

If the coast is steep and the water is deep to the foot of the cliffs, the waves will not be retarded as they approach and will keep their original direction.

Breaking of waves. When a wave approaches a shelving shore it keeps its form as a wave until it is near the land and then the top falls forward and the wave breaks. This, like the turning of the wave, is due in part to the fact that the wave travels more slowly as the water becomes shallower. The front of the wave is in shallower water than the back and therefore moves more slowly. The back gains upon it and the front becomes steeper, until in fact the crest is practically unsupported in front, and then, because of the forward movement of the particles at the top of the wave, the crest falls forward.

When the water is deep close up to the shore, the waves, if they break at all, break in a different fashion. They appear to throw themselves against the cliff and the water dashes up the face of the cliff, sometimes to a very great height. In front of the crest the particles of water in a wave are moving upwards and forwards, and when they strike the cliff the rebound continues to carry them upwards. Moreover, in a complete wave the upward movement of these particles is hindered by their cohesion with the lower part of the wave in front of them. But when they reach the cliff there is no part of the wave in front and this hindrance is removed.

Tides. From the earliest times it has no doubt been known to dwellers on the coasts of tidal seas that there is some connection between the tides and the moon. The tides are highest when the moon is either new or full, and at full moon the interval between the time of high tide and the time when the moon is on the meridian is approximately constant for each locality. But the interval is not everywhere the same. In some places it is one hour, in others two and so on, and, moreover, even at the same place it varies to some extent. It is easy, therefore, to see that there is some connection between the tides and the moon, but it is not easy to see what that connection is.

The cause of the connection was not understood until the time of Newton. Gravitation is a mutual attraction which exists between all particles of matter in the universe. The earth attracts the moon and the moon attracts the earth. The moon attracts every particle of matter in the earth. The amount of the attraction varies inversely as the square of the distance, and the moon, therefore, attracts the part of the earth which is nearest to it more strongly than the parts which are farthest away.

The diameter of the earth is about 8000 miles. Consequently the side of the earth facing the moon is 4000 miles nearer to the moon than the centre, and 8000 miles nearer than the opposite side. The moon, therefore, attracts the near side of the earth more strongly than it attracts the centre, and the far side less strongly. If the whole planet were solid the effect would be slight, but since the earth is almost surrounded by the liquid ocean, the consequence is that the waters yield and are heaped up beneath the moon. On the other side the attraction is less than at the centre and the waters being less attracted than the solid earth bulge outwards on the side away from the moon.

There is a difficulty in appreciating this explanation of the bulge upon the side opposite to the moon. The earth does not approach the moon and, therefore, we naturally look upon it as fixed in position with regard to the moon, and we imagine the moon as revolving round it. If this were so it would mean that the solid earth did not yield at all to the attraction of the moon. In that case there would be a bulging of the waters on the side facing the moon but none on the opposite side.

But, apart from its revolution round the sun, the earth is not fixed in position and it yields to the attraction of the moon. The moon does not revolve around the earth, but both earth and moon revolve around their common centre of gravity. Owing to the much greater mass of the earth this common centre is situated about 1000 miles beneath the surface of the earth. The relative positions of the earth and moon at different stages in a complete revolution are shown in Fig. 65, in which the common centre of gravity, *G*, is a fixed point, which does not alter its position. When the moon is at *a* the earth is at *A*; when the moon is at *b* the earth is at *B*; and when the moon is at *c*, the earth is at *C*. In this diagram, if the mutual attraction were suddenly to cease, the earth and moon would fly off in the direction of the arrows (*i.e.*, the direction in which they were actually moving at the time) and would separate from one another. But their mutual attraction

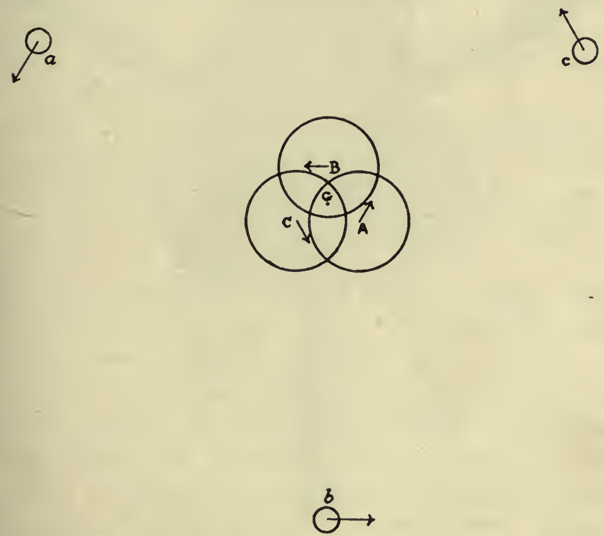


FIG. 65. Relative movements of the earth and the moon.

draws them together, with the result that they do not separate but revolve around the centre *G*. Thus although they do not actually approach one another they are continually yielding to each other's attraction and do not follow the course which they would otherwise take. When it is once clearly understood that the earth is yielding to the attraction of the moon it is easy to see that the solid earth will yield more than the water which is on the side away from the moon and, therefore, on that side the waters will bulge outwards, as well as on the side facing the moon.

We may also look at the question in a slightly different way. Take a lath or a strip of thick cardboard and fix it to a blackboard by means of a drawing pin *G*, round which it is free to turn (Fig. 66). The pin should be much nearer to one end than to the other. At the end of the longer arm fasten a small card disc to represent the moon. Take a larger disc to represent the earth and by means of a drawing-pin through

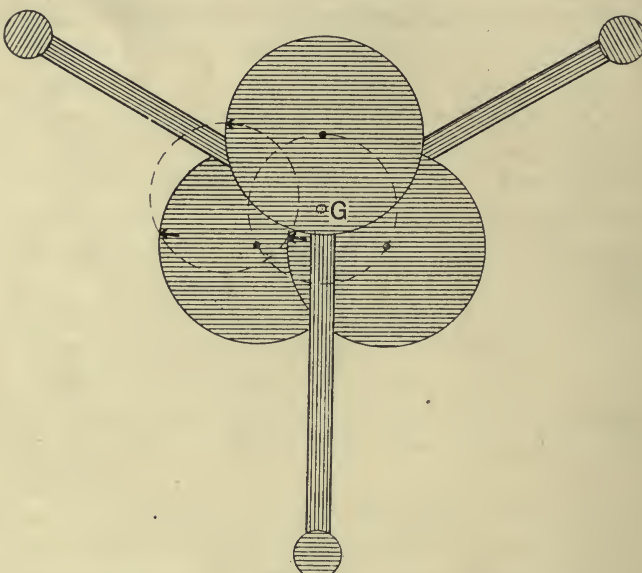


FIG. 66. Model to show the relative movements of the earth and the moon.

The model is shown in three positions of a complete revolution. The broken circles are the circles described by the centre of the large disc and by the point of the arrow marked upon its edge, when the large disc is not allowed to rotate on its own axis.

its centre fix it to the shorter arm of the lath in such a way that its edge overlaps the pin *G*. The disc representing the earth should be free to turn round the drawing-pin through its own centre.

Such a model as this will show very well the relative movements of the earth and the moon. The common centre of gravity is represented by the pin *G*, round which both discs can revolve; and the disc representing the earth can rotate about its own centre.

It is important to remember that although the earth *revolves* around the common centre of gravity it does not *rotate* about that point. Its rotation is round its own axis, and has nothing to do with the revolution round *G*, or with the production of the tides. If, therefore, we wish to consider the effect of the movement of the earth round the common centre of gravity we must put aside its rotation round its own axis and make it move round *G*, without rotating. To assist in doing this it is well to draw an arrow at the edge of the disc representing the earth and to remember that this arrow must always point in the same direction.

Take hold of the large disc, and without allowing it to turn on its own axis, make it move round the pin *G*. The lath will swing round and the smaller disc with it; and the movements of the two discs will be similar to those of the earth and the moon.

Next, while doing this, hold a piece of chalk against the edge of the large disc in such a way as to touch the board. As we make the revolution the chalk will draw a circle.

Repeat the experiment with the chalk at other points on the edge of the large disc, or passed through a hole in the disc. It will be found that it always draws a circle, and that all these circles are equal in size and are described in the same direction; but they are round different centres. Moreover, if two chalks are used at the same time it will be found that they are always at corresponding positions in their circles at the same moment.

Thus, putting aside its rotation round its own axis, every part of the earth during a lunar month describes a circle, and all these circles are of the same size and are described in the same direction. In every part, therefore, centrifugal force is developed and it is everywhere of the same amount. Moreover, since all parts of the earth are at corresponding points of their circles at the same time the centrifugal force is everywhere in the same direction. At the centre of the earth, which describes its circle round the point *G*, it is easy to see that the centrifugal force is directed away from the moon ; and everywhere else it will be parallel.

Accordingly at all points of the earth there are two forces, an attractive force directed towards the moon and a centrifugal force directed away from the moon. The latter is everywhere the same, but the former is greatest on the side facing the moon.

Since the earth and moon neither approach nor recede from each other, it is evident that at the centre of the earth the centrifugal force is just equal to the attractive force. On the side nearest the moon the attractive force is greater than at the centre, and is therefore greater than the centrifugal

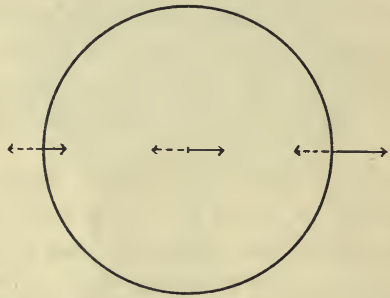


FIG. 67. Tide-producing force of the moon.

The moon is supposed to lie to the right of the diagram. The solid arrows represent the attractive force of the moon ; the dotted arrows represent the centrifugal force.

force ; and there is accordingly a surplus force pulling the waters towards the moon. On the side farthest from the moon the attractive force is less than at the centre, and is therefore less than the centrifugal force ; and there is a surplus force directed from the moon. Thus the bulging of the water beneath the moon is due to the excess of the attractive force above the centrifugal force, the bulging on the other side is due to the excess of the centrifugal force above the attractive force (Fig. 67).

In this way the moon causes the water to be drawn towards the part of the earth facing it and also towards the opposite side, and thus produces high tide (Fig. 68). But the water has been drawn *away* from the other parts of the earth, and about half way between the two high tides the sea is below its normal level and there is low tide. As the earth rotates on its axis every meridian comes in turn beneath each of the high tides and each of the low tides, and accordingly there are in most places two high tides and two low tides in the day.

FIG. 68. Tidal effect of the moon.

But while the earth is rotating on its own axis the moon is moving in the same direction round the centre of gravity of the earth and moon. Consequently after the earth has made a complete rotation it has still to turn a little farther before it brings the same meridian again beneath the moon (Fig. 69). Therefore high tide each day is later than it was the day before. The difference is roughly about 50 minutes, but it is sometimes more and sometimes less.

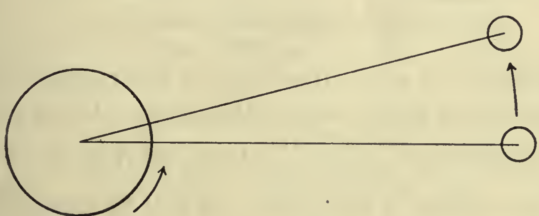


FIG. 69. The daily change in the time of high tide.

Effect of the sun's attraction. The moon is not the only heavenly body that draws the waters of the ocean towards it. The sun also produces a very distinct effect. But although the attraction exerted by the sun is far greater than that of the moon its influence upon the tides is less. For the tides are not determined by the amount of the attractive force, but by the *difference* in the attraction at the centre of the earth on the one hand and at the near side and the far side of the earth on the other hand. The centre of the moon is about 240,000 miles from the centre of the earth, or 236,000 miles from the nearer side. Its attractive force on the near side is to its attractive force at the centre in the proportion of $240,000^2$ to $236,000^2$, or nearly 31 to 30. The difference, which is the tide-producing force, is $\frac{1}{30}$th of the force exerted by the moon at the centre of the earth.

The distance of the sun is about 93,000,000 miles, and its attractive force on the nearer side of the earth is to its attractive force at the centre in the proportion of $93,000,000^2$ to $92,996,000^2$ or about 1,000,086 to 1,000,000. The difference, or tide-producing force, is $\frac{86}{1,000,000}$ of the attractive force of the *sun* at the centre of the earth. The mass of the sun is about 25,500,000 times that of the moon, but owing to its much

FIG. 70. Spring tides (at new moon).

greater distance its attractive force at the centre of the earth is only about 169 times that of the moon. Its tide-producing force is accordingly $\frac{86 \times 169}{1,000,000}$ of the attractive force of the *moon* at the centre of the earth. It will be found that this is about four-ninths of the tide-producing force of the moon, as given in the preceding paragraph.

Thus the tides are caused chiefly by the moon, but the sun has sufficient effect to modify them.

Spring and neap tides. When the earth, the sun and the moon are in the same straight line, which is nearly the case at full moon and at new moon, the bulging of the waters due to the sun is coincident with that due to the moon. The tides are, therefore, greater than usual, the high tide is higher and the low tide lower, and they are known as spring tides (Fig. 70).

When, however, the moon as seen from the earth, is 90° from the sun, as it is at the half-moons, the sun tends to produce high water where the moon would produce low water, and *vice versa*. Consequently there is a smaller difference than usual between high water and low water. High tide is lower and low tide higher than usual, and they are known as neap tides (Fig. 71).

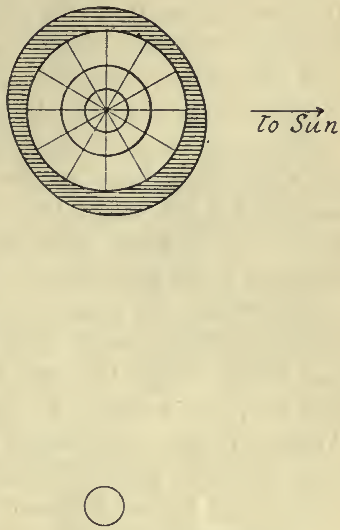


FIG. 71. Neap tides.

Influence of the continental masses. From the preceding account it might be inferred that at each place high tide should occur when the moon is on the meridian; and all places in the same longitude should have their high tide at the same time. But reference to an almanack will show that this is not the case. Liverpool and Leith are both about three degrees west of Greenwich but there is a difference of three hours or more in the time of high water.

Moreover, we might expect that the tide would go round the earth from east to west, following the apparent motion of the moon; but our tide reaches us from the Atlantic and travels from west to east.

These anomalies are due chiefly to two causes. In the first place it should be observed that in a globe completely surrounded by water the tides, as shown in Figs. 68–71, have the form of waves, the high tides being the crests and the low tides the troughs. The length of the wave is the distance from crest to crest, and in any given latitude this is half the circumference of the globe at that latitude. The tidal wave, like other waves, tends to travel round the earth at its own proper rate, independently of the sun and moon. Since the length of the wave, excepting close to the poles, is very great compared with the depth of the ocean, the rate depends on the depth of the water (see p. 140); and near the equator it is less than the rate at which the sun and moon appear to travel in their daily course. It is easy to see that this difference between the natural rate of the wave and the rate forced upon it by the sun and moon, will introduce complications, but these complications are too intricate to be considered here.

The movement of the tidal wave is also greatly influenced by the continental masses. If the globe were completely surrounded by water the tide would travel smoothly round from east to west. But the great land-masses stretch from north to south and prevent its progress. Only in the Antarctic Ocean is there an open and unobstructed course, and it is only in this southern sea that the tides are free to follow the moon and travel round the world from east to west, like two long but very low waves. As the wave passes the opening of the Atlantic it sets up a branch wave which travels up the Atlantic from south to north. Similar branch waves are formed in the Pacific and Indian Oceans.

The general course of the tidal wave in the Atlantic is shown in Fig. 72 by means of co-tidal lines. A co-tidal line is a line drawn through places which have their high tide at the same time, and the numbers placed against the lines indicate the Greenwich time of high tide on the days of full moon. But the map must not be regarded as anything more than a highly generalised approximation to the truth. In a large area like the Atlantic the complications are so great that the tide cannot be correctly represented as a simple progressive wave.

From this map it appears that the Atlantic tidal wave travels more rapidly in the deep water of mid-ocean than in the shallower water along the shore, and the crest becomes curved. As it passes northward the convexity of the curve increases, and when it reaches Europe the trend of the wave is from north to south and the wave approaches from the west. On the opposite side of the ocean the trend is also nearly north and south, but the wave approaches the coast from the east.

Tides in British seas. The tide accordingly reaches the British Isles from the west and its course within our seas is

FIG. 72. Co-tidal lines of the Atlantic (after Airy).

shown by co-tidal lines in Fig. 73. The wave is divided into two parts when it meets Ireland. The northern part sends a

FIG. 73. Co-tidal lines of the British Seas.

small off-shoot into the Irish Sea, but the main wave travels round the north of Scotland and gives rise to a wave passing from north to south along the eastern coast of Great Britain The southern part of the wave from the Atlantic is again divided by the Cornish peninsula, one branch going up the Irish Sea and Bristol Channel and another up the English Channel.

In the case of a sea where the tide can enter from both ends, there may be various anomalies. There may, for example, be four high tides in the day, two of them due to the tidal waves entering at one end of the sea and two due to the tidal waves entering at the other end. The best known example of this kind is the stretch of water between the Isle of Wight and the mainland, which is reached by tidal waves travelling up both the Spithead and the Solent. If, on the other hand, at any place the high tide due to the tidal wave from one end coincides with the low tide due to the tidal wave from the other end, the two will neutralise each other more or less completely and there may be no tide at all.

From the co-tidal maps it will be evident at once that the speed of the tidal wave is much greater in the open ocean than in shallow seas. In mid-Atlantic its velocity is six or seven hundred miles an hour, on the east coast of England it is about forty. The length of the wave, which on the map is the distance from any of the co-tidal lines to the next line marked with the same number, is great compared with the depth of the ocean, and therefore, as we have seen on p. 140, the velocity varies as the square root of the depth.

Height of the tides. In the open ocean, where there is nothing to interfere with the movements of the water, the tides are small, the difference between high and low water being only a few feet. But when the tidal wave enters a shallowing sea the front of the wave is retarded and the back gains upon it. The length of the wave (that is, the distance from the crest of one high tide to the crest of the next) decreases, but the height increases. Consequently in the shallow seas around our coasts the tides are usually considerable, the difference between high and low water amounting in many places to 20 or 30 feet or more.

When the tidal wave enters a gulf which not only shallows but also narrows inwards, the effect is still more marked. The wave as it travels on is confined within a continually decreasing space and to compensate for this it must increase in height. For this reason the tides in the Bristol Channel are exceptionally high. The difference between high and low water at Bristol during spring tides is 42 feet. In the Bay of Fundy, in the Dominion of Canada, the tides are even greater, the difference between high and low water being as much as 70 feet.

When, on the other hand, the tidal wave enters a sea which widens inwards, the wave as it travels on is spread over a wider space and its height accordingly diminishes. In the greater part of the Mediterranean, for example, the difference between high and low water is only a few inches. The tidal wave which enters from the Atlantic has very little effect. A small tide is produced by the action of the moon and sun upon the waters of the Mediterranean itself, but this is also inappreciable except in narrowing gulfs.

Bores. In the open ocean the tidal wave is symmetrical, but when it reaches shallowing water the front is retarded more than the back. The back gains upon the front and the front accordingly becomes steeper, as in the case of a wind-wave. If the retardation is sufficiently abrupt the front may become so steep that the crest begins to fall over and forms a line of tumbled waters like a broken wave upon a sandy beach.

It is in the mouths of rivers, where the retardation of the tide is great and sudden, that this most frequently occurs, and the broken crest is known as a bore or eger. The tide is no longer a simple wave, in which the particles of water move to and fro; the wave has broken and, as in a broken wave upon the shore, the water now moves bodily forward. Its front edge is the bore and appears as a wall of tumbling waters advancing up the river. In the Severn the bore at spring tides is often 3 or 4 feet high. In the Tsien-tang-kiang in China it is 12 feet.

Tidal currents. When high tide occurs at any place the level of the water is raised, and it is therefore evident that water has been drawn towards that place from some other part of the sea. At low tide the level of the water is lowered, and therefore water must have been drawn away from the place in question. Thus the existence of tides implies the existence of tidal currents.

Such currents must exist even in the open ocean, but high-water and low-water are there some thousands of miles apart, the rise and fall is only a few feet, and the currents are imperceptible. On our coasts it may be high-water in one place while it is low-water in another place less than a hundred miles away, the rise and fall may be 20 or 30 feet, and the currents accordingly are often strong.

In an ordinary wind-wave the particles in the upper part of the wave have a forward movement, in the lower part or trough the movement is backwards. Towards the middle of both slopes the particles move simply up and down. It is the same with the tidal wave. Consequently for some two hours before and two hours after high tide there are currents in the direction in which the tide is travelling; for some two hours before and two hours after low tide the currents are in the opposite direction. This may be looked upon as the normal condition, but near land these currents are liable to be modified by the form of the coast.

When the tidal wave advances at right angles to the shore the currents will flow directly to and from the shore. But when, as in the English Channel, the tidal wave travels along the coast, the currents will also run along the coast. It might be thought that these currents would neutralise each other's effects; that in the English Channel, for example, the forward currents would drift the sand and shingle from west to east, and the backward currents would drift it back again. But, at least in shallow seas, the forward currents are usually stronger than the backward currents, and the general tendency of the tidal currents is to drift material in the direction in which the tide is moving.

Consequently, in the English Channel there is a gradual drift of the sand and shingle along the coast from west to east; along our eastern coast there is a similar drift from north to south.

Tidal currents usually reach their maximum in straits, where the flow is concentrated within a narrow channel. They are exceptionally strong, for instance, in the Menai Straits and in the Pentland Firth.

CHAPTER XIII

CURRENTS

BESIDES the tidal currents already described and the general drift of the surface waters from the equator to the poles due to difference of temperature, there are also currents produced by other causes.

Currents due to difference of salinity. In the straits of Gibraltar a surface current passes almost constantly from the Atlantic into the Mediterranean. It is strongest when the tide outside the straits is rising; but it does not altogether cease even when the tide is falling, and, therefore, it is not caused entirely by the tide. Beneath the surface current there is another current flowing in the opposite direction, from the Mediterranean to the Atlantic. The lower current is more saline than the upper one.

In the Mediterranean evaporation is great and the supply of fresh water from rain and rivers is comparatively small. If the straits of Gibraltar were closed the sea would gradually dry up, but since they are open, water flows in from the Atlantic to make up for the loss by evaporation, and this is the cause of the surface current in the straits.

The cause of the lower current will be most simply illustrated by considering an imaginary case. Take a trough filled with salt water and divided completely into two compartments, A and B, by a vertical partition. If the water is of the same density throughout and stands at the same level in the two compartments, the removal of the partition will make no difference and there will be no tendency for the water to run from either compartment to the other. But suppose that some of the water in B evaporates but none of that in A. The density of the water in B becomes greater and the level of the surface lower. If an opening is made near the bottom of the partition there will be a slight flow from *A* to *B* on account of the loss of water in *B*, but this will soon cease; and because the water in *B* is denser than that in *A*, the level of the surface in *B* will still be lower than in *A*. If an opening is now made at the top of the partition a surface current will flow from *A* to *B*, owing to the difference of level. As soon as this happens the equilibrium at the bottom will be disturbed, because water has been removed from *A* to *B*. The pressure at the bottom becomes greater in *B* than in *A*, and a current will flow through the lower opening from *B* to *A*.

These two currents will continue to flow until the average density in the two compartments is the same and thus the level of the surface and the pressure at the bottom also become the same in both compartments. But if the difference of density is kept up by continual evaporation in *B*, the currents will be permanent. This is what happens in the Mediterranean; owing to evaporation the Mediterranean water is always denser than that of the Atlantic, and the currents are therefore permanent.

From these considerations we may conclude that wherever two seas of different salinity are connected, there will always be a surface flow from the fresher to the salter sea and an under flow from the salter to the fresher.

There are many other examples of currents due to the same cause. In the Mediterranean evaporation exceeds the supply of fresh water and the sea is therefore salter than the waters of the open ocean; in the Black Sea, with its numerous large rivers, evaporation is less than the supply and the water is fresher. Consequently there is a surface current of fresher water from the Black Sea to the Mediterranean and a deeper current of salter water from the Mediterranean to the Black Sea.

The water of the Baltic is considerably fresher than that of the North Sea. Therefore, there is an outward flow through the Skager Rak at the surface, while deeper down there is an inward flow of salter water from the sea outside.

Ocean currents. Excepting near the land, in shallow seas, or in polar waters the currents produced by the tides or by differences of salinity are generally too slight to be perceptible. But even in the midst of the great oceans there are currents which are often of considerable strength, and the origin of these was long a matter of dispute.

At one time it was suggested that the ocean currents were caused by the difference of temperature between the poles and the equator. As already explained this difference would produce a surface flow of warm water from the equator to the poles and a deeper flow of cold water from the poles to the equator. These flows would be deflected by the land-masses and also by the rotation of the earth, and it was thought that in this way the great ocean currents might be accounted for.

The coldness of the deeper water, even at the equator, indicates that these flows do actually exist; but the rate of movement must be very slow. It is only the upper layers of the water that are warmed and expanded by the heat of the sun and the gradient produced by the greater expansion near the equator must be very slight. There are, moreover, well defined currents flowing parallel to the equator. They are certainly not due directly to differences of temperature and they can hardly be accounted for by deflection.

There is little doubt now that the ocean currents are due mainly to the winds. Most of the principal currents flow in the direction of the prevalent winds, and where the winds change with the seasons, the currents also change.

The relations of the currents to the winds will be explained more fully in dealing with the currents of the different oceans.

Currents of the Atlantic Ocean (Fig. 74). North of the equator there is a current called the North Equatorial Current, flowing from east to west; south of the equator is a similar current known as the South Equatorial Current; and between the two is a weaker and less constant current called the Equatorial Counter Current, flowing from west to east.

The South Equatorial Current divides off Cape St Roque into two branches, one passing westwards along the northern shore of South America and the other southwards along its eastern shore.

Both the North Equatorial Current and the northern

FIG. 74. Currents of the Atlantic Ocean.
Polar currents indicated by dotted arrows.

branch of the South Equatorial Current reach the West Indies. A part of the water skirts the outer shores, but the greater part streams through the chain of islands into the Caribbean Sea and thence into the Gulf of Mexico. Where these seas are wide the movement is slow, but where they narrow it becomes more rapid.

From the Gulf of Mexico a current issues through the Florida Channel and, uniting with the branch which passed outside the West Indies, forms the Gulf Stream.

The Gulf Stream flows along the coast of the United States to Cape Hatteras and then, without changing its direction, passes out into the Atlantic. About 45° W. longitude it loses its character as a well-defined current, but there is still a general drift of the waters towards the east. This drift gradually spreads and divides, one branch flowing north-eastwards, and the other eastwards. The north-eastward branch, with various ramifications, reaches the coasts of Ireland, the British Isles and Norway, and sends some of its water into the Arctic Ocean. The eastward branch flows towards Spain and turning southwards skirts the coast of Africa and joins the North Equatorial Current.

From this description it will be seen that in the North Atlantic there is a complete circulation around a central area in which there is no movement. In this central area lies the Sargasso Sea, where floating sea-weed grows and is sometimes so abundant as to interfere with the progress of ships. The circulating waters are joined on the south by the northern branch of the South Equatorial Current, while on the north they send a branch into the Arctic seas. Thus the North Atlantic receives water from the South Atlantic and sends water into the Arctic Ocean.

The southern branch of the South Equatorial Current flows southwards along the eastern coast of Brazil to about latitude 30° S., where it leaves the coast and begins to bend eastwards, though some of its waters travel many degrees farther south before they turn. It flows across the ocean to South Africa, bends northwards along the African coast, where it is known as the Benguela Current, and rejoins the South Equatorial Current.

In the South Atlantic, therefore, as in the North Atlantic, there is a complete circulation around a central area which is free from currents. In the far south, where the Atlantic opens into the Antarctic seas, there is a continuous drift of the surface waters to the east, and this is called the Antarctic drift.

Besides the currents which have been described, all of which originate in the Atlantic Ocean and consist of Atlantic waters, there are others which enter the Atlantic from outside and which, because they come from Polar Seas, are necessarily cold.

A cold current flows down the eastern coast of Greenland sending a branch south-eastward towards the British Isles. But this branch soon sinks beneath the warmer waters of the North Atlantic drift.

Another cold current flows from Baffin's Bay past Labrador and Newfoundland, and keeping between the Gulf Stream and the American coast finally disappears about Cape Hatteras. This is called the Labrador current.

A similar current, occupying a similar position in the South Atlantic, flows from the Antarctic Ocean up the eastern side of South America, keeping between the southern extension of the Brazil Current and the coast of the continent. This is known as the Falkland Current.

These currents from the Polar Seas are usually less salt than the water into which they flow, because they are formed partly by the melting ice. They are in fact due, in part at least, to the difference of salinity which is thus produced. Because their water is fresher than that of the Atlantic, it floats at first upon the surface, and the currents are surface currents. But as they travel onwards into warmer parts of the ocean, the coldness of their water more than compensates for its lack of saltness; it is denser than the surrounding warmer water and accordingly these currents sink beneath the surface and disappear. This is what happens both to the Labrador Current and the Falkland Current.

It is not only in these currents from the Polar Seas that the surface water of the Atlantic is below the normal temperature for the latitude. Off the African coast, except for some few degrees on each side of the equator, the water is cold compared with that of mid-ocean in the same latitude, so cold in fact that reef-building corals do not grow upon this coast. The cause of this coldness has already been explained. It is due to the trade-winds, which blow the surface water towards the American side, so that the deeper and colder water wells up on the African shore.

It is probable that both the Labrador and the Falkland currents are due in part to a similar cause. The westerly winds blow the surface water to the east, and its place is taken partly by water from below, partly by water from the Polar Seas.

Relation of the Atlantic currents to the winds. In the region of the trade-winds the surface waters are continually blown towards the west, forming the North and the South Equatorial Currents. Against the American coast they are heaped up, so that the level there is slightly higher than on the eastern side of the Atlantic. Where the trade-winds are strong their tendency to blow the water westward is sufficient to maintain the difference of level and even to force the water up the slope; but in the equatorial calms, between the north-east and the south-east trades, there is no wind to influence the movement of the water, and in this space, therefore, some of the heaped up water flows back towards the lower level and forms the Equatorial Counter Current.

But the greater part of the water is still carried onwards by the trade-winds. It is forced by the shape of the land to divide into two branches, one passing northwards and the other southwards. The northern branch, in which both the North and the South Equatorial Currents have a share, finally becomes the Gulf Stream. The southern branch, which is formed by the South Equatorial Current alone, is the Brazil Current. Both branches skirt the shore until they come under the influence of the westerly winds, when they are carried back across the Atlantic to the opposite coast. Here again their course is determined by the shape of the land. In the North Atlantic a part of the water passes through the opening between Iceland and Scandinavia, a part turns down the Spanish coast, comes once more under the influence of the north-east trades and joins the North Equatorial Current. In the South Atlantic the current strikes the African coast, is deflected northward into the region of the south-east trades and joins the South Equatorial Current.

Thus, both in the North Atlantic and the South Atlantic the trade-winds blow the water towards the west, the westerly winds blow it towards the east. Each system of winds forces the water against a coast, and it is, at least in part, deflected along the coast until it comes under the influence of the other system. A complete circulation is accordingly established, around an area in which there are no currents. This area, both in the North and in the South Atlantic, is the region of the permanent tropical anticyclone, in which the air is almost calm ; and the currents follow very closely the course of the winds around the anticyclone.

The cold Labrador and Falkland Currents are probably due in part to the westerly winds, in part to the lower salinity of the Polar Seas. The westerly winds, blowing the water towards the east, tend to heap it up on the European and African coasts and to produce a slight depression of level upon the American shore. Towards this depression the polar waters tend to flow, and the flow is assisted by the difference of level due to the lower salinity of the polar seas. As already explained, the westerly winds, by blowing the surface waters towards the east, would cause an upwelling of cold water on the American shore, and accordingly the water of the Labrador and Falkland Currents may be in part derived from below.

Currents of the Pacific Ocean (Fig. 75). The circulation of the water in the Pacific Ocean is very like that in the Atlantic, but the shape of the coast-line introduces certain differences.

There is a North Equatorial Current and also a South Equatorial Current, flowing from east to west, and between them is an Equatorial Counter Current flowing from west to east.

The North Equatorial Current strikes the Philippines and turns northwards, flowing past the islands of Japan, where it is known as the Kuro Shiwo. About latitude 45° N. the current crosses the Pacific to the American shore. Some of its waters bend northwards along the coast of British Columbia, but the greater part turns southwards and rejoins the North Equatorial Current.

Owing to the numerous groups of islands in the South

FIG. 75. Currents of the Pacific Ocean.
Polar currents indicated by dotted arrows.

Pacific, the South Equatorial Current gives off several branches before it reaches the western margin of the ocean. All of these turn southwards into the region of the westerly winds, where the water is carried eastwards to the coast of South America. Here it turns northwards, and together with branches from the Antarctic drift forms the Peru Current and ultimately joins the South Equatorial Current.

Thus in the North Pacific and in the South Pacific, as in the North and the South Atlantic, there is a circulation of the water around a central area, and the circulation is due to the same cause, the trade-winds blowing the water towards the west and the westerly winds blowing it towards the east. At the southern opening of the Pacific, as in the case of the Atlantic, there is an easterly Antarctic drift.

But the shape of the coast-line causes certain differences. In the Atlantic, owing to the form and position of South America, the South Equatorial Current sends a large and important branch into the North Atlantic; in the Pacific there is scarcely any transference of the surface water from one side of the equator to the other. In the northern part of the Atlantic some of the water is driven between Iceland and Scandinavia into the Polar Sea; but the northern outlet of the Pacific is so narrow that the surface water finds little room for escape in this direction.

In the Pacific, as in the Atlantic, there are cold currents and upwellings of the deeper and colder water, and they occupy similar positions. From the Arctic Ocean, a current flows through the Behring Straits and along the fringe of islands off the coast of Asia, keeping on the inner side of the Kuro Shiwo. It corresponds with the Labrador Current. In the South Pacific, a current flowing north-eastwards between Tasmania and New Zealand represents the Falkland Current of the South Atlantic, but is less strongly marked because the land-mass of Australia and Tasmania barely reaches the region of the westerly winds.

In the Pacific there is upwelling of cold water on the coast of California and on the coast of Peru and Northern Chili, similar in position and origin to the upwellings on the coast of Africa.

Currents of the Indian Ocean. It is perhaps in the Indian Ocean that the connection between winds and currents is most clearly shown, for there the winds change with the seasons and the currents alter with the winds.

In the winter months of the northern hemisphere (Fig. 76), a current flows along the southern shore of Asia from east to west. It is produced by the north-east monsoon, just as the North Equatorial Current of the Atlantic or Pacific is produced by the north-east trade-wind.

South of the equator is the South Equatorial Current due to the south-east trade, and between this and the current already mentioned there is a weak and irregular Equatorial Counter Current flowing from west to east. The Counter Current is in the belt of calms, which at this season lies a little south of the equator.

The northern current, when it strikes the Arabian and

FIG. 76. Currents of the Indian Ocean (January).

African coast, is deflected to the south and joins the circulation of the southern part of the Ocean. The South Equatorial Current is similarly turned southwards by the land-mass of Africa, until it reaches the latitude of the westerly winds. Here the surface waters are driven eastwards to the Australian coast, where they are in part deflected to the north and rejoin the South Equatorial Current.

Thus at this season, when the winds are similar to those of the Pacific, the circulation of the water is also similar. The only important difference is that owing to the form of the coast, the North Equatorial Current is not deflected to the north but to the south, and it consequently crosses the equator and enters the southern part of the sea.

In the summer of the northern hemisphere (Fig. 77) the current along the southern shore of Asia flows from west to east. There is no north-east trade-wind and therefore there is no North Equatorial Current. Instead, the water north of the equator is carried eastward by the south-west monsoon. When it meets the peninsula of Further India it is deflected southwards and joins the South Equatorial Current.

South of the equator the south-east trade-winds and the westerly winds still blow as before, but are shifted somewhat to

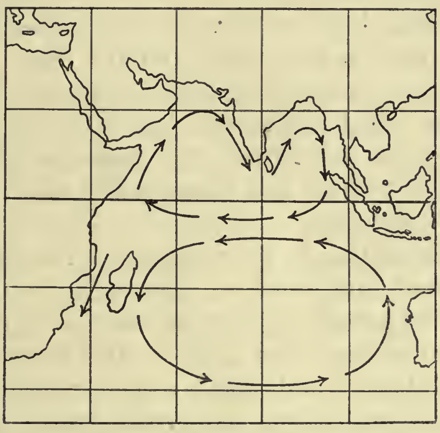


FIG. 77. Currents of the Indian Ocean (July).

the north. Therefore the general circulation is unchanged in direction, but is slightly altered in position. The South Equatorial Current is a little further north, and it is joined, as already mentioned, by the monsoon current. It is carried westwards by the south-east trade-wind; and, because it is so close to the equator, on the African coast a part of the water comes under the influence of the south-west monsoon and is turned towards the north; the rest is deflected southwards as in the season of the north-east monsoon.

During the prevalence of the south-west monsoon there is no equatorial belt of calms in the Indian Ocean, and, therefore, there is no counter current.

CHAPTER XIV

DEPOSITS ON THE OCEAN FLOOR

Terrigenous and pelagic deposits. Excepting near a rugged coast the bed of the sea is seldom formed of solid rock. The rock which no doubt exists beneath is usually covered by a layer of loose material which has been deposited upon the ocean floor.

There is a broad and general distinction between the deposits of the continental shelf and slope on the one hand and those of the deep-sea plain and the deeps on the other. The former consist mainly of material derived from the land, and are therefore often called terrigenous deposits; the latter, known as pelagic deposits, are formed to a large extent of the shells and skeletons of animals and plants which when alive float on the surface of the water.

But the distinction is not absolute. The deposits of the continental shelf and slope are not entirely of terrigenous origin; and the pelagic deposits are not entirely composed of the remains of animals and plants. The former include, for example, shell-banks and coral reefs; whilst amongst the pelagic deposits one of the most wide-spread types consists chiefly of the products of volcanic eruptions.

Moreover, there is no sharp line of demarcation. Pelagic deposits may extend far up the continental slope, and in seas that are free from sediment they are occasionally found even at depths of 200 fathoms.

Pelagic deposits, formed of the remains of floating animals and plants, are probably accumulating all over the ocean. Near the continental masses they are completely masked by the much greater amount of material which is brought down from the land, but where the proportion of terrigenous matter decreases the pelagic deposits begin to show.

Thus near the continents the deposits are mainly of terrigenous origin; away from the continents, even on the ridges and plateaux which rise nearly to the surface of the water, pelagic deposits predominate.

In the shallow waters of the continental shelf and slope both animals and plants flourish abundantly upon the bed of the sea and their remains are added to the material derived from the land. They sometimes constitute, in fact, the greater part of the deposit ; but such deposits must be distinguished from the true pelagic deposits formed of the remains of animals and plants which float upon the surface.

The larger fragments thrown out from volcanoes fall mostly on or near the land. When, however, they are of the nature of pumice, they may float for days and may be drifted into the open ocean before they become waterlogged and sink. The finer products of an eruption, known as volcanic dust, may be blown thousands of miles by the wind. It is said, in fact, that at the eruption of Krakatoa in 1883 much of the dust was carried three times round the globe before it fell. No part of the ocean, therefore, is beyond the reach of material of this kind, and volcanic dust is found even at the greatest depths.

The finer particles of soil from a desert are also sometimes blown far out to sea. In higher latitudes ice-bergs may carry sand and mud and even boulders into the deep ocean and on melting drop their burden on its floor. In these ways terrigenous material may be mingled with the pelagic deposits proper to mid-ocean, but only the volcanic dust occurs in any considerable quantity.

In general, therefore, the deposits on the continental shelf and slope consist mainly of :

(1) Material derived from the wear and tear of the land.

(2) The remains of animals and plants that live on the bed of the sea.

(3) Volcanic material.

The deposits on the deep-sea plain and in the deeps are formed chiefly of :

(1) The remains of animals and plants that float on the surface of the water.

(2) Volcanic material.

If, however, at any part of the continental shelf or slope there is a deficiency in the supply of the normal material, the pelagic type of deposit will appear.

Deposits of the Continental Shelf and Slope

Material derived from the wearing of the land. By far the greater part of the deposits on the continental shelf and slope consists of material brought down from the land by rivers or worn from its edge by the sea. The larger fragments are laid down close to the shore, for it is only in shallow water that the waves and currents have power to move them; but the finer particles may be borne a considerable distance out to sea, and the more minute they are the farther they can be carried. Thus, to a certain extent, the material is sorted according to size. The large blocks which have fallen from the cliffs are often too heavy to be moved at all, and they remain where they fell, until they are further broken up. Smaller boulders and pebbles form shingle-banks, usually between tide-marks, where the waves break violently, and being continually washed to and fro they are worn smooth and round and are gradually reduced in size. Sand is more easily transported and accordingly extends beyond low water mark, often, indeed, to the edge of the continental shelf. The finest material, which may be grouped under the general name of mud, is carried still farther and covers a large part of the shelf and of the slope beyond.

Thus from the shore outwards there is a gradual decrease in the coarseness of the deposit. But the distance to which it travels is not determined entirely by the size of the fragments; it depends also on the strength of the waves and currents, and since this is variable, the sorting of the material is not complete. Only a very small proportion, however, even of the finest particles, is ever carried beyond the continental slope.

It often happens that the strongest currents run parallel to the coast. In that case much of the material derived from the land is not carried outwards but is drifted along the shore. Therefore, the sand and mud brought down by a river is not always deposited at the river's mouth; it may travel along the coast for miles and find a resting place in some quiet bay. It is on account of this lateral drifting of the land-derived material that there is no very striking increase in the width of the terrigenous deposit opposite the mouths of rivers.

The *sand* that is derived from the wearing of the land may contain fragments of any of the rocks or minerals which help to form the land. But some of these are soft and are quickly ground to powder, producing mud instead of sand; others are easily decomposed and the product of the decomposition is usually friable; others again occur only in small quantities or are only locally abundant, and it is only here and there that they form any considerable proportion of the sand. Thus it comes about that by far the greater part of an ordinary sand is made of grains of quartz; for quartz is one of the most abundant constituents of the earth's crust, one of the hardest and most resistant, and one of those least liable to chemical change.

The *muds* are of finer texture than the sands. They consist to a large extent of minute particles of various rock-forming minerals, quartz being again the most abundant; but there is also, as a rule, a considerable quantity of impalpable clayey matter, the proportion of which increases with the distance from the land.

Blue mud is by far the most widely spread type, but Red mud is found in the Yellow Sea and off the coast of Brazil. In both the colour is due to compounds of iron, but in the Red mud the iron is in a more highly oxidised condition.

Green mud is found in some localities, especially on the continental slope off rocky coasts where no large rivers enter the sea. The colour is due to grains of a mineral called glauconite. Glauconite is a silicate of iron and it seems to be formed only in the presence of decaying organic matter. It often fills the chambers of the shells of Foraminifera, and when the shell dissolves, the casts of glauconite are left as little rounded grains. Where the accumulation of mud is rapid, the proportion of glauconite is too small to affect its general colour; but where the supply of mud is less the glauconite grains are relatively more abundant. Glauconite may also occur in sands in sufficient quantity to give them a greenish colour. Where warm and cold currents alternately invade each other's territory it is sometimes associated with phosphatic nodules, which also seem to require the presence of organic matter for their formation.

Organic deposits. On many parts of the continental shelf both animals and plants live and grow in multitudes, and their shells and skeletons may form the greater part of the deposit. On our own coasts, for example, there are oyster-banks and mussel-beds; in warmer seas corals and calcareous algæ flourish luxuriantly. The shells and skeletons may be broken up by the waves, forming sands and muds which differ from the ordinary terrigenous deposits in the fact that they consist almost entirely of carbonate of lime.

It is in the West Indian Seas that such organic deposits reach their greatest development. The Bahamas are formed almost entirely of shell sands and coral sands blown up by the wind, and sands and muds of similar constitution cover the bed of the surrounding sea and are also found extensively in the Gulf of Mexico and the Caribbean Sea.

Volcanic deposits. In volcanic regions the deposits of the continental shelf and slope consist chiefly of fragments thrown out during eruptions, and are not derived to so great an extent as usual from the wearing away of the land. There is the same sorting and distribution of the material, but there is a difference in its composition. Volcanic sands, for example, consist of fragments of lava, instead of grains of quartz.

Deposits of the Deep-sea Plain and of the Deeps

With the exception of the finer sorts of volcanic dust very little terrigenous material is carried beyond the foot of the continental slope, and the deep-sea plain is covered for the most part by pelagic deposits. Even on the slope itself, wherever the supply of terrigenous mud is deficient the deposit becomes more or less pelagic in type.

The pelagic deposits consist in part of the remains of animals and plants which float on the surface of the sea and in part of volcanic dust brought by the wind. On account of the low temperature of the water and for other reasons also the number of animals and plants which actually live at the bottom of the deeper parts of the ocean is comparatively small, and their remains help but little in the formation of the deposit.

When first brought up to the surface most of the pelagic deposits are in the form of a liquid mud, which is commonly spoken of as "ooze." When dried the ooze becomes a fine-grained and powdery mass, partly amorphous and partly made of little shells, sometimes visible to the naked eye and sometimes of microscopic size.

These shells belong to several different kinds of organisms, and the ooze is named after the predominant type. In some the shell is made of carbonate of lime, in others of silica, and the ooze accordingly may be either calcareous or siliceous.

Besides the organic oozes, formed of the remains of animals and plants, there is another type of deposit, known as Red clay, which consists mainly of inorganic material and which is apparently of volcanic origin.

The pelagic deposits may therefore be classified as follows :

Organic	Calcareous	Pteropod ooze
		Globigerina ooze
	Siliceous	Radiolarian ooze
		Diatom ooze
Inorganic		Red Clay

Pteropod ooze. In this the shells of a certain class of floating molluscs, known as pteropods, form the most conspicuous constituent. These shells are always thin and fragile and are often more or less conical in shape. They may be a quarter of an inch or half an inch in length and they are always formed of carbonate of lime.

Pteropod ooze occurs chiefly on the ridges and plateaux which rise from the deep-sea plain, where the water is comparatively shallow but at the same time distant from any continental mass. It is found, for example, at several places on the mid-Atlantic ridge. It is most typically developed at depths of about 800 or 1000 fathoms, but is often met with in shallower waters and it may extend downwards to about 1800 fathoms.

Pteropods flourish most abundantly where the surface water is warm and the annual range of temperature small, and the deposit is therefore found mostly within or near the tropics. But it is not very widely spread and it is principally in the Atlantic Ocean that it occurs.

Globigerina ooze. Globigerina ooze is made up chiefly of the calcareous shells of Foraminifera, Globigerina being the most abundant and widely distributed genus. These shells are always small, commonly about the size of the head of an ordinary pin, but often smaller and sometimes larger.

This is the most wide-spread type of ooze in the Atlantic and Indian Oceans and covers also a large area of the South Pacific. In general it is a deposit of either warm or temperate seas, but between Greenland and Norway it spreads beyond the Arctic circle. Here, however, the temperature of the surface waters is far above the average for the latitude.

It is at depths ranging from 1500 to 2000 fathoms that Globigerina ooze is most abundantly and most characteristically developed. But where it is not masked by too great an accumulation of terrigenous material it may extend into waters much shallower than this. Its lower limit varies. As the depth increases it gradually disappears, giving place to Red clay. Occasionally it has been found even below 3000 fathoms, but it does not occur in the actual deeps.

Diatom ooze. Diatoms belong to the vegetable kingdom and are in general of microscopic size. Their skeletons or "frustules" are made of silica. They flourish principally in the colder seas.

This type of ooze is found chiefly at depths varying from 600 to 2000 fathoms, but it may extend even to 4000 fathoms. It forms a broad belt in the Southern Sea outside the terrigenous deposits of the Antarctic continent, and a narrower band on the northern border of the Pacific Ocean.

Radiolarian ooze. The Radiolaria are minute organisms belonging to the same great division of the animal kingdom as the Foraminifera. Their shells or skeletons, however, are made of silica instead of carbonate of lime, and are characterised by their remarkable openwork structure, forming a kind of lattice supporting the body of the animal rather than enclosing it.

Radiolarian ooze occurs only in deep waters, and is seldom found at depths less than 2000 fathoms. It extends downwards to 5000 fathoms or more and is accordingly met with in the deeps as well as on the deep-sea plain.

It is confined to tropical seas and occurs in the Pacific and the Indian Oceans, but is not known in the Atlantic.

Distribution of the organic oozes. The distribution of the organic oozes is governed partly by the depth of the ocean, partly by the temperature of the surface water. The calcareous oozes are never found in the deepest parts of the ocean. At these great depths the carbonate of lime is dissolved, either in falling or soon after it reaches the bottom. Consequently, although Pteropods and Globigerina live at the surface and are not affected by the depth of the water beneath, their shells form no deposit in the deeps. The shells of Pteropods are thinner and more easily dissolved than those of Globigerina, and therefore Pteropod ooze does not extend downwards so far as Globigerina ooze. Pteropods, moreover, or at least the shell-bearing Pteropods, do not flourish to any great extent except in water that is warm, while Globigerina lives also in colder seas. For this reason Pteropod ooze is limited to tropical and sub-tropical regions, while Globigerina ooze spreads over both tropical and temperate seas.

Siliceous material is not so easily dissolved and the siliceous oozes therefore extend to greater depths than the calcareous deposits. But they can only occur where the organisms that form them live in abundance. Radiolaria flourish principally in the warmer seas and diatoms in the colder waters. Consequently Radiolarian ooze occurs within the tropics, Diatom ooze within and near the polar seas. Even in warm waters, however, Radiolaria are generally less abundant than Foraminifera, and it is accordingly only in deep water, where the foraminiferal shells are dissolved, that the proportion of

Radiolarian remains rises sufficiently high to form a true Radiolarian ooze.

Red clay. Red clay is the most widely spread of all the pelagic deposits. In composition it is a true clay, consisting mainly of hydrated silicate of aluminium, coloured by oxide of iron. It is not found above 2000 fathoms but it is the characteristic deposit of the deeper parts of the ocean. It covers more than half of the Pacific Ocean, and also large areas in the Indian and Atlantic Oceans, extending downwards to the greatest depths observed.

The Red clay is apparently formed by the decomposition of the volcanic material which is carried out to sea, either as pumice floating on the waves or as dust blown by the winds. Such material falls all over the ocean floor, but excepting near volcanic regions the supply is small and the rate of accumulation is extremely slow.

The calcareous oozes are deposited much more rapidly, and where the water is not too deep they conceal the comparatively small amount of volcanic material that falls with them. If, however, Globigerina or Pteropod ooze is treated with dilute hydrochloric acid, the carbonate of lime is dissolved and the material left is very like Red clay; the volcanic matter is there, but it is small in amount compared with the carbonate of lime. At great depths in the ocean a similar process goes on; the carbonate of lime is dissolved and nothing but Red clay is left. At intermediate depths there is a gradual passage from Globigerina ooze to Red clay.

Since silica is not so easily dissolved, the siliceous oozes may extend even into the deeps, and Radiolarian ooze has been found beyond 5000 fathoms. Whether the deposit formed is a Radiolarian ooze or a Red clay, depends not so much on the depth of the water but rather on the abundance of Radiolaria at the surface.

The extreme slowness of the deposition of Red clay in mid-ocean is shown by the fact that sharks' teeth, and ear-bones of whales, sometimes of extinct species, are frequently brought up in dredgings and soundings in the Red clay areas. They have lain so long that the remainder of the skeletons has been dissolved, and in the case of the extinct species they must have been there for many thousands of years; yet, at the most, they have been barely covered by the deposit. Moreover, spherules of iron and other minerals, like the dust that sometimes falls upon the earth in meteoric showers, have also been found. There is no reason to suppose that this dust falls more abundantly on the ocean than on the land. But on the land it is lost in the mass of other material, in the Red clay area it is buried very slowly and forms a much larger proportion of the whole deposit.

CHAPTER XV

CORAL REEFS AND ISLANDS

In many parts of the tropical seas coral grows in such profusion that it forms rocky reefs, often of great size, rising up to the surface of the water. Such reefs may fringe the shores of land that is not made of coral or they may form islands far removed from any other kind of land. Excepting where earth-movements have taken place it is seldom that a coral island rises more than ten or twenty feet above the level of high tide, and even this height is reached only by the heaps of broken coral and coral sand thrown by the waves upon the reef. The living coral does not grow above the water, scarcely indeed above low-water mark, and it is only where there has been elevation of the land that the reef itself stands up above the waves.

Coral reefs and islands are not made entirely of corals Other organisms play a very considerable part in their formation. Calcareous algæ are often at least as important as the corals themselves, and much of the deposit consists of the shells of Foraminifera.

The coral is formed by animals which are very like the sea anemones of our own coast. Some of them live apart, each on its own little cup of coral; but in the true reef-builders the bodies are attached to one another in groups or colonies. The coral is deposited at their bases and forms a firm and stony seat on which they rest and from which they never stir. It is composed of almost pure carbonate of lime.

Distribution of coral reefs. Coral reefs belong to the warmer seas and are almost entirely confined to the zone between latitude 30° N. and 30° S. In the Bermudas in latitude 32° N. there are reefs which are partly made of coral, but calcareous algæ and other organisms take a larger share than usual in their formation.

Even within the zone from 30° N. to 30° S. true coral reefs are unknown on the western shores of the continents. Masses of coral may occur, but they do not form typical reefs on the western coast of either America, Africa or Australia, although on the eastern sides of all these continents there are regions where reefs abound. It is the zone of the trade-winds and the western side is the leeward side, where there is an upwelling of cold water. Probably this is the principal cause of the absence of reefs; but it can scarcely be the sole one, for near the equator, even on these shores, the surface water is warm enough for corals.

Coral reefs are especially numerous in the Pacific and Indian Oceans. In the Atlantic they are abundant in the West Indian Seas and are also found off the coast of Brazil, but the only reefs in mid-Atlantic are those of the Bermudas.

The largest of all coral reefs is the Great Barrier Reef of Australia, which extends along the coast of Queensland for more than a thousand miles, with its outer edge at a distance from the mainland varying from 20 to 150 miles.

There are corals even in the northern seas and in deep waters, but they do not there form masses of any great size. The real reef-builders require a surface temperature that does not fall more than a degree or two below 70° F., and they do not grow freely at greater depths than 30 fathoms, though scattered colonies may spread downwards to 50 fathoms. Moreover, they are quickly killed by any deposition of sediment, and therefore they are found only where the sea is clear, and never where a river brings down mud from the land.

It is easy, accordingly, to understand why coral reefs are practically confined to the zone between 30° N. and 30° S., and why, within that zone, they favour the eastern shores of the land-masses. But there are thousands of coral reefs in mid-ocean, rising, to all appearances, directly from the floor of the deep sea, and it is not easy to prove how the foundations of these reefs were built, at depths far below those in which the corals can live. This is a problem concerning which there are still great differences of opinion.

Structure of coral reefs. Three kinds of coral reef are generally recognised as more or less distinct, viz., fringing reefs, barrier reefs and atolls.

A fringing reef is one that lies close to the shore of some continent or island. Its surface forms a rough and uneven platform round the coast, about the level of low water, and its outer edge slopes downwards into the sea. Between the coral platform and the land there is sometimes a shallow channel or lagoon which is filled with water even at low tide (Fig. 78).

FIG. 78. Fringing reef.

When the lagoon is wide and deep and the reef lies at a distance from the shore and rises from deep water it is called a barrier reef (Fig. 79).

FIG. 79. Barrier reef.

An atoll is a reef in the form of a ring or horse-shoe with a lagoon in the centre (Fig. 80). Sometimes there is a small

FIG. 80. Atoll.

island in the middle of the lagoon and the reef is really a barrier reef around it. But in the true atoll there is no central island, only a ring-shaped reef enclosing a lagoon.

There are reefs which are never uncovered even at the lowest tides, and their jagged crests, hidden beneath the waves, are a serious danger to passing ships. But as a rule, when circumstances are favourable, a coral reef grows both upwards and outwards until it reaches low-water level. Its upward growth then ceases, but it may still continue to spread outwards. Therefore, whether it is a fringing reef, a barrier reef or an atoll, the top of the reef is a rugged and uneven flat which at low tide is just awash. It is traversed by fissures, especially near its margin, through which the sea advances and retreats as the waves break over its outer edge. The outer edge itself is often slightly raised, forming a low rim, which is made of calcareous algæ rather than of coral.

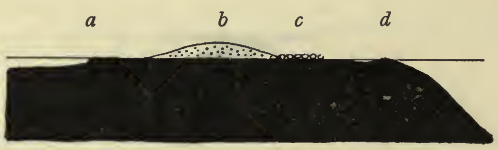


FIG. 81. Section of a reef-flat.
a, lagoon; *b*, mound of sand; *c*, boulder zone; *d*, outer edge of reef.

The seaward face of the reef is steep, but at the top it is usually rounded or bevelled, so that for some 20 or 30 fathoms the slope is moderate (Fig. 81). Perhaps the growth of coral is most luxuriant some distance beneath the surface, and in any case the deeper colonies are less liable to be torn off by the waves. At greater depths the angle of slope may be 40 or 50 degrees or even more.

Towards the lagoon the flat may slope gently beneath the water, or it may end in a sort of step of no great depth. But often the shore of the lagoon is sandy and the reef itself is completely concealed. In the lagoon the water usually deepens very slowly except at the step already mentioned, and even in the deepest lagoons soundings of more than 50 fathoms are scarcely known.

On the reef-flat the coral is mostly dead. In the lagoon there may be living corals, and calcareous algæ and Foraminifera are abundant. But the reef-building types of coral do not thrive except on the seaward face of the reef, where no sediment can settle and the supply of food is ample. The growth of a reef is therefore mainly outwards. In some cases the lagoon appears to be filling up, in others it is deepening.

Coral islands. Some little distance within the outer edge of the reef-flat there is a zone of boulders, consisting of masses of broken coral thrown up by the sea and sometimes standing above the water even at high tide. If the reef is narrow the boulder zone may border the lagoon; the smaller fragments, forming coral sand, find no resting-place upon the flat, excepting in the hollows, and are swept across into the lagoon or backwards into the sea. But if the reef is wide, the waves are unable to wash the sand across and it collects in mounds behind the boulder zone, forming islands which rest upon the coral flat. Sand from the lagoon may also be added to these mounds, if the direction of the wind is favourable. Foraminifera often form a large proportion of the sand, in addition to the fragments of coral and calcareous algæ. The whole is easily consolidated by percolating water, which dissolves and re-deposits carbonate of lime, cementing the particles together.

Many coral islands are nothing more than such mounds of sand, resting upon a reef-flat. In that case they may be several miles in length and may even be of considerable breadth, but they are always low. When, however, there has been elevation of the land, the reef itself may be raised above the sea, and the island may attain an altitude of several hundreds of feet.

In course of time seeds brought by winds and currents, or carried by birds, will establish themselves; and may clothe the island with a luxuriant growth of vegetation almost to the water's edge. The low-lying islands covered with trees which from a little distance seem to grow upon the water itself, form one of the most striking features of the coral seas.

Mode of formation of coral reefs. A fringing reef is formed by corals, with the help of other calcareous organisms, growing upon the floor of the shallow seas around a continent or island

The corals spread outwards from the shore to a depth of about 30 fathoms, but beyond that depth there are practically none. The coral masses grow upwards until they reach low-water level, and thus a platform is formed, which ends within the position of the original 30 fathom line, and the edge of the platform rises from about that depth. The outer corals grow more freely than those which are closer to the shore, probably because they are more favourably situated for receiving supplies of food from the sea, and because they are less liable to be covered by sediment. Consequently, the outer part of the reef may reach the surface first and a shallow channel or lagoon may be left between the platform and the land.

So much is generally agreed ; but in barrier reefs and atolls the outer side of the reef rises from depths where no reef-building corals live. There are only two explanations possible.

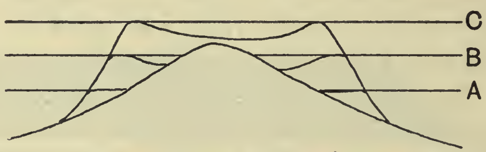


FIG. 82. Darwin's theory of the formation of atolls. *A*, *B*, *C* represent the level of the sea at successive stages in the growth of the reef.

Either the foundations of such reefs were not built by corals, or else the sea has deepened since the corals built them. Each of these explanations has its advocates ; and it is probable that both are true, that some reefs have been formed in the one way and some in the other.

Darwin's theory. The theory of subsidence was first proposed by Darwin. He imagined that all reefs began as fringing reefs around a continent or island (Fig. 82). As long as the relative level of land and sea remains unchanged, the reef will continue as a fringing reef. But if the land and sea floor subside[1] so that the reef sinks beneath the water, the

[1] The same effects will be produced by a rise in the level of the sea. It is very difficult and often impossible to distinguish between sinking of the land and rising of the sea, and in the following account the term subsidence must be understood to apply to both.

corals will again begin to grow upwards. The growth, as already explained, is most vigorous at the outer edge of the reef, while nearer to the shore it is extremely slow. Accordingly, it will often happen that only the outer part of the reef can keep pace with the subsidence, and the inner part is flooded by the sea, thus forming a lagoon between the reef platform and the land. As the subsidence goes on, the width and depth of the lagoon increase and the reef becomes a barrier reef. If the reef is round an island the island may at length be entirely submerged, and nothing will be left but a ring of reef enclosing a lagoon.

According to this view, barrier reefs and atolls can only occur where there has been subsidence, and every atoll marks the position of a former island. Moreover, the sea outside an atoll will usually be deep, for the subsidence must have been sufficient to submerge the central island.

Murray's theory. If atolls and barrier reefs rising from deep water do not owe their formation to subsidence in the manner described by Darwin, the only alternative is that their bases were not built by living coral; and this is the view supported by Agassiz, Murray and others.

Sir John Murray points out that in course of time a fringing reef may spread beyond the original 30 fathom line; for after this limit is reached fragments of coral, both large and small, will accumulate at the foot of the reef, and upon the pile of talus the corals will continue to grow outwards. In this way, without any subsidence the reef may extend into deep water; but only the upper 30 fathoms will be built by living coral, in deeper water the foundations will consist of the debris from the reef, cemented together by the action of the water.

As the reef grows outwards the inner corals die, and Murray believes that the dead coral is slowly dissolved. Thus a lagoon is hollowed out between the rim of the reef and the shore, and the fringing reef is gradually converted into a barrier reef.

Atolls, according to Murray's view, are formed on the tops of plateaux and hills which rise from the sea floor to depths at which reef-building corals live. Colonies of corals will establish themselves and grow both upwards and outwards, many colonies perhaps uniting into a single mass or reef. In any such mass the outer corals will grow most freely and will reach the surface first, forming a ring or atoll enclosing a lagoon. When the reef has reached this stage, or even at an earlier period, the living coral is practically confined to its seaward face and the interior of the reef is dead. The lagoon will be enlarged and deepened by solution of the dead coral, while the outward growth will still go on. The ring of coral will open out more widely and the size of the atoll will increase with age.

Murray's theory requires a very large number of submarine hills or plateaux, all of them reaching to about the same level; their tops must all have been some 15 to 30 fathoms below the surface of the water. Such a remarkable coincidence in height needs explanation, and Murray suggests two ways in which it may have been brought about, even if the original heights were different.

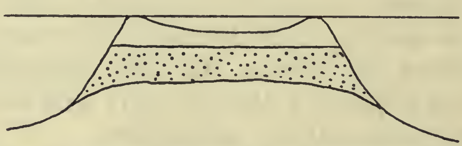


FIG. 83. Murray's theory of the formation of atolls. The dotted portion represents pelagic and other deposits, reaching upwards to the level of coral growth.

Most islands in mid-ocean, except the coral islands themselves, are of volcanic origin, and he believes that the elevations on the ocean floor were mainly volcanic. Many of them rose above the sea, and many still stand above the waves; but since volcanoes are often nothing more than piles of fragments thrown out during eruptions, many would be quickly worn away by the sea, till only a shoal was left to mark the site. This has actually happened in the case of one or two new volcanic islands which have appeared and disappeared during the last hundred years.

Other hills may not originally have reached the 30 fathom line. Pelagic deposits would accumulate upon them, more quickly than in the deeper sea around because there would be less solution of the carbonate of lime. In time their summits would be raised to the level at which shells and deep-sea corals live; and finally they would reach the depth where reef-building corals can begin to grow (Fig. 83).

In these two ways, by the wearing away of hills which stood above the sea and by deposition upon those which lay too deep beneath the surface, a sufficient number of elevations may be produced on which reef-building corals can live.

Discussion of the evidence. In the limited space that is here available it is impossible to do more than indicate a few of the arguments that have been advanced in support of these rival theories.

The evidence is strong that some atolls and barrier reefs have been formed in areas of subsidence. The coast of Queensland within the Great Barrier Reef shows the drowned valleys and outlying islands characteristic of a sinking land. In many ring-shaped reefs with a central island, the island has a similar indented coast-line, and it is evident that a little further depression would convert the reef into an atoll.

It is equally clear that other atolls have been formed without subsidence. Darwin himself recognised that in shallow seas the natural outward growth of coral masses would lead to the formation of ring-shaped reefs, and he thought that the atolls and barrier-like reefs of the West Indian Seas grew in this fashion upon banks of sediment accumulated by currents. But he distinguished between these and the deep-rooted reefs of the Indian and Pacific Oceans, which he believed to have a solid foundation of coral.

Even in the Pacific, however, many atolls are found in areas where there is no evidence of subsidence, and where, on the contrary, there is definite proof of elevation. In the Pelew Islands, and in many other island groups of the Pacific, there are coral reefs which have been raised above the sea, pointing clearly to elevation of the land; and yet in the same groups there may be atolls. Moreover, atolls are sometimes found in water of no great depth, rising from a plateau and not apparently from sunken peaks.

According to Darwin's view barrier reefs and the reefs of atolls must be of very considerable depth, often 2000 feet or more. According to Murray's view only the upper 30 fathoms (or less than 200 feet) of such reefs is made of coral which has grown in place. It has been observed that in many raised reefs the thickness of coral is not more than 200 feet, and it has even been maintained, though on very insufficient evidence, that this thickness is scarcely ever exceeded. There are several examples of atolls which have been raised above the sea, and in those which have been examined the depth of the coral deposits is less than 200 feet. Santa Anna, in the Solomon Islands, and probably Christmas Island, south of Java, are instances.

On Murray's hypothesis the steepness of the external slope of many reefs is somewhat difficult to explain. It is easy to understand that a mass built up solidly of growing coral may have almost perpendicular walls, but according to him it is only the upper 30 fathoms of a reef that are formed in this way. The lower part consists of pelagic deposits, the shells and skeletons of animals and plants which live on the floor of the sea below 30 fathoms, and the talus of the reef. It is, in fact, an accumulation of fragments ; and it is difficult to believe that any such accumulation can have an external slope even approximating to 75°. Yet slopes as steep as this or even steeper are met with on the seaward faces of coral reefs at depths far greater than 30 fathoms. It is possible, however, that calcareous algæ or other sedentary organisms might form a solid deposit,—that, in fact, reef-builders other than corals may grow in deep waters.

The depth of the lagoon in many reefs is also not very satisfactorily accounted for by Murray's explanation. The reef is supposed to be built upon a platform which has been lowered by erosion or raised by deposition to a depth of 30 fathoms. Therefore the natural depth of the lagoon should be less than 30 fathoms ; but it frequently exceeds 40 fathoms. Murray ascribes the deepening to solution of dead coral by the water of the sea ; but in order that the depth may reach 40 fathoms, not only must the whole thickness of coral in the lagoon have been removed, but also 10 fathoms of the foundation on which the corals built. This foundation was in general either volcanic or formed by deposition in the sea, and there seems to be no reason why it should be dissolved more easily by the water of the lagoon than by that of the open ocean. It may be argued that deposition went on more rapidly in the open sea than in the lagoon, and so exceeded the rate of solution; but the evidence on this point is somewhat conflicting. Observations indicate that in some atolls the lagoon is filling up, in others it may be growing deeper.

Of recent years the destruction of dead coral, which undoubtedly goes on, has been attributed largely to boring animals and plants, and the hollowing of the lagoon to the action of currents as well as of solution.

Funafuti. In the hope of finally settling the question it was determined to put down a boring through the rim of a well-marked atoll. According to Murray's view only the upper 30 fathoms should consist of coral, and this should rest either directly upon a foundation of volcanic rock, upon a talus of broken coral or upon an accumulation of pelagic and other organic remains of a different type from the reef-building corals. According to Darwin's theory, the corals and other reef-building organisms should extend far below 30 fathoms and should rest directly upon the flanks of the peak round which the atoll was formed.

The atoll of Funafuti in the Ellice group in the South Pacific was selected, and after several attempts a boring was put down to a depth of 1114½ feet. Much of the material passed through, especially in the upper part of the boring, was not sufficiently consolidated to withstand the action of the borer and came up in fragments. About one-third, however, was firm enough to yield a solid core. The detailed examination of the material showed that the whole was made up mainly of Foraminifera, corals, and calcareous algæ, and the same types extended from the top to the bottom. Reef-building corals were found throughout. Even in the lower part of the boring they appeared to be in the position of growth, but this conclusion is open to doubt owing to the imperfect state of preservation of the coral remains. So far as the evidence goes, therefore, it is decidedly in favour of Darwin's theory. It is possible that the lower part of the reef may be made of talus, but there is nothing to support this view; and there were no traces of any other kind of deposit, such as Murray's theory demands.

Two borings were put down in the lagoon, where the depth of the water at low tide was 101 feet. One of the borings reached 113 feet and the other 144 feet below the floor of the lagoon, and at these depths they met solid limestone too hard to be penetrated by the apparatus used. In each the upper 70 feet was made chiefly of fragments of calcareous algæ and was similar in general character to the deposit now forming in the lagoon.

General conclusions. Owing to the conflicting nature of the evidence it is difficult to draw any general conclusions, excepting that atolls and barrier reefs may be formed in various ways. Some have grown in areas of subsidence, others in regions where no depression has taken place.

There is no proof that the majority of atolls surround submerged islands. On any extensive shoal a number of ring-shaped reefs may be formed by the outward growth of coral masses; and if the shoal is slowly depressed each of these may continue to grow upwards, finally becoming an atoll rising from the floor of the deep sea.

There is but little evidence that the accumulation of pelagic deposits has done much to raise the summits of submarine elevations to the level of coral growth. Globigerina limestones and similar deposits have been found at the base of raised coral reefs; and in the Solomon Islands a reef has been described resting upon a material very like the Red clay of the deeps. But this is evidence of elevation of the sea floor by earth-movement rather than by deposition. It can hardly be maintained that any deposition of oceanic Red clay could raise the floor of the sea to the level at which reef-building corals live.

It may be added that there is no sufficient proof that in lagoons generally the deepening due to solution and erosion is greater than the shallowing due to deposition.

CHAPTER XVI

MATERIALS OF THE EARTH'S CRUST

In the ordinary sense of the word the term rock implies something which is hard and resistant, but by the geologist the meaning of the word is extended so as to include all the solid material of the earth's crust[1], whether it is hard like granite or soft like clay.

There are many different kinds of rock, but the different kinds are not sharply distinguished from one another. A rock is not a definite chemical compound but is usually a mixture of various minerals. A limestone is made up chiefly of carbonate of lime, and a clay consists mainly of silicate of alumina; but many limestones contain a large proportion of clay, and many clays contain a large proportion of carbonate of lime.

Rocks may be divided into two main classes according to their mode of origin, and these two classes are known respectively as Igneous and Sedimentary. A third class may be added consisting of rocks which have been so greatly altered, by heat or pressure or by both combined, that their original characters are completely lost. Such rocks are known as Metamorphic.

The igneous rocks have at one time been molten and have solidified from the molten condition. In many of them the various minerals have crystallised separately and the rock is a mass of large or small crystals interlocking with one another. Such rocks are called crystalline.

The sedimentary rocks have been laid down by rivers or other agents, or they are formed of the shells and skeletons of animals or plants. They consist for the most part of fragments, such as grains of sand or pieces of shell, usually bound together

[1] The term "earth's crust" requires a little explanation. It appears to have originated when it was believed that the interior of the earth was molten, and it was applied to the outer solid crust resting upon the liquid interior. There is now a considerable amount of evidence that the interior of the earth is not liquid; but its condition must be very different from that of the exterior and the term earth's crust is still employed to denote the outer part of the solid earth, in which the rocks are more or less similar to those at the surface.

by some cementing material. Generally, owing to their mode of formation, they are deposited in layers known as beds or strata, and hence are often spoken of as "stratified."

Igneous rocks. Igneous rocks have at one time been molten in the interior of the earth. Sometimes the molten material has been poured out upon the surface of the earth, as in volcanoes. Sometimes it has solidified deep down beneath the surface, sometimes in the channels which connected the molten reservoirs with the exterior.

In the interior the cooling is necessarily slow, allowing time for the crystallisation of the different minerals, and a rock which has solidified deep down in the earth is always completely

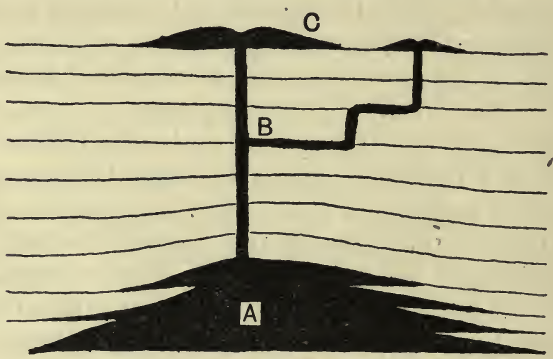


FIG. 84. Igneous rocks.
A. Plutonic rocks. *B*. Dyke rocks. *C*. Volcanic rocks, or lava.

crystalline. Such rocks are called plutonic rocks, and granite is a good example (Fig. 84).

The molten material sometimes finds its way to the surface through clefts, or it may melt a passage for itself; and it may solidify on the way. It will then form vertical walls or dykes cutting through the beds; or more or less horizontal sheets, or cylindrical necks. In these the cooling is more rapid than in the great masses below, and the rocks are usually less completely crystalline than the plutonic rocks. They are often spoken of as dyke rocks.

When the molten rock escapes upon the surface it is known as lava. A stream of lava may flow for miles, but naturally it will cool and solidify more quickly than in the interior of the earth. The different minerals composing it have less time to crystallise. Many lavas, indeed, are not crystalline at all, but resemble glass. Obsidian is a good example. In other lavas the crystals are too small to be visible to the unaided eye ; but sometimes they are of moderate size.

Sedimentary rocks. The sedimentary rocks are commonly divided into four classes according to their composition, as follows :

Arenaceous rocks, *e.g.* sandstone, grit.
Argillaceous rocks, *e.g.* clay, shale.
Calcareous rocks, *e.g.* limestone.
Carbonaceous rocks, *e.g.* coal, lignite.

A rock which is made of pebbles of other rock, like a consolidated gravel, is called a conglomerate. A rock which is made of angular fragments of considerable size is called a breccia. Such rocks hardly come into any of the four classes given above.

Arenaceous, or sandy, rocks are usually composed chiefly of grains of quartz ; but fragments of other minerals are commonly found in them, and may even form the bulk of the material.

Argillaceous, or clayey, rocks are typically made of clay, a hydrated silicate of alumina. But very fine particles of other minerals may form a deposit which in ordinary language would be called a mud or clay. There are, for example, calcareous muds, consisting chiefly of minute particles of carbonate of lime.

Calcareous rocks or limestones, consist mainly of the shells or skeletons of animals or plants, and are formed of carbonate of lime. Deposits of carbonate of lime are sometimes produced when water containing a large proportion of that mineral evaporates, leaving a film of carbonate of lime where it stood or flowed. It is in this way that petrifying springs apparently convert objects into stone. They cover them with layer after layer of carbonate of lime. Deposits of carbonate of lime formed in this manner are called travertine or calcareous tufa.

Carbonaceous rocks usually consist of the remains of plants, converted, apparently by heat and pressure, into coal. Lignite is a similar kind of deposit in which the change has not been so complete.

Metamorphic rocks. In some regions, owing to heat, pressure, or the movements to which they have been subjected, both igneous and sedimentary rocks may be so greatly altered that they entirely lose their original character and appearance. They are then said to be metamorphic. A metamorphic rock is usually crystalline, but differs from an ordinary igneous rock in the fact that the different minerals are generally arranged

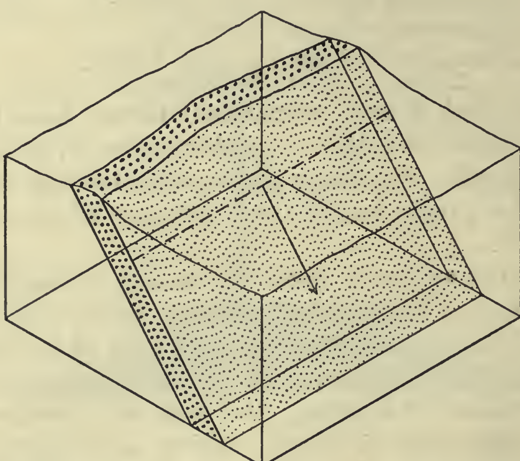


FIG. 85. Dip, strike and outcrop.

The dip is shown by the arrow; the strike by the broken line; the outcrop is the exposed edge of the bed on the upper surface of the block.

in layers. Metamorphic rocks often form the central axis of a great mountain chain.

Folding and faulting. When first deposited the stratified rocks are usually laid down in almost horizontal beds or strata; but subsequently, owing to movements of the earth's crust, they are often tilted out of their original position.

When a bed is not horizontal but has been inclined, the direction of maximum slope is called the dip, and the angle which the dip makes with the horizontal is called the angle of dip.

S. H. Reynolds, photo.

Fig. 1. Bedding and Joints.

Limestone and shale of the Lower Lias, Penarth.

S. H. Reynolds, photo.

Fig. 2. Bedding and Joints.

The Schrammstein, Cretaceous sandstone, Saxon Switzerland.

A line on the surface of the bed at right angles to the dip is called the strike (Fig. 85).

If a bed is inclined and the ground does not slope at the same angle, the edge of the bed will come out on the surface in a more or less regular band, which is called the outcrop.

Often the beds, which were originally horizontal, are bent into a series of folds, as in Fig. 86. The arches of the folds are called anticlines and the troughs synclines.

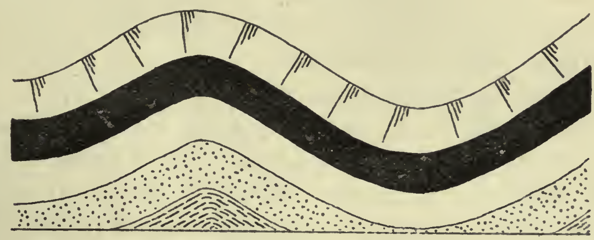


FIG. 86. Anticline and syncline.

Sometimes the strata break instead of folding, and on one side of the break the beds are dropped, relatively to those on the other side. Such a fracture is called a fault (Fig. 87).

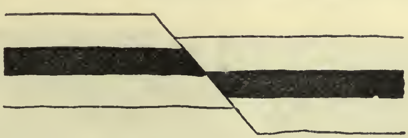


FIG. 87. Fault.

When the fault is nearly horizontal and the beds above it have been pushed forward over those below, the fault is called a thrust-plane (Fig. 96, p. 203).

Joints. Most rocks, whether igneous or sedimentary, break more easily in some directions than in others. Many kinds of stratified rock split readily along the bedding, that is to say, between the layers or beds of which they are composed; but usually they will also break fairly easily along two sets of planes which are at right angles to the bedding. These planes of weakness are known as joints and exert a considerable influence upon the forms of crags and cliffs where the rock is exposed to the weather (Pl. V, figs. 1, 2). Most commonly one set of joints is parallel to the dip of the beds and another set is parallel to the strike.

In igneous rocks the form of the joints varies. Basalt, as in the Giant's Causeway, often breaks into a series of hexagonal columns, and there are also often transverse joints across the columns (Pl. VI, fig. 2). In granites the joints frequently make the rock break into rectangular blocks, which on exposure to the weather become rounded, producing the characteristic forms of the tors of Devonshire and Cornwall (Pl. VI, fig. 1).

Cleavage. The presence of joints will cause a mass of rock to break into blocks of various shapes and sizes. But some kinds of rock, which have been subjected to great pressure, also split or "cleave" easily into thin sheets or slabs. Occasionally the direction of cleavage is parallel to the stratification, but more often it is inclined or at right angles to the bedding. Cleavage, like jointing, is due to planes of weakness in the rock; but cleavage-planes are different in their nature and origin from joint-planes. Ordinary roofing slate is a rock in which the cleavage is well developed. Such rocks have often lost all tendency to split along the bedding.

CHAPTER XVII

EARTH MOVEMENTS

Elevation and Subsidence

Changes of level. On many coasts there are legends of invasions by the sea and stories of ancient towns which now lie beneath the waters. When these legends have any real foundation, the invasion has usually been due to the wearing away of the land by the waves, but in some cases there is evidence of an actual change of level. Either the land has sunk or the sea has risen.

Occasionally, on the other hand, a town that was once upon the coast is now some distance from the sea and the land has

S. H. Reynolds, photo.

Fig. 1. Joints in granite.
Near Lustleigh, Dartmoor.

S. H. Reynolds, photo.

Fig. 2. Joints in basalt.
Giant's Causeway, Antrim.

evidently grown outwards. Generally this is due to deposition of material in the sea, but sometimes it is the effect of an actual elevation of the land relatively to the sea.

Changes of level are usually attributed to upward or downward movements of the land, but an alteration in the level of the sea would produce a similar result. There is, however, one important difference. The surface of the sea is not absolutely level; on the continental coasts it is somewhat raised by the attraction of the land-masses, and this attraction is not everywhere the same. An alteration in the level of the sea need not, therefore, be world-wide; but it must affect a very large area simultaneously, and for long distances the change of level will be practically uniform throughout.

Movement of the land, on the other hand, is not likely to be uniform. Some parts will rise or fall more than others, and there may, in fact, be a rise in one place and a fall in another near by.

It is, however, often very difficult to determine to which cause the change of level is due, and on this account some observers speak of an elevation of the land relatively to the sea as a negative movement, and a depression of the land relatively to the sea as a positive movement, whether the cause is an actual movement of the land or an alteration in the level of the sea. In the following pages, however, the terms elevation and subsidence (or depression) will be used with their ordinary significance, meaning elevation and subsidence relatively to the level of the sea.

Changes of level may be either sudden or gradual. Sudden changes take place only during earthquakes. On the sea coast their effect is conspicuous, even if they are small in amount, and they have accordingly often been noticed. Sir Charles Lyell in his *Principles of Geology* describes many examples. In the New Zealand earthquake of 1855, an upheaval amounting to 9 feet was recorded. After the Chilian earthquake of 1822 the coast for a long distance is said to have stood 3 or 4 feet higher than before.

Inland such changes are not so easily observed; but, to mention only a single instance, during the Japanese earthquake of October 28th, 1891, a fracture of the surface occurred and on one side of the fracture the ground sank 20 feet relatively to the other side.

A gradual change is much more difficult to prove. It might be thought an easy matter to make a mark upon a sea-cliff and to measure the height of this above the surface of the water. But on our own coasts, for instance, the level of the sea is continually fluctuating. It varies not only with the tides but also with the direction and strength of the winds, and a long series of observations is necessary to determine the mean sea-level. In the Baltic the difficulties are not so great, on account of the absence of tides and the number of sheltered inlets. In the early part of the eighteenth century Celsius came to the conclusion that the waters of the Baltic were slowly falling; but several objections to his views were advanced by other writers. It was pointed out, for example, that the lower part of the town of Danzig still lay at the level of the sea and had done so since 1000 A.D. The interest aroused by these discussions led to the placing of marks upon the rocks, indicating the level of the water on a calm day, and these marks have since been examined from time to time. The observations have shown conclusively that there has been a change in the relative level of land and sea; but since the change varies from place to place, it has been due to movement of the land rather than of the water. At Stockholm, the land has risen at the rate of about 18½ inches in a century; farther north at twice that rate, and in southern Sweden only about half as fast.

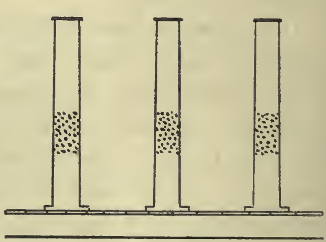


FIG. 88. The Temple of Serapis.

In the Baltic the gradual elevation of the land has thus been shown by actual observation and measurement; but in most cases changes of level can only be proved by inference. One of the most famous examples is that of the Temple of Serapis at Pozzuoli near Naples (Fig. 88). The pavement of the building is now a little below high water and upon it three of the original pillars are still standing. Below the present pavement excavations have shown that there is an older one, at a depth of about 5 feet. The pillars are smooth up to a height of 12 feet, and the next 9 feet are pierced by numerous holes bored by a species of shell-fish which does not live above high-water mark. The upper part of the pillars is free from perforations.

When the first pavement was constructed, the land, presumably, stood at least 5 feet above its present level. When the holes were bored the land had sunk, and the second pavement was submerged to a depth of 21 feet. To account for the absence of holes in the lower part of the pillars, it is supposed that the building had been previously buried to a depth of 12 feet by volcanic ash thrown out during some eruption of Vesuvius.

Thus the remains point to an oscillation of the land of at least 26 feet, from 5 feet above the present level to 21 feet below it.

Geological evidences of elevation. Besides the evidence afforded by the works of man, the sea itself often leaves indications of its former position. But it is important to distinguish between the effect of deposition and the effect of elevation, for either may lead to an apparent retreat of the sea. In Romney Marsh and in the Wash, for example, the land is gaining on the sea, but this is due simply to the deposition of sand, mud and gravel brought down by rivers or swept along the coast by currents. Similarly, owing to the gradual silting up of the head of the Adriatic, the town of Adria, which was a port in ancient times, is now 14 miles from the sea. In such cases as these the new land is always low lying and is formed of mud and sand and other similar deposits.

There are, however, various kinds of evidence which prove that in some parts of the world the sea stood higher or the land lower than it does now.

Above the present beach and beyond the reach of the waves in the greatest storms there is sometimes a kind of terrace covered with sand or gravel like that of the sea shore (Fig. 89). If, as often happens, sea-shells are found within the sand, it is clear that at one time the terrace was the actual beach and the land must since have risen. Raised sea-beaches of this kind are found in many parts of Great Britain (Pl. VII, fig. 1). In Scotland there are sometimes four or five, one above another, at heights varying from 25 to 100 feet. Many of the fiords of Norway show similar terraces up to a height of at least 600 feet.

Behind the present shore-line and above the level of the highest tides there is occasionally a line of cliff, sometimes with caves hollowed out at its base (Pl. VII, figs. 1, 2). At one time the cliff marked the limit of the tides, and its present position is due to elevation. It must not be assumed, however, that every sea-cliff that is not touched by the waves, has been raised. Sometimes it is simply the accumulation of beach material at its base that prevents the sea from reaching it.

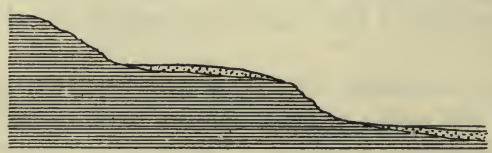


FIG. 89. Raised sea-beach.

The shells of animals such as barnacles, which live in the sea but fix themselves to rocks, are sometimes found attached at heights which are not now reached by the water. In such cases the land must have risen, relatively to the sea. Similar reasoning applies to the various kinds of boring shells such as those in the pillars of the Temple of Serapis.

Reef-building corals cannot stand exposure to the sun and air for more than a few hours. When, therefore, a coral reef is found above the sea, it is clear proof of a change of level. There are many examples of raised coral reefs in the Pacific and Indian Oceans and also in the West Indies.

Elevation of the land is sometimes indicated by the form of the coast. A widespread elevation will raise a part of the continental shelf above the sea, and the land will, accordingly, be fringed by a coastal plain. The outline will be smooth and free from indentations, and behind the coastal plain there will

S. H. Reynolds, photo.

Fig. 1. Raised beach and sea-cliff.
Tormore, Arran.

S. H. Reynolds, photo.

Fig. 2. Old sea-caves.
Cushendall, Antrim.

usually be a sudden rise or even a cliff, marking the former position of the shore-line. This subject will be discussed more fully in a subsequent chapter.

Geological evidences of subsidence. Subsidence is in general more difficult to detect and prove than elevation, because the evidence of the former position of the sea is destroyed or hidden. The old sea-beaches and the old sea-cliffs sink beneath the waves and are lost to sight. It is, moreover, often difficult to distinguish between the effects of subsidence and those of erosion. The sea may cover the site of a former town, but this in itself is no proof of subsidence. It is often due to the wearing away of the land by the waves. This is the case, for example, at Cromer, Dunwich and other places upon the East Anglian coast, where the remains of the old town now lie beneath the water.

Upon our own coasts one of the most plausible arguments in favour of recent subsidence is the presence of "submerged forests" and accumulations of peat or leaf-mould at or below low-water mark. Only the stumps of the trees are left, but in some places these are in the position of growth and afford clear evidence that the forests or woods were invaded by the sea. Submerged forests are found along the shores of Devon and Cornwall and in other parts of Great Britain.

It is, however, possible for peaty and other accumulations of vegetable material to be formed below the level of the sea in a lagoon cut off from the open water. It has also been suggested that the gradual removal of a sandy substratum by water flowing underground, may cause a growing wood to sink beneath the sea. Such a subsidence would be purely local and would not imply a depression of the land as a whole.

According to Darwin's theory of coral reefs, both barrier reefs and atolls are evidence of subsidence; but this theory is not now universally accepted. The question has already been discussed (Chap. XV).

As a rule the most conclusive proof of subsidence is afforded by the form of the coast. Whenever a land-mass sinks, the sea will enter the valleys, forming inlets which will frequently branch inwards. A stream or river will usually flow into the head of each inlet, and each inlet will be the direct continuation of a valley in the land. The indented outline produced in this way is well shown in the west of Scotland, on the Essex coast, and elsewhere. The whole subject of the effects of elevation and subsidence on the form of the coast-line is dealt with in a later chapter.

Nature of Earth Movements

It has been shown in the preceding pages that the surface of the solid earth is not still. Even within the last few hundred years the Scandinavian peninsula has risen perceptibly, the Temple of Serapis has sunk and risen again and similar changes have gone on in other parts of the world. In course of time such movements must produce very great effects upon the continents and oceans; and geology shows that many parts which are now land were once beneath the sea. Many of the rocks which now form part of the land contain the remains of marine animals and were evidently deposited beneath the sea. In the chalk of southern England there are sea-urchins; in the limestone of Derbyshire corals and shells are abundant. In the Alps and the Himalayas beds with marine fossils are found at a height of many thousands of feet.

The stratified rocks which form so much of the surface of the globe were originally laid down horizontally or nearly so, and for the most part beneath the sea. They serve, therefore, as an index of the nature of the movements which have taken place. These movements are of two kinds, which may be distinguished as vertical (or radial) and horizontal (or tangential), according to the principal direction of the movement. In vertical movements there is a simple elevation or depression of a large area of the earth's crust, the beds remaining nearly horizontal. The effect is to form plateaux or to raise a large mass of land above the sea, and this type of movement is accordingly known as plateau-building or continent-building or epeirogenic.

Horizontal movements are due to forces in the earth's crust acting more or less tangentially to the surface. The originally horizontal strata are compressed from side to side, with the result that they crumple or fold. The crumpling takes place where the crust of the earth is weakest, and the effect is to raise up a narrow belt of folded strata, forming a mountain chain. This type of movement is known as mountain-building or orogenic.

Vertical movements. The general characteristic of these movements is that there is no crumpling of the strata, which for the most part remain horizontal; but as will be seen in the accompanying diagrams, the beds may in places be tilted even into a vertical position.

FIG. 90. Vertical movement with gentle bending.

Sometimes a part of the earth's crust is raised in the form of a broad flat arch, or depressed in the form of a wide and shallow basin. The strata will be scarcely disturbed but will have a very slight dip towards or from the centre (Fig. 90). Such a movement may cause the elevation of a continent, or if it produces a depression beneath the ocean, it will lower the level of the sea and cause an apparent rise of the surrounding land.

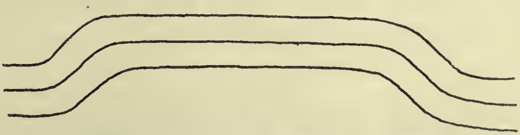


FIG. 91. Vertical movement with local abrupt bending.

Sometimes the boundary between the elevated and the depressed area is very sharply marked, and we have then the form shown in Fig. 91. The strata are in general horizontal, but at the edges of the areas which have been raised or depressed they may be almost vertical.

In many cases the beds have been unable to stand the amount of bending required, and they have accordingly fractured. Large blocks have been raised, and others depressed, relatively to each other, and these blocks are bounded by faults (Fig. 92). Frequently there is a certain amount of

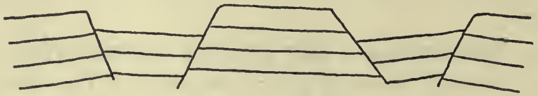


FIG. 92. Vertical movement accompanied by faulting.

tilting of the blocks and the strata are no longer horizontal, but they are not crumpled.

Occasionally a strip of country is let down between two faults and a long and narrow depression is formed (Fig. 93). Such a depression, produced directly by earth movements and not, like most valleys, by erosion, is often called a "rift-valley." The Jordan flows in a valley of this type.

FIG. 93. "Rift-valley."

In all the examples given above the strata were horizontal until the earth movements occurred. But the same type of movement sometimes affects areas which had been previously folded.

Horizontal movements. These are characterised by the crumpling or folding of the strata, but there is great variety in the forms and structures produced.

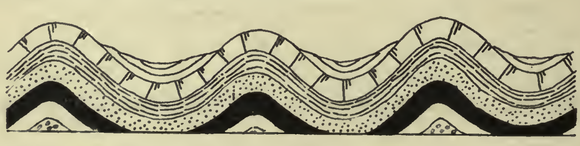


FIG. 94. Folding of the Jura type.

If the amount of compression is small, the beds may be thrown into a series of simple arches and troughs, as in Fig. 94. This is the case in a considerable part of the Jura mountains.

When the compression is greater the centre of the folded belt is raised and the marginal folds lean outwards. The structure becomes that shown in Fig. 95, and because the beds in general dip inwards from both sides, it is called fan-structure. During the process of folding the whole is subject to denudation, and the outer parts of the folds are worn away as they are produced.

FIG. 95. Fan-structure.

Sometimes the lateral compression causes the crust to fracture, and one side is pushed over the other, crumpling in the process. This is well shown in the Ardennes (Fig. 96),

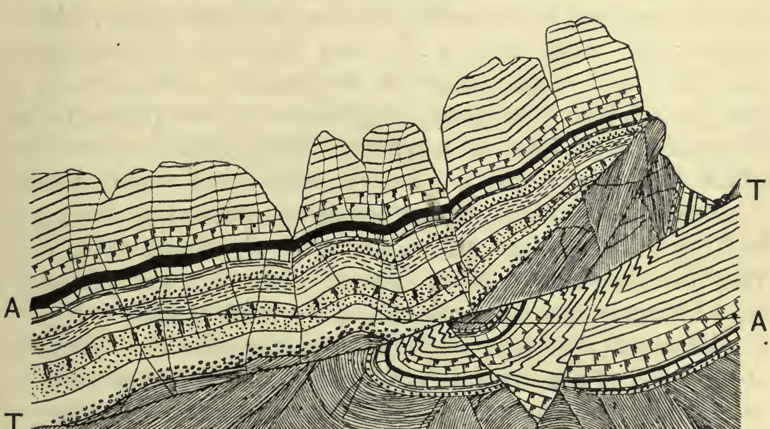


FIG. 96. Section of the Ardennes and the Belgian coal-field. (After Cornet and Briart.)

AA. The present surface of the ground. The rocks above this line have been removed by denudation. *TT*. The great thrust-plane. The mass of rock above *TT* has been pushed forward over the mass below.

which are the remains of an ancient mountain chain, now very greatly denuded so that its internal structure is exposed. In this case, moreover, our knowledge of the structure is increased by the borings which have been put down for coal.

The folding and fracturing are often much more complex than in any of these diagrams; but the examples given will serve to show the general nature of the movements which give rise to mountain chains.

Earthquakes

An earthquake, as the word implies, is a shaking of the crust of the earth. Sometimes it is accompanied by a permanent elevation or depression of the ground; but often no lasting effect is visible at the surface except the damage done by the shaking.

In many cases earthquakes are undoubtedly closely connected with the earth movements of the types already described. They are most frequent where there is evidence of recent folding or faulting and where it is probable that the movements have not yet ceased. In the Japanese earthquake of October 28, 1891, and in the San Francisco earthquake of April 18, 1906 visible faults were formed at the surface of the ground, but these may have been the consequence rather than the cause of the earthquakes. The actual movements to which earthquakes are due are probably more deeply seated.

Earthquakes, however, are not always caused by folding or faulting. They are sometimes the result of volcanic explosions. They are common in most volcanic districts; and eruptions are often preceded or accompanied by earthquakes.

Whatever the cause may be an earthquake is a vibration of the crust of the earth. Sometimes actual waves are seen to travel along the ground, like waves on the surface of a sheet of water. More often the waves are too long and low to be visible to the eye, but the rocking of the ground may be felt as they pass. Buildings sway to and fro and fissures sometimes open in the ground and again close up.

The earthquake usually originates some miles beneath the surface, and from the origin or "seismic focus" the vibrations spread in all directions (Fig. 97). They reach the surface first at the point immediately above the origin and this point is called the epicentre. It is at the epicentre that the shock of the earthquake is first experienced, and on the ground it seems to spread outwards as waves spread from a stone thrown into a pool of water.

If the time of arrival of the earthquake is observed at a number of different places it is usually possible to determine the position of the epicentre. Lines, called homoseismal lines, are drawn through places which were affected by the earthquake at the same moment. These lines are generally elliptical in shape, and the middle of the ellipse is the epicentre. The actual place of origin of the earthquake is below the epicentre.

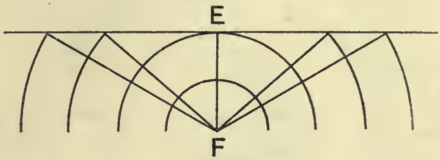


FIG. 97. Seismic focus (*F*) and epicentre (*E*) of an earthquake.

Observations of the time of an earthquake, however, are seldom very accurate, and other methods are therefore usually employed. As the vibrations spread outwards from the origin their intensity diminishes. Far away from the centre the movement may be sufficient to displace ornaments on brackets or shelves but not to injure buildings. Nearer to the centre tall chimneys may be overthrown while houses are but little affected. Nearer still the shock may be so great that not even the most solidly constructed buildings remain standing. By classifying the damage in this way it is possible to draw lines through places which have suffered equally, and such lines are known as isoseismal lines (Fig. 98). Isoseismal lines are often very irregular in form, because the damage depends in part upon the nature of the foundations on which the buildings rest, as well as upon their distance from the origin. Nevertheless if a number of the isoseismal lines can be drawn, it is usually possible to determine approximately the position of the epicentre.

It might be expected that the greatest amount of damage would be at the epicentre itself, but this is not always the case. At the epicentre the intensity of the movement is at its maximum, but it is here an up and down movement, which does less injury to buildings than a shake from side to side. The greatest damage is done where the earthquake wave emerges obliquely, but is still near enough to the origin to retain a considerable proportion of its energy.

FIG. 98. Isoseismal lines of the Inverness earthquake of 1901 (Davison).

In recent years it has been shown that in an earthquake at least three distinct kinds of vibrations or waves are set up. There are longitudinal waves, like those of sound in air, in which the particles move to and fro in the direction in which the wave is travelling. There are transverse waves, like the waves which run along a rope which is fastened at one end, stretched fairly taut, and shaken at the other end. In such waves the particles move to and fro at right angles to the path of the wave. Finally there are surface waves, which travel along the ground like the waves on a sheet of water. In these the motion of the particles is also transverse, but they differ from the transverse waves in the body of the earth.

Close to the epicentre all these waves reach the observer so nearly together that they cannot be distinguished. Farther away there is an interval of time between them, which increases with the distance from the origin. The waves which pass directly through the earth reach the place of observation before the surface waves, partly because their path is shorter and

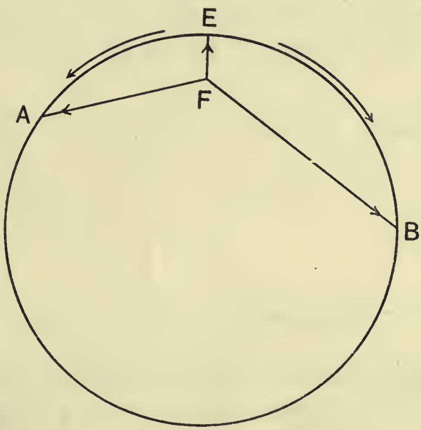


FIG. 99. Paths of direct and surface waves.
F. Seismic focus. *E.* Epicentre. *A, B.* Places of observation.

partly because they move more rapidly (Fig. 99). Consequently at a considerable distance from the epicentre there are as a rule at least three distinct sets of disturbances, the first two due to the longitudinal and transverse waves which travel through the earth and the third to those which travel along the surface; and the greater the interval between these disturbances, the greater is the distance of the origin.

Instruments are now made of such extreme delicacy that they will record the vibrations due to an earthquake on the opposite side of the globe, and thus it is possible to say, from observations made in England, that an earthquake has occurred a thousand or two thousand or, it may be, as much as eight or nine thousand miles away. The distance of the origin can be given, approximately, and even, with certain forms of instrument, the direction in which it lies.

Depth of origin. It is a comparatively easy matter to find the epicentre of an earthquake, but it is difficult to determine the depth of the centre or origin.

At one time it was supposed that the cracks in buildings served as an indication of the angle of emergence of the earthquake wave. It was assumed that they were formed at right angles to the path of the vibrations, and in that case observations at two or three points suitably placed would be sufficient to determine the position of the origin (Fig. 100). But the

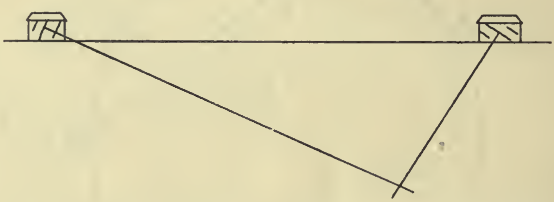


FIG. 100. Determination of depth of seismic focus.

results obtained were often not concordant, and there can be no doubt that the structure of the building and other circumstances have a considerable influence upon the direction of the cracks.

Other methods have been tried, but none are completely satisfactory. There is little doubt, however, that the depth of the origin is always small compared with the diameter of the globe, and probably never exceeds 30 miles.

Earthquake waves in the sea. When an earthquake originates either below or near the sea so that the vibrations of the crust are still intense on the floor of the sea, the water above is greatly disturbed and waves of much larger size than are ever due to wind may be started and may travel enormous distances. Such waves are often popularly, but quite incorrectly, spoken of as tidal waves.

In many of the great earthquakes of the past more loss of life has been caused by these sea waves flooding the coast than by the earthquake itself.

FIG. 101. Seismic regions of the world (Milne).

Distribution of earthquakes. In some parts of the world, such as Japan, there is hardly a day without an earthquake of greater or less intensity; but there are large areas also in which earthquakes are extremely rare. The principal earthquake regions are shown in Fig. 101. In most of them there is geological evidence of recent earth movements; and in many, but not in all, there are numerous active volcanoes.

CHAPTER XVIII

SHORE LINES

THE shape and character of a coast are determined partly by the movements of elevation and depression which have affected the earth's crust, partly by the action of the sea and of other denuding agencies. They depend also to a large extent upon the nature of the materials which form the land.

Destructive action of the sea. The waves of the sea continually falling upon the shore gradually wear away the land, whether its margin is a shelving beach or a steep and rocky cliff. If the coast consists of loose material, the waves themselves may wash away the fragments; but if it is formed of firm and solid rock their action is indirect. The pebbles and boulders that they throw against a cliff serve as battering-rams, and the face of the cliff is slowly broken up, especially at its base (Pl. VIII, fig. 2). Most rocks, moreover, are fissured, at least to some extent, and the fissures are filled with air. When a wave rises against a rock or cliff, the air in the fissures is compressed; when the wave falls away the air expands. By this alternate compression and expansion the fissures are enlarged and masses of rock are broken off.

The blow-holes which are met with on some coasts have been formed in this way. A cave is hollowed out either by the boulders and pebbles thrown against the cliff or by the compression and expansion of the air in a fissure. If the mouth of the cave is closed by the water at high tide, every wave that rises outside increases the pressure within the cave, and as the wave falls the pressure is again diminished. In this way the rock which forms the walls and roof of the cave may be broken up although it is not reached by the waves themselves. The cave is gradually enlarged and finally a hole is opened through

S. H. Reynolds, photo.

Fig. 1. **Wave-cut rock-platform.**
North of Stonehaven.

S. H. Reynolds, photo.

Fig. 2. **Caves in New Red Sandstone.**
Near Paignton.

the roof, perhaps some hundreds of yards inland. During storms at high tide the waves may now enter the cave and perhaps dash upwards through the blow-hole in the roof.

If the shore slopes gently into the sea, the action of the waves is almost entirely confined to the part between tide-marks, extending only a little below low water and in heavy storms a little above the true high-water level. Consequently this part of the land is worn away and the material so worn is deposited between tide-marks and below low-water mark. A kind of nick or terrace is thus produced as in Fig. 102 (see also Pl. VIII, fig. 1). The sea will continue to act for some time upon the cliff which forms the inner edge of the nick, and the beach will gradually be widened. At length, however, the beach becomes so broad that the waves no longer reach the cliff; and unless the beach material is removed in some way, erosion by the sea will practically cease.

FIG. 102. Formation of a beach.
The horizontal lines show the level of high water and low water.

When the coast is abrupt and the sea washes against a cliff, the action of the waves is still most effective near the level of the water. Consequently between tide-marks the rock is worn away, and if the sea were the only erosive agent the cliff would overhang. But while the sea is working at the base the ordinary agents of weathering are wearing away the top.

The form of the cliff will depend in part upon the relative rapidity of these two actions. If the weather acts more rapidly than the sea the face of the cliff will slope towards the water. If the sea acts more rapidly than the weather the cliff will overhang.

The form of the cliff is also greatly affected by the stratification and jointing of the rock, for it is along the planes of bedding and along the joints that rocks usually break most easily. If the beds dip seawards and the joints landwards, large blocks of rock will easily be detached and will fall upon the beach and the cliff will rise in a series of overhanging steps (Fig. 103). If on the other hand the beds dip landwards and the joints seawards, the blocks, even when loosened, will not fall until they are actually undermined. Erosion at the base is in this

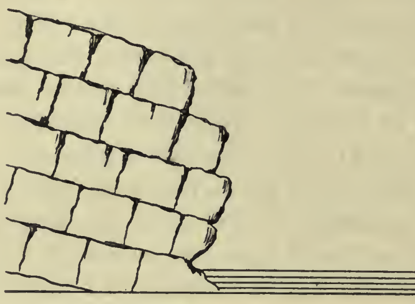


FIG. 103. Erosion of cliff (beds dipping seawards).

case less effective and the face of the cliff will usually slope towards the sea (Fig. 104).

Soft rocks are in general more easily eroded than hard ones; and consequently, if the coast is made of rocks of different kinds, the soft beds are worn into bays or gulfs, while the harder beds stand out as promontories. But there is a limit

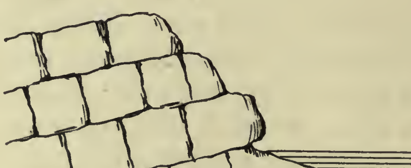


FIG. 104. Erosion of cliff (beds dipping landwards).

to the landward extension of the gulf, for the water within it is sheltered from the wind and at its head the waves are much less violent than at the extremities of the promontories. The more deeply the gulf is cut into the land, the weaker are the waves that reach its head and the slower is the erosion. Moreover, in its quiet waters sand and shingle may collect and serve as a protection to its shores. In time accordingly the rate of erosion at the head of the gulf becomes equal to that at the ends of the promontories and the gulf no longer increases in length.

The long narrow gulfs of the south-west of Ireland probably owe their formation in part to this kind of action, though other causes may have helped in their production (Fig. 105). Each gulf is cut along a strip of Carboniferous Limestone, while the intervening promontories consist of Old Red Sandstone. The erosion of the limestone, however, has not been due entirely to the action of the sea, for in some cases the depressions which

FIG. 105. South-west coast of Ireland.
Old Red Sandstone and more ancient rocks, black; Carboniferous Limestone and later rocks, dotted.

form the gulfs are continued inland as river valleys. Subsidence of the land may have assisted the invasion of the valleys by the sea.

The broad bays of the south coast of England are similarly hollowed out in soft beds while the capes are formed of harder rocks. But the width of the soft beds is greater than in south-western Ireland, and the harder rocks are not so resistant. There is still another reason for the difference in the form of the bays. In the south-west of Ireland the tide approaches and recedes at right angles to the coast, and helps to keep the gulfs free from loose material. In the south of England the tide runs along the coast and sweeps the sand and shingle into the bays. The deposit formed in this way helps to protect the shores of the bays from erosion, while the projecting capes are exposed to the full force of the waves. In the south of England, moreover, on account of the softer nature of the rocks, there is far more loose material to form protecting beaches than in the south-west of Ireland.

Constructive action of the sea. The material worn away from the land must be deposited somewhere. Sometimes it is carried below low-water mark and helps to form the continental shelf. It is in fact definitely removed from the land and placed beneath the sea. But this is not always the case. If a current runs along the shore the loose material will be drifted laterally, and although it is removed from one part of the land it may be added to another part. On the eastern coast of England for example the sea in many places is encroaching on the land, while in the Wash, in Romney Marsh and elsewhere the land is gaining on the sea, and the gain is partly due to the action of the sea itself.

It is on the southern and eastern coasts of England that the lateral drifting of the loose material is most conspicuous. Generally, though not invariably, it follows the direction of the tides, for on these coasts the currents are mostly of tidal origin. Accordingly on the eastern coast the drift is usually from north to south, on the southern coast from west to east. On the Atlantic coast the tide approaches nearly at right angles and there is but little lateral drifting.

The lateral movement of the beach is clearly shown at many seaside towns, where groynes have been thrown out to protect the cliffs. A groyne is a wall of wood, stone or brick, built outwards at right angles to the shore. The sand and shingle drifting along the shore is piled against one side of the groyne. On the other side, where there is no heaping up, the beach is usually at a lower level for some distance, until in fact the effect of the next groyne begins to be felt. The object of the groynes is to prevent the loose material from travelling onwards and thus to cause the accumulation of a beach which will prevent the waves from reaching the cliffs.

When the shape of the coast is such as to deflect the lateral current outwards the sand and shingle will be carried outwards also and in time will form a spit of low-lying land projecting into the sea. Spurn Head for example is a spit formed by the loose material which has been swept down the coast of Yorkshire.

Sometimes the spit is carried completely across the opening of a bay and converts the bay into a lagoon. This is the case with Chesil Bank (Fig. 106), which is a shingle spit connecting

FIG. 106. Chesil Bank and Portland Island.

the Isle of Portland with the mainland. The Bank is nearly parallel to the original coast, but is separated from it by a narrow channel.

Where the sand and shingle is drifted across the mouth of a river, the current of the river may be strong enough to keep the opening clear. Frequently, however, the river is unable to remove the drifting material and a submerged bar is formed across the mouth, or even an actual spit of land. In the latter case the mouth of the river is turned to one side, and as the spit continues to grow the river is more and more deflected. Both the Yare and the Alde show such a deflection and the latter is an especially striking example. The Alde flows seawards until near Aldeburgh it is only about a hundred yards from the beach; but instead of entering the sea at this point it turns southwards and for eight or nine miles it runs nearly parallel to the coast before it reaches the sea.

When the river opens into an estuary the formation of the

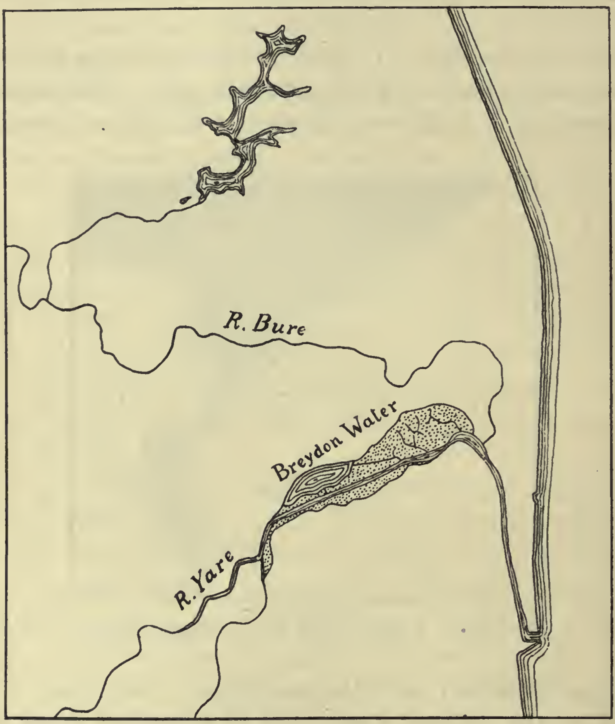


FIG. 107. Coast of Norfolk near Yarmouth.
Breydon Water is shown as it was in 1885. Practically the whole of it is now uncovered at low tide.

spit across it converts the estuary into a lagoon. This is the case for example with the Alde above Aldeburgh and the Yare above Yarmouth (Fig. 107). Usually the lagoon will communicate with the sea; but if the flow of water in the river is small, the mouth may be completely closed and the water escapes by percolating through the spit of sand and shingle.

Shores formed by depression. When a large area of the earth's crust is slowly depressed and a long established continent quietly subsides, the sea spreads over the edge of the land and enters the river valleys. The coast-line will then become irregular, and its form will depend upon the character of the flooded land.

Coast-line of subsided lowland region. If the land is low-lying and consists of soft rocks the valleys will be broad

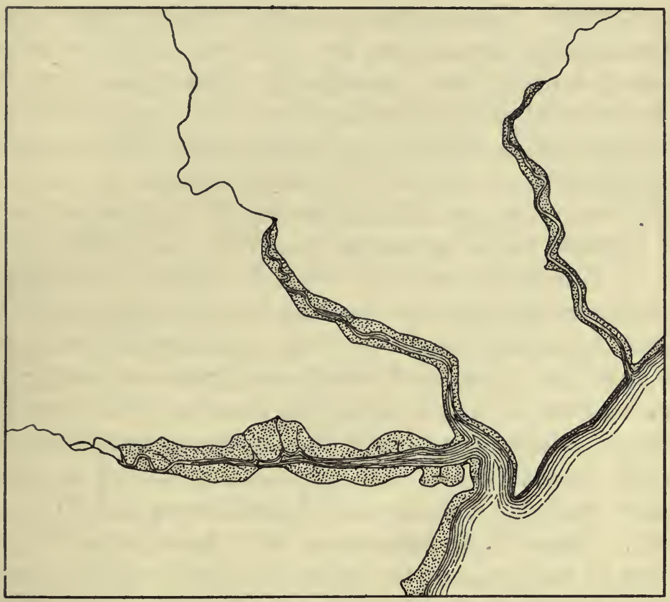


FIG. 108. Mouths of the Deben, Orwell and Stour. The dotted areas are uncovered at low water.

and shallow and often meandering. A very slight depression of the land will cause the sea to spread up the valleys and the coast-line will become excessively winding. On account of the gentle slopes of the flooded land the shores will be low and often there will be salt marshes which are covered at high tide and uncovered when the tide is low. In Essex and the southern part of Suffolk much of the coast is of this character. The sea has spread far up the valleys of the Orwell, the Stour, the

Blackwater and other rivers, and the shore is low and often marshy (Fig. 108).

A coast of this kind, however, seldom maintains its character for long. Because the rocks are soft the sea quickly wears away the prominent points, and the deposits brought down by the rivers gradually fill up the estuaries.

If tidal or other currents run along the shore, sand spits will be formed across the mouths of the irregular inlets, cutting them off from the open sea and converting them into lagoons or backwaters. Thus by the erosion of the promontories and the formation of spits across the inlets the coast-line becomes smooth and even ; but immediately behind the coast there will be backwaters of irregular shape. In course of time the backwaters will be filled up by deposits and will be converted first into marshes and later into low-lying flats.

The various stages in the process are well shown on the coast between the Thames and the Wash. In Essex and the south of Suffolk the sea has evidently entered the valleys of the rivers and the outline of the land is very irregular (Fig. 108). Prominent points such as the Naze are gradually being worn away, while at Landguard Point near Felixstowe there is the commencement of a spit across the mouths of the Orwell and the Stour. Farther north the sand spits are more fully developed, as in the case of the Alde already described : all the irregular inlets have been cut off from the sea and the coast is smooth. Some of the backwaters have been completely filled up, but in many cases the process is not complete and the partially filled lagoons remain as the " Broads " characteristic of this part of the country (Fig. 107).

Coast-line of subsided highland region. If the land is high and rises abruptly from the sea, and the rocks are hard, the valleys will usually be narrow and deep. They will not as a rule meander, but there will be branches or tributary valleys. When such a region is depressed the new coast-line will be deeply indented. The flooded entrances of the valleys will become long and narrow gulfs, often branching inwards. The hill tops near the old sea margin may be completely surrounded by water and form groups of islands fringing the shore of the mainland. The coast will everywhere be steep and rocky and there will be little or no beach. In Norway and the west of Scotland this type of coast is beautifully shown.

The characteristic features of such a coast may long remain unaltered by the sea. Owing to the hardness of the rocks even the most exposed points offer great resistance to the waves and the amount of fragmental material produced is comparatively small. Partly for this reason and partly on account of the steepness of the seaward slope there is either no beach or the beach is very narrow. Lateral currents find but little loose material to carry and spits of sand or shingle are seldom formed. Consequently the general character of the coast changes much more slowly than when the land is low-lying and is composed of softer beds. Nevertheless even the hardest rocks give way in time; the promontories are gradually worn back, the gulfs are filled with deposits brought down by the rivers, and at last the irregularities of the coast-line may disappear.

In spite of the small amount of beach material, lateral currents may help to smooth the outline of the coast, for they may distribute river deposits along the shore, or they may even bring material from some other region where the rocks are soft.

Wales, like Scotland, is an ancient rocky highland partially submerged, but while the west of Scotland is deeply indented the shore of Cardigan Bay is comparatively smooth. In the west of Scotland the tide approaches and recedes at right angles to the general trend of the coast, and the ebbing tide helps to keep the firths and lochs free from deposit. In Wales the tidal wave runs along the shore and sweeps the loose material into the sheltered inlets. It is because nearly all the inlets are filled with deposit that the outline of Cardigan Bay is smooth.

Coast-line of subsided mountain range. If the subsiding coast is formed by a range of mountains similar to the Alps, the sea will enter the longitudinal valleys between the individual chains which constitute the range. The higher hills of the outer chains will be represented by lines of islands off the mainland and the valleys will form narrow gulfs almost parallel to the general direction of the shore-line. The Dalmatian coast is one of the best examples of this type.

Shores formed by elevation. A continental area is usually surrounded by a continental shelf. The shelf is smooth and slopes gently outwards, and generally it is covered by mud, sand and other soft materials. If the land is raised bodily, a part of this shelf is exposed to view and forms a fringe of low-lying and often marshy ground around the original mass of land. Such a fringe is called a coastal plain. It slopes very gently to the sea and the slope is continued beneath the water to the edge of the continental shelf. The shore-line is smooth and even except where it is broken by the mouths of rivers. Inland the coastal plain often ends abruptly against the foot of the cliffs which originally formed the coast.

In the United States a well defined coastal plain extends from the neighbourhood of New York southwards along the Atlantic shore, attaining a width in Georgia of about a hundred miles. In the north on the landward side the plain is bounded by a steep declivity, which was the former coast-line and is now the edge of an elevated plateau. Down this declivity the rivers form falls and rapids, and the edge of the plateau is accordingly known as the Fall-line. It is a feature of considerable importance. Up to the Fall-line the rivers are navigable, while the falls themselves are valuable as a source of power. Hence many of the most important towns of the eastern United States are situated upon the Fall-line.

The seaward margin of the American coastal plain is not so smooth as might be expected if the plain were simply an elevated portion of the continental shelf. In the north, especially, it is marked by numerous irregular inlets. After the plain was raised above the sea it was eroded by rivers and a slight subsequent depression has allowed the sea to enter the valleys worn by the rivers.

Movements of the mountain-building type. When the earth movements are of the mountain-building type an entirely different kind of coast may be produced. Such movements have often taken place near the margin of a continental mass and they may result in the elevation of a mountain range beneath the sea. The higher parts of the range will form chains of islands outside the coast of the mainland. The West Indies and the festoons of islands off the eastern coast of Asia have been formed in this way. It must not be supposed however that in such cases the original coast of the mainland remains unaltered, for the elevation of a mountain chain is often accompanied by the sinking of large blocks of the earth's crust, especially on the inner side of the mountain arc. The Sea of Japan and the other bordering seas of the east of Asia appear to owe their origin in part to this cause.

Coasts formed by faults. When the earth's crust collapses like a breaking arch, blocks of the crust sink downwards and other blocks perhaps are raised. If the falling blocks subside beneath the sea and the upstanding blocks remain land, the coast will coincide with the fracture or fault. In plan such faults are usually straight or gently curved, in section they are nearly vertical. The coast will therefore be nearly straight and very steep, the land rising abruptly from the sea.

Parts of the coast of China seem to have been formed in this way ; but perhaps the western coast of India from Bombay southwards is a still better example. There is reason to believe that the land-mass of India once extended much farther towards the west, but in geologically recent times the western extension sank beneath the sea. The coast has since been modified by denudation, but it is still remarkably straight and abrupt.

CHAPTER XIX

DELTAS AND ESTUARIES

WHEN a river enters the sea its current slackens and finally ceases and its load is deposited upon the floor of the sea. The coarser and heavier material is laid down first and the finer and lighter material is carried farther out ; but eventually the whole must fall. Off the mouth of the Amazon the sea is sometimes discoloured by mud to a distance of one or two hundred miles, but even this is only a small fraction of the total width of the ocean ; and in general the river deposits are confined to a comparatively narrow belt near the coast.

If there are tides or currents in the sea where the river enters, some of the material brought down by the river may be carried farther out or may be drifted along the coast and laid down many miles away. The tides and currents may be sufficiently strong to prevent any considerable deposition, and the mouth of the river will then be kept open and will form an estuary.

If on the other hand the sea itself is still, the river will deposit its burden near its mouth and will form a delta.

Accordingly we find in general that deltas are characteristic of nearly tideless seas and estuaries of seas in which the tides are great.

Estuaries. According to the derivation of the word an estuary is the tidal portion of a river, but the term is usually confined to rivers that have a single mouth, and it is applied more particularly when the mouth opens widely out towards the sea. The mouths of the Thames and the Severn for example are typical estuaries; the Hugli, although it is tidal, would hardly be called an estuary, because it is only one of the branches of the Ganges.

Since a river naturally tends to deposit its burden at its mouth, an estuary cannot be permanent unless the deposit is removed by some other agent, and the removal is usually due to the tide. The river is dammed by the rising tide at its mouth, and its current may even be reversed for some distance upwards ; but when the tide is falling the pent-up water escapes and the current is increased beyond its normal strength. The tide in fact flushes the river's mouth and helps to keep it free from sediment.

The flushing takes place whatever the direction of the tidal wave may be, but it is most effective when the tide runs straight into the river's mouth. When, on the other hand, the tidal wave travels across the opening of the river it has two opposite tendencies. By drifting sand and shingle in its course it attempts to form a bar or spit across the mouth ; by flushing the river it helps to keep the opening clear. Which of these tendencies prevails will depend upon the relative strength of the two kinds of action.

When the land is made of soft rocks the funnel-like shape of a typical estuary may be due to the scouring action of the tide; but more often it indicates depression. Most estuaries which have this characteristic form occur in regions where there is other evidence of subsidence. The sea has flooded the river valleys and the shape of the estuaries has been preserved, rather than actually made, by the action of the tides.

Deltas. Deltas are formed when the deposits of a river are not removed by tidal or other currents, and they are especially characteristic therefore of tideless seas. Even where the tides are great, however, a river may build a delta if it brings down more material than the tides can carry away; but such a delta will usually be abnormal and irregular in shape as, for example, in the case of the Rhine.

Deposition of coarse material. The normal form of a delta depends to a large extent upon the nature of the river's burden. The coarser material that is rolled along the bed, is dropped immediately at the river's mouth, where the current slackens, and it forms a bank upon the floor of the sea. The deposit will be greatest where the supply of material is greatest; and this will be where the current is strongest, which is usually near the middle of the river. The sand and pebbles are swept along the bank and are tipped over its end and sides, and the bank grows outwards like a railway embankment in course of construction. But as the bank increases in height and length it obstructs the current more and more, until it becomes easier for the stream to divide and flow to each side. Each branch will then begin to form a new bank of its own. In time these branches will again divide and subdivide and so the deposit grows outwards in the form of a fan. The surface of the fan slopes gently seaward and is marked by irregular branching channels. The outer edge has a steeper slope, corresponding with the angle of rest of the material that is tipped over it. The process of formation may often be watched upon the sea-shore when a little stream runs down the sand into a quiet pool.

Evidently the river can never raise any part of the delta above its own flood level, but the deposits formed during floods may at normal times stand up above the water, and the branches of the river may then be confined to definite channels. Vegetation may grow upon the higher parts and in course of time the gradual accumulation of vegetable matter may even raise the surface above the level of the floods.

Deposition of suspended material. Most large and long-established rivers, however, drop the coarser part of their burden far above their mouths, and it is only the finer silt that reaches the sea. Most of this is carried in suspension and the mode of deposition is not quite the same as in the case of the material that is rolled along the bed.

One of the reasons for the difference is the peculiar action of salt water upon suspended mud. If pure and finely levigated clay is shaken up with fresh water and is then allowed to stand, the water will remain muddy for many hours. But if a few drops of brine are added, the clay settles rapidly and the water clears. With calcareous or siliceous muds the effect of the brine is much less marked or is even absent altogether. In most rivers, however, a large proportion of the silt consists of clay; and accordingly the mixture of the fresh water of the river with the salt water of the sea assists deposition.

When a river enters the sea its current is continued outwards for some distance and the mud is carried with it. In the middle of the current the speed is at its greatest and the water at its freshest, and deposition is comparatively slow; towards the edges, where the velocity is retarded and the fresh water mixes with the sea, the silt falls much more quickly. Deposition takes place throughout, but it is most rapid at the margins of the current, and on each side a muddy bank is built beneath the sea (Fig. 109). In course of time these banks are raised to the flood level, and when the river is at its normal height they stand above the

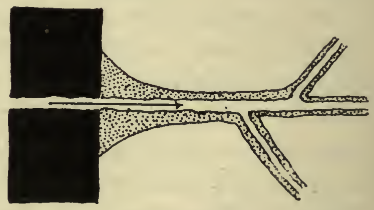


FIG. 109. Diagram of mud-delta.

water as natural embankments running out to sea. The river no longer mixes with the sea until it escapes beyond the embankments. Its current therefore reaches outwards still farther than before, and it proceeds to extend its embankments beyond their former limits.

The river now runs in a channel upon a muddy tongue projecting out to sea. In floods it may break through its bank at some weak spot and a branch of the current will escape through the breach. This branch will build embankments of its own and thus the original tongue-like delta becomes forked. On a small scale the commencement of the forking may be seen in the delta of the Dee where it enters the southern end of Bala Lake (Fig. 110).

FIG. 110. Delta of the Dee at the south-western end of Bala Lake.

By the repetition of the process a branching delta like that of the Mississippi may be developed. For many miles the Mississippi flows out into the Gulf of Mexico upon a tongue of alluvial deposit. Towards the extremity of the delta it divides and subdivides and each branch runs between narrow banks of mud (Fig. 111).

The major branches of the river will usually diverge from one another, but it may happen that two subsidiary branches meet. Their embankments will then unite, enclosing a lagoon, which may be gradually filled up by deposits formed during floods, perhaps assisted by growing vegetation.

It is only at the surface that a muddy delta assumes the branching form described. Beneath the sea its shape is different. The embankments are not vertical walls but, being made of mud, their outer slopes are very gentle. Consequently at a very small depth they become confluent with one another and the branches are no longer separate. The deposit spreads far beyond the embankments and upon the sea floor it forms a tongue which perhaps may be irregularly lobed. The slopes

FIG. 111. The terminal portion of the Mississippi delta.

are everywhere extremely gentle and the embankments are only the highest and visible portions of the delta.

A branching delta can only be constructed where the river is free from external interference. If there are currents belonging to the sea itself the shape is liable to be greatly modified,

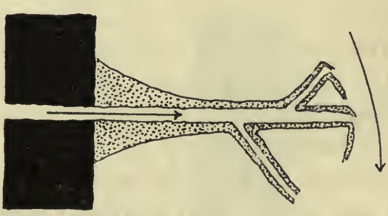


FIG. 112. Effect of currents upon the form of a mud-delta.

for the mud will no longer be distributed by the river alone. If a current in the sea runs across the mouths of the river, as in Fig. 112, the mud will be deflected to one side. Banks or spits of mud will be formed across the interspaces between the branches and in time may convert them into lagoons. This has happened, for instance, in the case of the Nile (Fig. 113). Both the Rosetta and the Damietta branches have built out tongues of land; but in this part of the Mediterranean a current flows from west to east and the mud brought down by the river is swept towards the east. In consequence of this a spit of land has been built eastwards from the Rosetta branch, enclosing Borollos Lake; and another spit from the Damietta

FIG. 113. Delta of the Nile.

mouth shuts off Lake Menzala. Other smaller lagoons have also been formed in a similar way. It is in fact to this current that the delta of the Nile owes its present smooth and regular outline.

CHAPTER XX

EARTH SCULPTURE

General nature of the processes of earth sculpture. The main features of the land-masses are determined primarily by the movements of elevation and depression already described. But as soon as a part of the earth's crust is raised above the ocean it is exposed to various agencies which tend to modify its form. Its margins are worn by the sea, its surface by rivers and other agents; the material removed from one place is laid down in another; and thus the form of the land is slowly changed by the two processes of wearing away or denudation and laying down or deposition. The modifications produced on the surface are often spoken of collectively as land sculpture or earth sculpture.

Under the influence of atmospheric changes many kinds of rock seem to decay and even the hardest are slowly broken up. The process is known as weathering, because it is, in fact, due to the weather. The general result is to produce a layer of loose material resting upon the solid and unaltered rock. It is in this way that much of the soil has been formed.

The loose material produced by weathering is gradually carried downwards, a process known as transport. It may be washed down a slope by rain, it may be rolled by a river along its beds, or it may even slide downwards under the action of gravity alone. But however it travels it wears the surface over which it moves, and this wearing action of moving material is called corrasion.

From time to time the transported material finds a temporary resting-place and is laid down, sometimes in the sea, sometimes on the land. If it is deposited in the sea it may alter the shape of the coast-line, if on the land it will change the form of the surface.

There is still another process by which the land is worn away. Water, and especially water which contains carbon dioxide, has a very considerable solvent action upon the earth's crust. Pure rock-salt is readily and completely soluble; pure limestone is dissolved more slowly but with equal completeness. In the case of most rocks, however, it is only some of the constituents that are dissolved, while the rest remain behind, and this is one of the principal causes of weathering.

Earth sculpture therefore is the effect of denudation and deposition, and the processes of denudation may be grouped under the heads of weathering, transport with its accompanying corrasion, and solution.

Weathering. Dry air has very little chemical effect upon the materials of the earth's crust, but few rocks can stand indefinite exposure if the air be moist. Either they gradually decay, which means that some of their constituents are decomposed, or they break up without any chemical change. The decay will go on even when the water is entirely in the form of vapour, but it proceeds more rapidly and vigorously when the vapour condenses as rain or dew. The condensed vapour will not be pure water, but will always contain gases dissolved from the atmosphere, and of these oxygen and carbon dioxide are the most important aids in weathering.

Some of the chemical changes are due to the oxygen of the air, which in the presence of moisture acts powerfully upon many minerals, especially those containing iron. Basalt, for example, when exposed to the air, becomes covered with a brown crust, consisting largely of oxide of iron.

But water containing carbon dioxide appears to be the chief agent of disintegration. It dissolves carbonate of lime with comparative ease, and in course of time a limestone may be entirely removed except the clay and other insoluble matter which it usually contains. In many sandstones, the grains of sand are cemented together with carbonate of lime. This may be dissolved and the rock will then fall to pieces.

Water containing carbon dioxide acts also on other minerals. It decomposes felspar, carrying away some of its constituents in solution and leaving the rest in the form of clay. A granite may accordingly become a mass of clay with quartz and mica scattered through it. Instead of being a firm and solid rock it will then be loose and friable, and will easily be washed away.

The longer the rainwater is kept in contact with the rock the more rapid will be the disintegration. If the rock is free from pores and cracks, and its surface smooth, the rain will run off and will have little chance of causing any chemical change. But if the surface is rough and there are crevices into which the water can penetrate, it will have a longer time to produce its effects. A polished slab of granite standing vertically will resist the weather longer than a rough block lying horizontally.

The surface soil may both aid and hinder the process of decay. It acts as a sponge and keeps the rock beneath it moist after the rain has ceased. Lichens and mosses produce a similar effect and beneath a patch of moss the stone is sometimes more decayed than on the exposed surfaces. On the other hand both soil and moss impede the removal of the decayed material and so prevent the exposure of a fresh surface to the action of the weather.

Weathering due to moist air or to clinging drops of water tends to round the corners and edges of rocks and to produce convex surfaces. If we imagine a large cube of rock as made up of a number of little cubes fitted closely together, it is evident that a little cube in the middle of the side of the large cube will expose only one of its faces to the atmosphere, at an edge of the main cube it will expose two faces and at one of the corners it will expose three faces. Therefore the weathering is most rapid at the corners of the large cube, less rapid at the edges, and least rapid on the faces; and the cube becomes rounded like a cube of sugar dropped into a glass of water.

Of all the various kinds of rock the least liable to chemical change are those consisting chiefly or wholly of silica. In its crystalline form of quartz, silica is practically unaffected by water even when the water contains carbon dioxide or the acids produced by decaying vegetation. In its non-crystalline form it is soluble, but only to a very slight extent.

But even the most resistant rocks are gradually disintegrated. Changes of temperature, causing alternate expansion and contraction of the rock itself, will break up the surface. In desert regions this is one of the most important processes of denudation; but in temperate climates a much more powerful influence is frost. No rock is absolutely impervious to water, and when the water in the pores or crevices is frozen, it expands and exerts great pressure upon the walls of the space in which it is confined. By alternate thawing and freezing the cracks are gradually enlarged and the rock is broken up. The screes or heaps of broken rock at the foot of crags in our hilly districts are due chiefly to this cause (Pl. IX, figs. 1, 2).

Effects of running water. Weathering is largely the effect

P. Lake, photo.

Fig. 1. Screes at the foot of the Cwm Clwyd Crags.
Ceiriog valley, North Wales.

S. H. Reynolds, photo.

Fig. 2. Screes on Mynydd-y-Gader.
Near Dolgelley.

of standing water, that is to say, of water which rests in hollows or cracks or which has soaked into the soil or rock. It breaks up the surface but does not remove the loose material, and if no other causes came into play, the layer of broken or decayed rock would increase in depth till the covering was sufficiently thick to protect the underlying rock from further action. Gravity alone will cause some of this material to fall to a lower level, but without the aid of water its effects are largely neutralised by friction.

When rain falls it assists the action of gravity in two ways. It acts as a lubricant, enabling the broken fragments to slide more easily upon one another. This is clearly illustrated by the tips or waste-heaps of a slate-quarry. When the useless material from the quarry is thrown down the sides of the heaps in fine weather it quickly comes to rest, and as long as the weather remains dry there is little further disturbance. But when rain falls the heaps become wet, pieces of rock which were at rest begin to slide downward and the movement usually continues until the heap becomes dry again. In this way rain helps the downward movement even of large masses of rock.

On smaller fragments the effects of rain are greater, for its own velocity as it runs down a slope enables it to carry mud and to roll down grains of sand or even little pebbles. As none of the ordinary agents of transport in a climate such as ours, excepting the wind, carry particles upwards, the effect of each shower is added to that of the previous one and consequently the soil gradually travels downhill.

Grass and trees, by preventing the free flow of the water and by binding the loose particles together, serve as a protection to the soil; and on a gentle grass-covered or wooded slope, the movement will be slow and may be imperceptible. The direct influence of rain is greatest in a hilly district where the slopes are steep and where there is little vegetation. In such circumstances the effect of a single storm of rain may be conspicuous.

When rain falls upon the ground, some of it is evaporated, some sinks into the earth and some runs over the surface. The portion which sinks in is not lost, but sooner or later, in most cases at least, it reappears, usually in a spring, and helps to form a stream. The part which runs over the surface may flow for a time as a sheet of water, but it will presently form little runnels and these will join into larger streams and eventually the water will find its way into a river. There is therefore no hard and sharp line between the work of rain and the work of rivers. The work done before the water has been concentrated into definite channels, and even the work done in a channel which is only temporarily occupied during a rainstorm, may be called the work of rain. The work done by a more or less permanent flow of water down a definite channel may be called the work of a stream or river.

CHAPTER XXI

RIVERS

It is evident to the most casual observation that a river does three kinds of work. It wears away its banks and bed; this is erosion. It carries the material brought into it by rain or by its own erosive action; this is transport. It drops this material, sometimes on its own bed, sometimes in a lake, sometimes in the sea, but always at a lower level than the point at which the material was received; this is deposition.

So far as the form of the land is concerned the erosive action is on the whole the most important, but the erosion depends to a large extent upon the material transported, and transport and erosion cannot be considered altogether independently.

Transport. Some of the material carried by a river is in solution, the rest consists of solid fragments of various sizes and shapes. Pieces of wood may float upon the surface. The smaller particles of rock sink very slowly and for most of their journey remain suspended in the water. The larger fragments are rolled along the bed; by knocking against one another they become rounded and are formed into pebbles, by rubbing against the banks and bed they wear away the channel of the stream.

The size of the pebbles that a river can move depends upon the velocity of the current. It depends also to a considerable extent upon the shape and material of the pebbles. If, however, the pebbles are all spherical and of the same specific gravity, then the diameter of the pebbles that a river can just move varies as the square of its velocity, and their volume as the sixth power of the velocity.

The solid material carried by a river is called its load. Often the term is used to include also the material in solution, but this in fact forms part of the transporting agent itself; and for the present at least it will be more convenient to restrict the term to the solid fragments which float upon the surface or are carried in suspension or are rolled along the bed.

The transporting power of a river evidently depends upon its velocity and its volume. A rapid stream can carry more material and can carry fragments of larger size than a slow one. A large river can carry more than a small one which has the same velocity.

It is often stated that for any particular river or part of a river (assumed to have a constant velocity and volume) there is a limit to the load that it can bear, and when this limit is reached the load is called the full or maximum load of the river. But without some further qualification the statement is not strictly correct, for as long as the water is moving at all it is capable of carrying a greater load, provided that the material is reduced to a sufficient degree of fineness. The effect of adding more material to the load of a river is to reduce the velocity of the current—the river has a heavier load to carry and cannot carry it so fast. The decrease of velocity diminishes the size of the pebbles or boulders that the river can move. If, in its former load, there were any pebbles or particles which it cannot move with its diminished speed, these pebbles will be dropped, but if they were broken up sufficiently small, the river would still be able to move them and they would still remain a part of its load.

Thus the full load of a river depends not only on the velocity and volume of the river but also, and to a very important extent, upon the size of the particles constituting the load. There is no definite limit to the possible load except for particles of a definite size. If the particles are small enough the river may become a mass of moving mud before the maximum load is reached. If the particles already in the river and the particles brought into it from outside are of various sizes the limit to the possible load is indeterminate.

Further complications are introduced by the fact that owing to friction with its channel the velocity of a river at the bottom and sides is considerably less than in the middle of the stream. But still it remains true that when the bulk of the load consists of particles which are approximately of the same size, there is a fairly definite limit to the possible load; and when this limit is reached, any further addition to the river's burden involves the dropping of an equivalent portion of the original load.

Erosion. The erosion which a river accomplishes is of two kinds, chemical and mechanical. It may dissolve the rocks over which it flows, if they are of a suitable nature. This is chemical erosion or solution. It may break off fragments from its bed and banks. This is mechanical erosion or corrasion. In most cases the corrasion is far greater than the solution, and it is accordingly with the corrasion that we are chiefly concerned.

Corrasion may be vertical or lateral. Vertical corrasion is corrasion of the bed of the river, deepening its channel. Lateral corrasion is corrasion of the banks, and leaves the bed untouched.

Clear water does very little corrasion of solid rock; but if the banks and bed are made of loose material, even a stream of clear water becomes an effective corrasive agent. The greater part of the corrasion done by a river is due, however, to the pebbles and sand that it rolls along its bed.

In order to see how the corrasion and the load interact upon one another, it will be convenient to assume at first that all the particles which form the load and which the river breaks off from its banks and bed are of the same material and the same size. There will then be a definite limit to the load that the river can carry.

In these circumstances, when the river has no load, it does no corrasion. When it has a full load it corrades its banks and bed; but, because it already carries as much as it can, for every particle that it breaks off it must drop one of those that it was carrying. Deposition will then be equal to corrasion, and the vertical corrasion is ineffective.

Thus with no load there is no corrasion, with a full load corrasion is equalled by deposition. Therefore there is some intermediate load for which the effective corrasion is at its maximum.

A similar law holds good even if the material that forms the load is of various degrees of fineness. The full load is not then a definite quantity; but the effective corrasion will still increase with the load up to a certain point; and beyond that point it will decrease as the load increases, and will finally cease. This is the fundamental principle of river erosion.

Grading of the river-channel. If we consider a single reach or portion of a river, in which it may be assumed that the slope of the bed and the velocity and volume of the stream are uniform, the effect of the load will become evident. For simplicity it will again be convenient to suppose that the particles of the load and the particles which the river breaks off are all of the same size and material.

The load that enters the reach depends upon what has happened above, and not in any way upon the reach itself. It may therefore be exactly the full load for the velocity and volume of the river in the reach or it may be more or less than this.

If the load that enters the reach is just the full load for the river in the reach, there will be no corrasion and no deposition, or, to be more exact, corrasion and deposition will be equal. The slope of the bed will remain unaltered and this part of the river is said to be "graded" (Fig. 114 *a*).

If the load that enters the reach is more than this full load, the river will be unable to carry the whole of it and the surplus material will be deposited at the head of the reach. The slope of the bed will accordingly be increased, and the velocity of the stream in the reach will therefore also be increased. This will go on until the velocity is just sufficient to enable the river to carry the load that enters the reach. In such a case the river is said to be " aggrading " its bed (Fig. 114 *b*).

If the load that enters the reach is somewhat less than the full load, the river will wear away its bed more than it deposits. In doing so it adds to its load. Accordingly the effective corrasion will be greatest at the head of the reach and will diminish downwards as the load increases. The slope of the bed, and with it the velocity of the stream, will be decreased; and this will go on until the velocity is just sufficient to enable

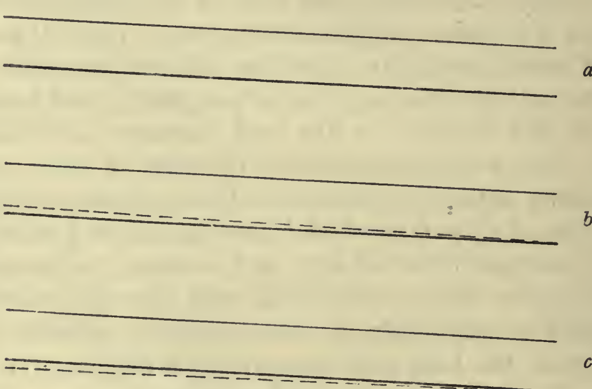


FIG. 114. Grading of river-channel.

the river to carry the load that enters the reach. In this case the river is said to be " de-grading " its bed (Fig. 114 *c*).

Thus throughout its course the river is continually engaged in adjusting the slope of each part of its channel to the load which enters that part.

If the load consists of particles of various sizes, the river tends to adjust the slope of the reach to the size of the largest particles that enter the reach in sufficient quantity. If the velocity in the reach is just enough to enable it to roll these particles along its bed, they will pass through. If the velocity in the reach is not sufficient to move them, they will be deposited at the head of the reach, increasing the slope. If on the other hand the velocity is more than is required to move the largest particles, there will be corrasion until the slope and velocity are sufficiently reduced.

Even in a single reach, however, the volume and velocity are not constant, and both are considerably increased during a flood. If most of the material enters the reach only during floods, and consists of boulders or pebbles so large that the river can move them during a flood but not in its normal state, the slope will be adjusted to the flood conditions, and at other times the water will be practically clear and will have scarcely any effect either of transport or corrasion. This is often the case with mountain streams.

If, on the other hand, a large part of the material that enters the reach is sufficiently fine to be moved by the river in its normal condition, the slope will in general be adjusted to this normal condition and the river will never be free from moving sediment excepting perhaps during a drought. This is usually the case in the lower part of a river's course.

Curve of water-erosion. We may next consider the effect of these principles upon the general slope of the river from its source to the sea. Suppose that a river starts upon a slope which is uniform from the watershed to the sea and consists of rock of the same hardness throughout. At the top the river receives only the rain that falls at the top, and its volume is therefore small. Lower down it receives not only the rain that falls at that point, but also the rain that falls on the slope above. The volume therefore increases from the source to the sea, except in dry climates where the loss by evaporation is great or in porous rocks where the water sinks into the ground.

At its source the river has no load and therefore does little or no corrasion. As it flows onwards the loose material produced by weathering is carried into it by rain and other agencies; the load increases, and with the load the corrasion also will increase up to a certain point. If the slope be long enough the river will at length attain its full load, at least for the bulk of its material, and deposition will become equal to the corrasion. At the top, then, there is no corrasion, somewhere towards the lower end deposition and corrasion are equal, and therefore at some intermediate point the effective corrasion is at its maximum.

The wearing away of the bed is greatest at this point and decreases both upwards and downwards, and the river converts the original uniform slope of its channel into a curve such as is shown in Fig. 115.

This curve, which running water tends to produce, is called the curve of water-erosion. Sometimes it has been termed the base-line of erosion, but the expression is inaccurate, for erosion does not cease when the channel has assumed this form[1].

It should be remarked that if the slope on which the river starts, is short and steep, the river may reach the sea before it has attained its full load. In such a case corrasion will be effective down to sea-level, and may in fact continue to increase

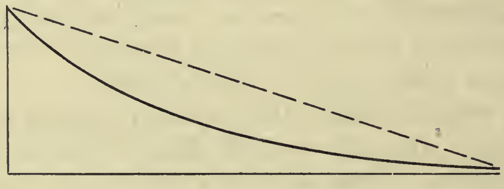


FIG. 115. Curve of water-erosion.

throughout the whole length of the river. In course of time, however, the deposits at the mouth will form a flat over which the river flows and the typical curve of water-erosion will finally be completed.

After the curve of water-erosion is produced it continues to develop. In Fig. 115 the original slope was uniform throughout. When the curve is formed the slope at the top is steeper than before, the slope near the sea is gentler than before. The tendency therefore is for corrasion to increase in the upper part of the river's course and for deposition to increase in the lower part, and the point of maximum corrasion moves up the stream. The upper part continues to become steeper and

[1] A river may correctly be said to have reached its "base-level," either temporarily or permanently, at any part of its channel where deposition is equal to corrasion. But as long as it transports any material it is evident that in some part of its course corrasion is greater than deposition.

the lower part flatter, and the curve becomes more and more accentuated (Fig. 116). The river itself cannot erode its watershed, but rain and other agencies will do so and the

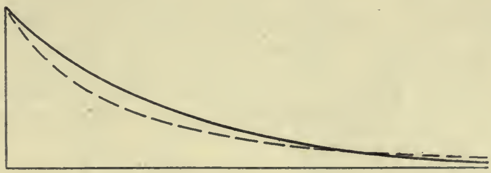


FIG. 116. Further development of the curve of water-erosion.

watershed is gradually lowered, and the average slope of the whole is reduced.

Convexity of watersheds. The concave curve of water-erosion does not as a rule extend up to the watershed. If it did, the watershed between two rivers would be sharp as in

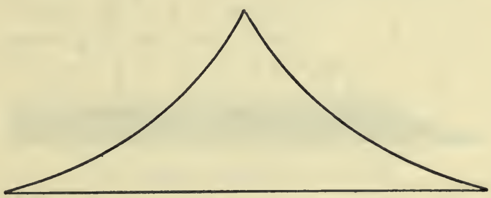


FIG. 117. Cuspate watershed.

Fig. 117; but usually the watershed is more or less rounded, and towards the top the slope becomes convex (Fig. 118).

A river does not start at the top of a hill, because there is

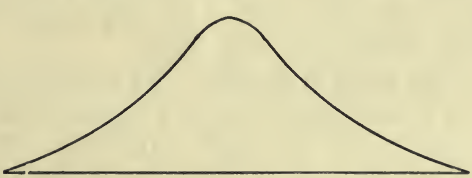


FIG. 118. Normal watershed.

no supply of water there except when it is actually raining. Some of the rain soaks into the hill and escapes as a spring lower down the slope. The upper part of the hill serves as a kind of sponge which holds the water that soaks in, and only allows it to escape gradually. Before the supply is exhausted the next shower of rain may come. The spring will then be permanent and will be the source of a permanent stream. It is about the level of such springs that the typical curve of water-erosion begins. Above this level erosion is due to weathering rather than to corrasion, and the slope is usually convex. The loose material produced by weathering is washed away during showers of rain.

The chalk hills of south-eastern England often show very clearly the change of curve at the level of the springs. The chalk itself is very porous and beneath it there is usually a bed of porous sandstone called the Upper Greensand, but the latter rests upon an impervious bed of clay. A large proportion of the rain that falls upon the chalk sinks into the rock instead of

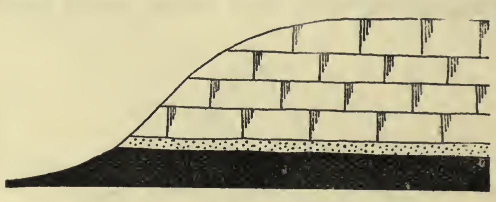


FIG. 119. Form of chalk hills.

running over the surface. But it cannot pass through the clay and therefore escapes on the side of the hill at the junction of the Greensand and the clay, forming a line of springs. Below this level the hill slope is usually concave but above the springs the slope is convex. The form of the lower slope is due to running water, the form of the upper part of the hill is due to solution of the chalk (Fig. 119).

Erosion at the heads of streams. No river can erode above its source, but nevertheless there is a tendency for the valley to be prolonged beyond the head of the river. The curve of erosion produced by the river is here at its steepest, and consequently the corrasive effect of rain at this point is at its maximum; it acts not only on the sides of the valley but also at the head, in the same direction as the stream itself, and consequently the valley is continued beyond the source of the permanent stream. As the erosion goes on, the springs are tapped farther back, and thus not only is the valley cut backward at its head, but also the source of the river recedes.

An example will show the stages of the process more clearly. Fig. 120 represents a hill the top of which is made of porous rock resting upon an impervious bed. At *A* the water which soaks into the porous rock escapes, forming the source of a stream. The curve of erosion due to the stream begins at *A*. Above this point the rain at first does little corrasion, because for the most part it sinks into the porous rock, and the curve at the top of the hill is accordingly produced by weathering and is convex. But, as the curve of erosion is developed, the slope at *A* becomes steeper and

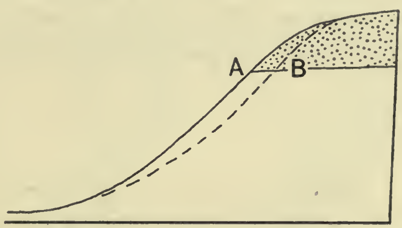


FIG. 120. Cutting back at the head of a stream.

steeper; the rain begins to run over the surface so rapidly that it has not time to sink into the ground, and it becomes an effective corrasive agent as well as an agent of weathering. Therefore there is now erosion at *A*, the hill slope is cut back to the dotted line, and the stream begins at *B* instead of *A*.

Whatever may be the cause of the springs which form the sources of the permanent streams, a similar process will go on, and therefore there is always a tendency for rivers "to cut backwards at their heads"—this is the common and convenient mode of expression, but the cutting backwards is done by other erosive agents rather than by the river itself.

Development of the river valley. As the curve of water-erosion develops, the difference between the upper and lower parts of a river's course becomes more and more clearly defined. In the former corrasion increases and exceeds deposition, in the latter deposition increases and at length exceeds corrasion. In the upper part of its course therefore the river tends to deepen its channel and this part is accordingly often known as the Valley tract. In the lower part the river spreads out the material that it carries, and tends to form plains, and this part is often called the Plain tract. Owing to the gradual flattening of the lower part of the curve the Plain tract spreads slowly up the stream and the extent of the Valley tract diminishes.

Valley tract. In the upper part of the river's course corrasion is greater than deposition. The river therefore deepens its channel and forms a valley. If the river were the only agent concerned the sides of the valley would be vertical. But while the river is cutting downwards, rain and frost and other agencies wear away the sides and the valley becomes **V**-shaped. The steepness of the sides depends upon the rate of this lateral wearing as compared with the rate of deepening.

FIG. 121. Cross-sections of river-valley in the Valley tract.

If the rocks in which the valley is cut are hard and resistant the sides will be steep and the **V** will be narrow. If the rocks are soft and easily worn away, the valley will form a wide and open **V**. The steepness of the sides is also influenced by the amount of rain. If there is but little rain, the sides of the valley will be steep, even though the rocks may be comparatively soft. If, on the other hand, the rainfall is heavy, the slope of the sides will be gentle unless the rocks are very resistant (Fig. 121).

It is partly on account of the dryness of the climate that the walls of the Colorado cañon are so nearly vertical and the cañon itself so narrow.

Plain tract. In the lower part of a river's course, the river deposits more than it corrades. Consequently the valley is not deepened and there is no vertical corrasion. But the river can still wear away its banks and lateral corrasion still goes on.

The corrasion will be balanced, or more than balanced, by deposition in some other part of the bed, but it may result in a shifting of the channel.

If for any reason one part of the bank is more easily worn away than the rest, the course of the river will become slightly curved, and when once a curve is formed the river tends to accentuate it. In Fig. 122 *a* the flow of the river is shown by the arrows. The current is directed towards the concave side of the bend at *A*, while on the convex bank at *B* the water will be slack. Consequently the bank will be worn away at

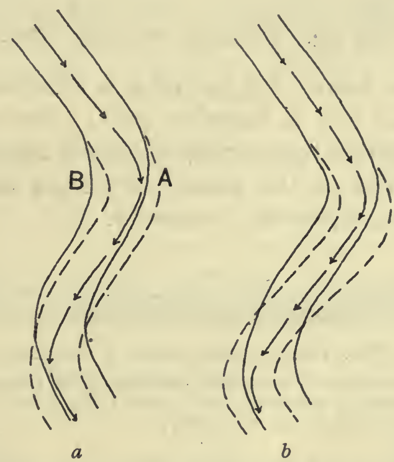


FIG. 122. Stages in the development of meanders.

A while at *B* there will be deposition, and the bend will become more pronounced, as shown by the dotted lines. Such curves in the course of a river are known as meanders. The same process will still go on until the meander forms almost a complete circle as in Fig. 123 *a*. At length the river will cut through the narrow neck at *C* and its course will again become almost straight. The former meander will remain as a backwater for some time; but the entrances to it will gradually be silted up (Fig. 123 *b*), because the current now goes by the shortest route and in the meander the water is still. The backwater will be completely cut off from the main channel and will form what

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is known as an ox-bow lake. Such lakes are very common at the sides of the Mississippi.

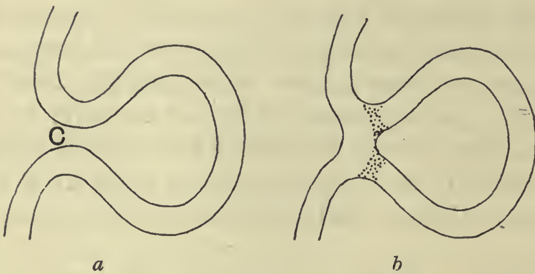


FIG. 123. Formation of ox-bow lakes.

The concave bank of a meander is continually being cut into by the river and is therefore steep; the convex bank is formed by deposition by the river itself and consequently slopes gently downwards to the water. Fig. 124 shows a typical section across the river in a meander.

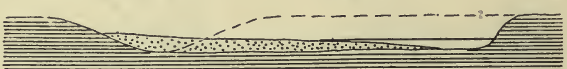


FIG. 124. Section across a meander.

The broken line shows the original position of the river-channel. The dotted portion represents the material deposited by the river on the convex bank of the meander.

Since, in the Plain tract, the river is continually corrading laterally, forming meanders and cutting them off, its course is always changing. It swings, as it were, from side to side and thus produces a nearly level flat of considerable width. The

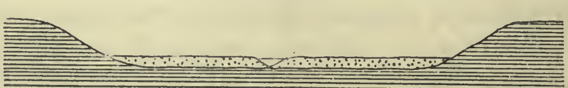


FIG. 125. Section of river-valley in the Plain tract.

limits of its lateral movements are marked by rising ground on each side of the broad and even floor through which the river meanders, and the valley therefore has the form shown in Fig. 125. The channel of the river is cut within the floor, but because there is no vertical corrasion it is sunk but little beneath the level of the surrounding flats.

In flood time the water rises above its banks and spreads over the floor of the valley, which is accordingly often called the "flood-plain." When this happens the current is strong within the channel but over the flooded area the water is practically still. Much mud is deposited at the margin of the current, because there the speed is slackened; and a bank of mud is therefore formed on the edge of the channel (Fig. 126). When

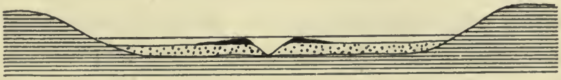


FIG. 126. Formation of natural embankments (shown black) during floods.

the flood ceases the water in the channel subsides, the water on the flooded plain partly escapes into the river and partly sinks into the ground; but the mud deposited at the edge of the channel remains, raising the river-banks above the level of the plain. The river, now reduced to its normal condition, again begins to deposit in its proper channel, raising the level of its bed. Consequently, although the banks are higher than before, their height above the *bed* may be no greater and the river will

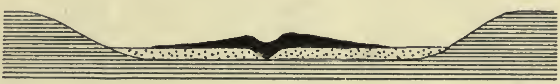


FIG. 127. River-bed raised by deposition above the level of the flood-plain.

not be less liable to overflow. If this goes on from year to year the banks may be raised to a considerable height and the bed of the river may lie above the level of the surrounding plain (Fig. 127).

The natural embankments which a river forms in this way for itself, improved and strengthened artificially, are known on the Mississippi and elsewhere as levees.

Examples of rivers which have raised themselves above the surrounding country are to be met with in many low-lying districts. Many of the rivers of the Fens are above the level of the land around; in the lower part of its course the Po is above the Plain of Lombardy; and for hundreds of miles the channel of the Hoang-ho stands high above the plain through which it flows.

But in all such cases, unless the natural embankments are strengthened artificially and are continually repaired, the river will not keep to a definite channel. In floods it is always liable to break its banks, to overspread the surrounding country and to forsake its old course for a new one. This has happened several times with the Hoang-ho. At one time it flowed into the Yellow Sea; but in the year 1852 it broke its banks, flooded an area of thousands of square miles, drowning, it is estimated, nearly a million people, and took a new course into

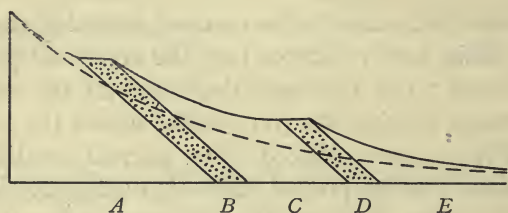


FIG. 128. Influence of differences of hardness on the curve of water-erosion.
A, C, E. Soft beds. *B, D.* Hard beds.

the Gulf of Pe-chi-li. The distance between its old mouth and its new one is nearly 300 miles. On several other occasions the Hoang-ho has caused similar disastrous floods and has changed its course.

Influence of differences of hardness upon the development of river-channels. In the account of the development of the curve of water-erosion it is assumed that the slope on which the river starts is made throughout of rock of uniform hardness. If, as is usually the case, it consists partly of soft beds and partly of hard beds, the form of the curve will be modified.

Suppose, for example, that upon the slope two beds crop out which are harder than the rest (Fig. 128). The river tries to form the simple curve of water-erosion shown by the dotted line. But because the beds *B* and *D* are hard, they are corraded less rapidly than they would be if they were as soft as the rest. The curve is therefore bulged up at the outcrops of these beds and becomes three separate curves instead of a single simple one.

Except for small inequalities the bed of a river must everywhere slope towards the sea. Therefore the soft bed *A* cannot be eroded below the level of the outcrop of the hard bed *B*. But since the soft rocks are more easily worn away, the slope of the river above the outcrop of each hard band decreases while below the outcrop it grows steeper. Over the soft beds the current will now be slow, over the hard beds there will be waterfalls or rapids. The increased speed of the river at the hard beds will, to some extent at least, compensate for their extra resistance, and the curves will gradually be cut down-

FIG. 129. Further development of the curve of water-erosion in hard and soft beds.

wards as shown in Fig. 129. As the slope near the sea diminishes, the area of deposition spreads upwards, and in time the hard beds will be buried beneath the material laid down by the river and will cease to show. But as long as a hard bed is in a part of the river where corrasion exceeds deposition, there will be rapids through it and a smoother reach above.

Moreover, since the harder beds will also offer more resistance to the weather, the sides of the valley will be steeper than in the softer beds. In the latter the valleys open out, in the former the river often flows through narrow and rocky gorges.

Waterfalls. If a hard bed dips gently down the stream as in Fig. 128 the river in passing over it will generally form rapids rather than waterfalls. But if the hard bed is horizontal, or dips gently up the stream, the river will wear away the softer rocks beneath it, the hard bed will overhang and a waterfall will be produced (Fig. 130). The falling water will continue to undermine the hard bed, and from time to time blocks of the latter will be broken off and the waterfall will gradually recede up-stream. This is what has happened in the case of the Niagara Falls. The top of the fall is formed of limestone, which rests upon a bed of shale. The fall was once at the front edge of the plateau shown in Fig. 131. It has gradually cut backward in the manner described, to its present position, and it is still receding. Below the fall is the gorge which the fall has cut, and the

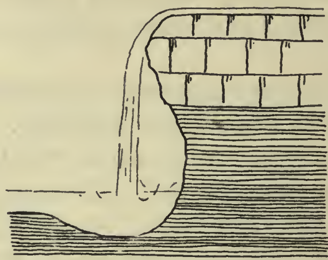


FIG. 130. Waterfall in horizontal strata.

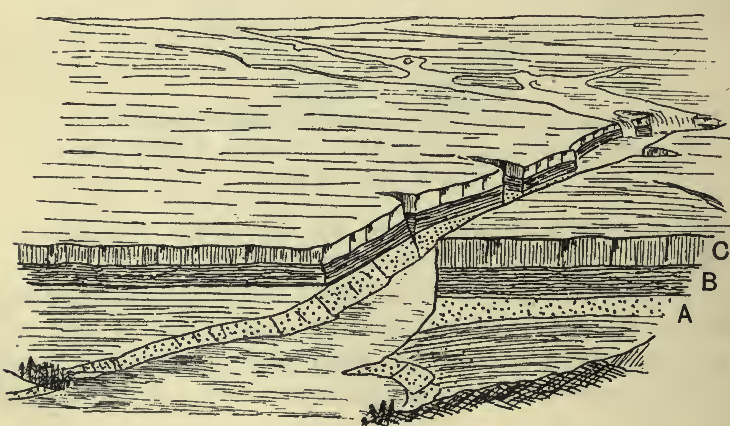


FIG. 131. Niagara falls and gorge. (After Lyell.)
A. Medina sandstone. *B*. Niagara shale and Clinton group.
C. Niagara limestone.

sides of the gorge consist of limestone above and shale and sandstone below. The backward erosion of the waterfall is comparatively rapid, because it depends upon the erosion of the soft rock at its base. The erosion of the sides of the valley below the fall is slow; it is due to the action of rain and other weathering agencies, and the hard rock above protects the softer rock beneath. Therefore the valley below the fall has steep sides and forms a gorge.

Whenever a waterfall cuts backwards there will usually be a gorge below it, for the backward erosion is generally rapid compared with the lateral erosion of the sides of the valley (Pl. X, fig. 1). But there are cases in which there is scarcely any backward erosion at all. If a hard bed, or an igneous dyke, runs vertically across the river, the soft rock on the down-stream side will be rapidly worn away and a waterfall will be formed (Fig. 132). But no undermining of the hard bed is possible. The river will gradually cut its channel deeper, both in the hard and the soft rock, and the waterfall will not alter its position.

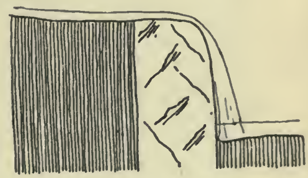


FIG. 132. Waterfall over a vertical hard bed.

Rejuvenation. A river in its Plain tract has ceased to corrade its bed, and as the curve of water-erosion becomes more and more developed the Plain tract extends farther and farther up the stream. In a long-established river, or, as it is often called, a mature river, vertical corrasion is accordingly limited to the tributary streams which form its head-waters, and throughout the main part of its course the river meanders through the plain that it has formed by lateral corrasion and deposition.

After this stage is reached, it sometimes happens that the volume or velocity of the river is in some way increased. The rainfall may become greater, or earth movements may give a greater slope to the country through which the river flows. Vertical corrasion will begin again and the river is said to be rejuvenated.

The channel will be deepened, and the flood-plain will now become a flat alluvial tract considerably above the level of the river and out of reach of the greatest flood. It will form a kind of platform overlooking the river, and such a platform is called a "river-terrace" (Fig. 133, see also Pl. X, fig. 2).

After equilibrium is restored and vertical corrasion ceases, the river will again corrade laterally and form a new flood-plain at a lower level than the old one. Possibly for some reason vertical corrasion may begin again and in this way a series of river-terraces may be formed one above another.

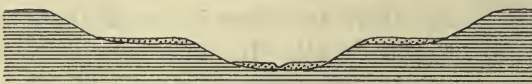


FIG. 133. River-terraces.

If the rejuvenation of a river is sufficient, the vertical corrasion may become so effective that the lateral corrasion is relatively unimportant. The river will then sink its channel deeply into the ground while still preserving all the turns and windings of its original course. In this way are formed the deep meandering valleys of the Wye below Ross, of the Dee at Llangollen, of the Wear at Durham, and of other rivers.

Meanders such as these, cut deeply into the rock, are called "incised meanders."

CHAPTER XXII

DEVELOPMENT OF RIVER-SYSTEMS

General principles. Up to the present we have considered only the changes that a single river produces in its own valley. But there may be many rivers upon the same slope and it often happens that the growth of one affects the development of others. It is with the changes which are thus produced that we are now concerned.

If a series of horizontal strata is bent into the form of an arch which rises above the sea, the arch, as soon as it appears above the water, will be worn by the waves. If the rate of elevation and the rate of erosion are properly adjusted, the curve of the arch will be planed off and the land will be like a low roof, with the two sides sloping gently from a central watershed. On each side the beds will dip in the same

Plate X

T. J. Roberts, photo.

Fig. 1. Waterfall and gorge.

Nant Ffrancon, North Wales.

P. Lake, photo.

Fig. 2. Low river-terrace.

Ceiriog valley, North Wales.

direction as the surface of the ground, but at a steeper angle (Fig. 134).

Upon each slope a system of rivers will be developed, and since the direction of the rivers is determined directly by the

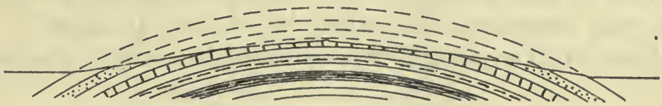


FIG. 134. A denuded anticline.

uplift which raised the arch, the rivers are said to be consequent on the uplift, and the whole forms a consequent drainage system.

The rivers will begin to flow almost as soon as the land

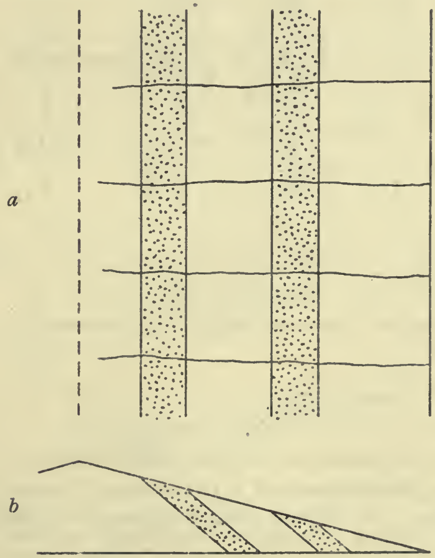


FIG. 135. Development of a river-system: first stage.
a. Plan. *b.* Section.

appears above the sea; but in order to show how they develop it will be simpler to suppose that the slope is completely formed before the rivers start.

Suppose that in the manner described an uplift produces a region sloping directly and uniformly from a straight watershed to the sea and that the outcrops of two hard beds run parallel to the watershed, the beds dipping in the same direction as the slope but at a steeper angle. Fig. 135 shows a plan and section of a part of such a region. For the present we need not consider what happens on the other side of the watershed.

Upon this slope a number of parallel rivers will be developed,

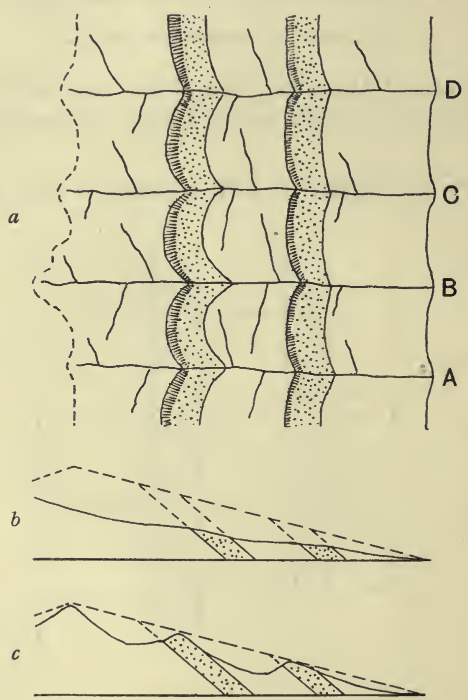


FIG. 136. Development of a river-system: second stage.
a. Plan. *b*. Section along a consequent stream.
c. Section between two consequent streams.

flowing directly from the watershed to the sea. These rivers are the immediate consequence of the uplift and are therefore known as consequent rivers.

As soon as the consequent rivers have excavated their valleys, water will enter these valleys from the sides, and tributaries will accordingly be formed, flowing more or less at right angles to the main streams. Such a direction is practically *along* the original slope, and is therefore impossible until the valleys of the consequent rivers have been carved out. The tributaries are accordingly called subsequent rivers,—they are formed after the consequent rivers (Fig. 136).

The consequent streams flow in the direction of the dip, and are sometimes known as dip streams. The subsequent streams run approximately in the direction of the strike and are sometimes called strike streams.

Because the soft beds are more easily eroded than the hard ones, the tributary valleys will be formed in the soft beds, and the subsequent streams will flow along the outcrops of these beds. The hard beds, away from the actual channels of the consequent streams, will not be directly affected by river-erosion and will therefore stand up as ridges overlooking the valleys of the subsequent streams. A section along the channel of a consequent stream is shown in Fig. 136 *b*, while Fig. 136 *c* is a section parallel to a consequent stream, but not along it. The projecting edge of a hard bed is called an escarpment.

From Fig. 136 *c* it will be seen that as the subsequent stream deepens its valley, a slope of soft rock will be formed beneath the edge of the hard bed and this slope will gradually grow steeper. The soft foundation on which the hard bed rests is weakened; the edge of the bed breaks into blocks which slide slowly or rapidly down the slope in front; and the escarpment gradually recedes.

Both the consequent and the subsequent streams will cut backwards at their heads. The former will eat into the main watershed, the latter into the watersheds between the consequent streams. If, as will usually be the case, one of the consequent rivers, *B*, is larger or swifter than the rest, it will erode its valley more deeply. Its tributaries will therefore have a steeper fall than the tributaries of the other consequent streams, and the erosion at their heads will be more rapid. In time their valleys will be cut back into the valley of the neighbouring consequent river, and the headwaters of the latter will find their way into the larger river *B* (Fig. 137). This process is known as beheading or river-capture.

The lower part of the beheaded river, fed by the tributaries below the point of capture, will continue as a stream of diminished size as in the case of *D*. Before long, however, the increased erosion of the subsequent stream *a*, strengthened by its capture of the headwaters of *C* and *D*, will lower the level of the soft bed in which it flows, to so great an extent that all the water on its own side of the escarpment will run into it. In the figure this has already happened in the case of the rivers *A* and *C*. *Obsequent* streams, or streams flowing in the opposite direction to the consequent rivers, have been formed at *x*, *x*, in the original valleys of the beheaded rivers. The lower parts

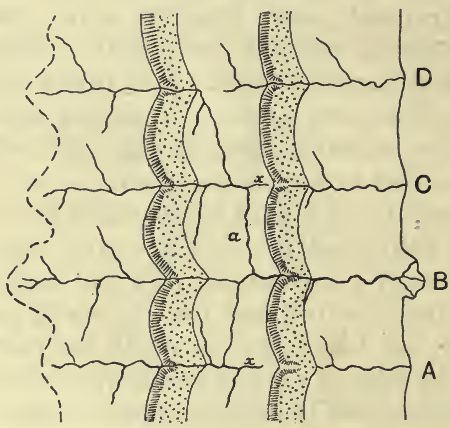


FIG. 137. Development of a river-system: third stage.

of *A* and *C* now take their rise on the other side of the escarpment and the gaps through which they originally flowed have become dry, forming what are known as wind-gaps. The valleys of the diminished streams *A* and *C* were made for a larger flow of water. They seem too large for their present streams and are accordingly often termed misfits.

If there is no interference by earth movements or other causes, the largest of the original consequent rivers may capture in turn the headwaters of all the rest. The tributaries which effect the capture may become considerably larger than the original consequent river into which they flow, and the latter may be by comparison an almost insignificant stream down to its junction with its tributaries.

A consequent river may even capture the headwaters of a river on the other side of the main watershed. If the slopes on the two sides of the watershed are unequal, the erosion at the heads of the rivers will be more rapid on the steeper side. The valleys on this side will gradually eat their way backwards through the watershed, and in time will tap the upper waters of the rivers on the gentler slope.

The rivers of Northumberland. The rivers of Northumberland afford a good example of the development of a river-system in the manner described (Fig. 138).

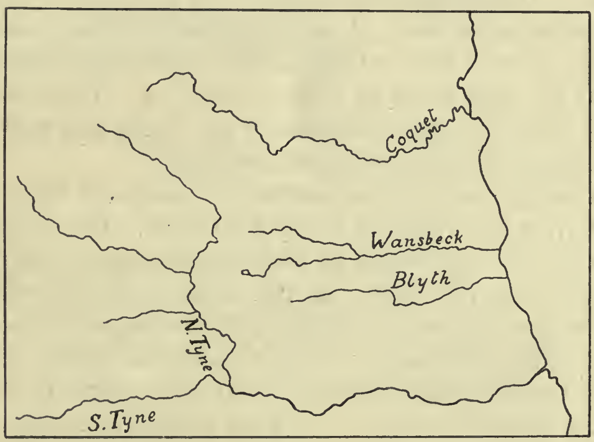


FIG. 138. The rivers of Northumberland.

To the west lie the Cheviots and the Pennines, forming the watershed between the North and the Irish Seas ; and from the watershed there is a general slope towards the east. Upon this slope flow the Coquet, the Wansbeck, the Blyth and the Tyne. The Coquet and the branch of the Tyne known as the South Tyne, rise near the watershed and flow directly to the sea. But the Wansbeck and the Blyth have their sources on the middle of the general slope.

The Tyne receives a tributary from the north, called the North Tyne, and on its western side the North Tyne receives tributaries coming straight down from the watershed. One of these tributaries if it continued its course would enter the valley of the Wansbeck, another would flow into the Blyth.

Originally the four rivers, the Coquet, the Wansbeck, the Blyth and the Tyne, flowed straight from the watershed to the sea. The Tyne was the largest of these rivers and its tributary, the North Tyne, cut backwards until it had captured the head-waters of the Blyth and the Wansbeck. It has not yet reached the Coquet, but in time this river too will be beheaded.

The Humber. The Humber is really the mouth of the Aire, but the tributaries, the Ouse on the north and the Trent on the south, are now more important than the Aire.

The Pennine Chain lies to the west and there is a general slope towards the east. A large number of rivers start in the Pennines close to the watershed and begin to flow eastwards; but only the Aire keeps its course to the sea. The Calder joins the Aire, the rest enter the Ouse or the Trent and their waters are deflected from their original direction.

Originally there were a number of consequent rivers flowing directly from the Pennines to the North Sea. One of these was the Aire. In a soft band of rock its tributaries, the Ouse on the north and the Trent on the south, cut backwards and captured the headwaters of the other consequent streams.

In Northumberland it is easy to trace the former continuations of the beheaded streams, in Yorkshire and Lincolnshire it is more difficult to do so. In East Yorkshire, however, there is a large valley, the Vale of Pickering, which appears to have been the valley of the Swale and Ure when those rivers flowed directly to the sea. The present Witham was probably the continuation of one of the original consequent streams in the south. It now rises in low ground and flows through a gap in the Lincoln Hills; and it is said that during floods some of the water of the upper Trent still finds its way into the Witham.

The rivers of the Weald. The Weald (Fig. 139) is the area that lies between the North Downs and the South Downs, stretching westwards from the high land of Salisbury Plain to the sea.

The Downs and Salisbury Plain are formed by the outcrop of the chalk, which is comparatively hard and therefore forms high ground, with an escarpment overlooking the Weald. In the middle of the Weald there is an upland area, formed by the outcrop of a series of sandstones, and between this and the escarpment of the chalk is a broad depression. On a map the

FIG. 139. Map of the Weald.
Dotted areas, sandstone. Rectangular ruling, chalk.
White areas (on the land), clay.

depression is U-shaped, with the bend of the U lying against the chalk of Salisbury Plain and the open end of the U against the sea. This depression is hollowed out in a bed of clay.

Some of the rivers rise in the central area of high land, others in the depression itself; but almost all have cut their

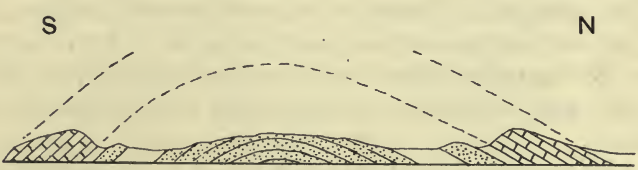


FIG. 140. Section of the Weald.

valleys through the Downs, although the Downs are higher than the sources of the rivers.

At first sight it is difficult to understand how a river can cut its way through hills which are higher than its source; but the difficulty is explained by the geological structure. Fig. 140 is a section across the Weald from north to south. From this it is evident that the beds are in the form of an arch or elongated dome, and at one time the chalk extended over the whole. The rivers rose at the top of the arch and ran down towards the north and the south. Gradually these rivers and their tributaries removed the chalk from the whole of the central area, exposing the beds beneath; but during the process they still kept their original directions. Except in the actual courses of the rivers, the chalk resisted erosion much more strongly than the clay and stands up as a ridge, through which the valleys are cut. The ridge is higher than the present sources of the streams but it is not higher than their sources originally were.

Antecedent drainage. In the cases already described the river-system or system of drainage is the result of the earth movements to which the tilting of the surface and of the beds is due. It is therefore called consequent drainage, and in such drainage the direction of the rivers is closely related to the geological structure.

But it may happen that earth movements occur after the drainage system is established, and if the earth movements take place so slowly that erosion proceeds as fast or faster than the uplift, the rivers may continue to keep their original courses and in time may show no relation to the geological structures developed by the earth movements. Such drainage is said to be antecedent.

If for example an anticline is raised across the line of the river, but so slowly that the river erodes its bed downward as fast as the anticline rises, the direction of the river will be unaltered and it will flow through the ridge formed by the anticline.

It has been suggested that this is why the Indus and the Brahmaputra break through the chain of the Himalayas. They may have existed previously to the elevation of the chain and have simply kept their original courses. But another explanation is possible. They may have begun upon the southern slopes, after the mountains were raised, and may have cut their way back through the watershed in the manner described on p. 240.

It is in fact very difficult to prove a case of antecedent drainage. Usually some other explanation is also possible.

Superimposed drainage. There is another way in which the direction of the rivers may become independent of the geological structure of the country through which they flow. The courses of the rivers may be determined by an uplift and may be related in the usual manner to the structure. But in time their valleys may be cut down into an older series of rocks which have been affected by an earlier system of folding quite independent of the later movements to which the rivers are due.

This is the case in the Lake District. The older rocks which form the greater part of the region were thrown into a series of folds running from east-north-east to west-south-west, the oldest rocks of all appearing in the northern part of the area. The rivers rise near the middle of the Lake District and

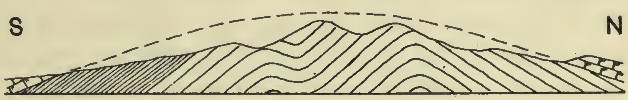


FIG. 141. Section of the Lake District.
The broken line represents the base of the later beds which once covered the whole district but which are now left only round its edge.

radiate outwards in all directions, quite independently of the arrangement of the beds on which they flow. But the older rocks of the Lake District are completely surrounded by a ring of later beds. Originally these later beds covered the whole area, forming a dome with its centre about the present position of Helvellyn (Fig. 141). The rivers started near the top of the dome and flowed outwards in all directions. Erosion has gradually removed the later beds and exposed the older folded rocks beneath. The courses of the rivers were determined by the dome-like form of the later beds upon which they began to flow, and as they cut their valleys downwards they have kept, in general, to their original directions, in spite of the different structure of the rocks into which they cut.

Such a system of drainage as this is said to be superimposed or superinduced. It has been imposed upon the district not by the rocks which now form the region, but by the later beds which once spread over them.

General result of erosion by running water. In the previous chapter it was shown that in every part of its course a river attempts to adjust the slope of its channel so that the velocity of the current is just sufficient to enable it to carry the load that enters that part. When this condition is attained, corrasion and deposition are equal and the river is said to be graded.

It is possible for the adjustment in a single reach to be perfect; but if any load is brought into this reach it implies that above the reach corrasion is greater than deposition and that the adjustment there is not complete. It is conceivable that in time the channel might be graded throughout. Everywhere corrasion would then be equal to deposition, no load would pass from any part of the channel into the part below and the river would cease to carry any material downwards. But it is evident that this condition cannot be reached until the velocity of the river is so low that it is unable to move even the loose particles on its banks and bed. Until this condition is reached there must be some corrasion.

It is the same with running water generally, even when it is not confined to definite channels, but flows as a sheet, as sometimes happens during a thunderstorm.

Running water therefore wears away the land until it has produced a slope so gentle that its velocity down the slope is not sufficient to move the particles produced by weathering or corrasion. Any projection above this slope will be eroded, any depression below it will be filled up by deposition.

Owing to earth movements, variations in the rainfall and other causes, such a condition is never actually attained; but erosion by rain and rivers often produces an approximation to it. The surface of the ground will then be nearly a plain and is called a peneplain. Often the harder rocks, not yet completely worn down, still stand up as hills in the midst of the plain-like islands in the sea. But, given time, they too would be eroded and reduced to the common level.

CHAPTER XXIII

UNDERGROUND WATER

Origin of underground water. Some of the rain that falls upon the ground sinks into the rocks beneath. The proportion that disappears in this way depends to a large extent upon the nature of the rock. Chalk for example is porous and allows the water to percolate through it readily. Even a firm and solid rock like granite, which in itself is practically impervious, often has open joints to a considerable depth and the water can penetrate along them. Clay is almost impervious, and since it is usually free from cracks, a bed of clay generally forms a very effective barrier to the movement of the water underground.

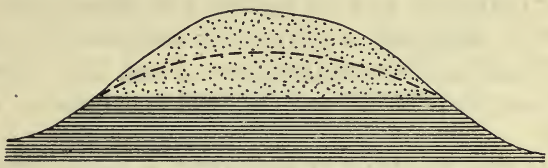


FIG. 142. Level of saturation.

There is reason to believe that water is often produced by chemical changes in the heated interior of the earth, and in volcanic districts and elsewhere this water may rise to the surface. But in a country such as ours most of the underground water is simply rain which has percolated from above.

This underground water is the source from which springs and wells draw their supplies, and a simple case will serve as an illustration.

Level of saturation. Fig. 142 represents a hill formed of a horizontal bed of clay capped by a layer of porous sandstone. When rain falls, it percolates through the sandstone, but cannot penetrate the clay. If the rain is long continued, all the pores of the sandstone may be filled with water and the bed of sandstone will be saturated. When the rain ceases the sandstone holds the water like a sponge, but like a sponge it allows it to trickle out slowly at its base and sides. The water therefore will issue from the hill-side at the junction of the clay and sandstone, forming a spring or a series of springs. But as the water at the base of the sandstone oozes out the water at the top sinks slowly downwards, and the sandstone dries from the top downwards. The sandstone near the outlet is drained most easily, and accordingly the surface of the water in the sandstone takes the form shown by the broken line in the figure. Below this line the sandstone is still saturated, immediately above the line it is damp, and higher up it is practically dry.

The level below which the rock is saturated is called the "level of saturation" or the "water-table."

If the weather remains dry the level of saturation will still continue to fall, until there is not sufficient pressure to force the water outwards, and the springs will then dry up. But if the layer of sandstone is sufficiently extensive the supply of water will last until the next shower of rain falls; some of the rain will sink into the hill and the level of saturation will again be raised.

In this way the level of saturation is continually varying according to the weather. In a dry season it falls, in a wet season it rises. But usually there is a limit below which it never sinks and this is called the permanent level of saturation. There will also be a limit above which it never rises. This may be the surface of the ground itself, but more usually it lies some distance beneath. In the particular case illustrated in Fig. 142 the water can escape so freely at the sides of the hill that, even in the wettest seasons, it is not likely that the whole of the sandstone will be saturated.

The fluctuations in the level of saturation cause corresponding fluctuations in the supply of water in wells and springs.

If a well is sunk to the wet season level but not to the permanent level of saturation, it will have water in it in a wet season but not in a dry one. If it reaches the permanent level there will always be water in the well.

In Fig. 143, in which the cap of sandstone slants to the left, the springs on the left slope of the hill will be permanent because they lie below the level of permanent saturation ; the springs on the right slope will flow only during a wet season

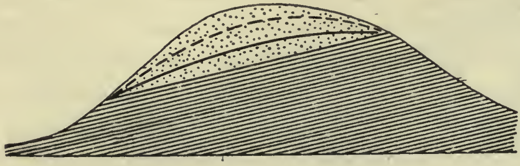


FIG. 143. Permanent and intermittent springs.
The full line shows the permanent level of saturation ; the broken line shows the wet-season level.

and are called intermittent springs. They lie above the level of permanent saturation but below the wet season level.

Springs are usually at the junction of a pervious bed with an impervious one. It may be that the pervious bed lies on

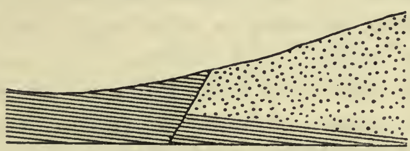


FIG. 144. Springs due to a fault.

the impervious one as in the examples already given ; but the two beds may be brought into contact by a fault as in Fig. 144. In such a case as this there will be a spring or a line of springs at the fault, because this is the lowest point at which the water in the pervious bed can escape.

FIG. 145. Artesian well.

Artesian wells. If a porous bed like chalk lying between two beds of clay is bent into the form of a basin as in the London district (Fig. 145) the rain falling on the outcrop of the chalk will sink downwards. There is no escape for it below, and the chalk will therefore be saturated with water to the rim of the basin. The level of saturation is shown by the broken lines. There will be springs around the edge, but only the water above the edge can escape. If a well is sunk through the upper layer of clay into the chalk, the water in the chalk will flow into the well. If the chalk offered no resistance to its movement it would tend to rise to the level of the rim of the basin; but the chalk is an obstruction and it does not therefore rise so high. If, however, the level of saturation at the outcrop of the chalk is sufficiently far above the well, the water in the well will rise above the ground and flow out over the surface or even form a fountain. Such a well as this is called an artesian well.

Underground water in limestone districts. The water which percolates through the rocks dissolves some of their constituents. In most cases the amount dissolved is small, the water is soon saturated, or holds as much as it can, and no further solution takes place. But ordinary rainwater, containing carbon dioxide taken up from the air, can dissolve a large quantity of carbonate of lime. In passing through a bed of limestone or chalk, therefore, the percolating water removes a considerable amount of material in solution, and the effects produced are often of great importance.

The water naturally takes the easiest route and flows through cracks, where they are present, rather than through the smaller pores and openings. The cracks, which are usually along joint-planes, are gradually enlarged by solution and become still easier channels than before (Pl. XI, fig. 2). Gradually the water concentrates in these channels, and in a limestone the underground water often forms streams of considerable size (Fig. 146).

A crack at the surface may in time be enlarged so much that it forms a wide opening sinking deeply into the ground. Such openings are known as swallow-holes and are common in limestone districts. Occasionally a surface stream of considerable size falls into a swallow-hole and disappears. Gaping Ghyll on the side of Ingleborough is a well-known example.

P. Lake, photo.

Fig. 1. High Tor, Matlock.

A limestone crag overlooking the Derwent.

S. H. Reynolds, photo.

Fig. 2. Eroded surface of limestone.

Ben Suardal, Broadford, Skye.

The channels beneath the surface are often widened to so great an extent that they form a series of caverns running for miles into the ground. In Derbyshire and other limestone districts such caves are common.

The roof and floor of the cave are frequently covered with stalactites and stalagmites. A stalactite is a column of carbonate of lime hanging downwards from the roof. The water which trickles through the cracks in the ceiling contains carbonate of lime in solution. While a drop is hanging, it partly evaporates and leaves a little carbonate of lime behind. The next drop leaves a little more and so a small lump is formed and gradually grows downwards, the water tending to trickle down its side and hang in drops at the lower end.

Stalagmites are similar lumps formed on the floor of the cave by the water that falls from above. Naturally they grow

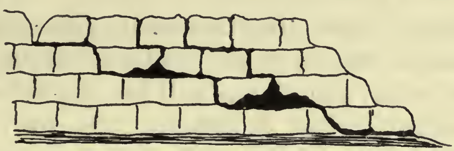


FIG. 146. Underground streams and caverns in limestone.

upwards instead of downwards and they are usually shorter and thicker than the stalactites. Often a stalagmite is produced by the water that drips from a stalactite, and sometimes the stalagmite and stalactite meet, forming a complete pillar from the floor to the roof of the cave.

The water that trickles down the side of the cave or over the bottom, deposits its carbonate of lime in layers encrusting the walls and floor. This deposit is also often known as stalagmite.

Caverns in limestone are usually developed along the joints, and often therefore they are long compared with their width. It sometimes happens that the roof falls in and the cavern is then converted into an open gorge, or perhaps simply an irregular cavity in the ground.

Although chalk is carbonate of lime, caverns and underground channels are not formed in chalk to so great an extent as, for example, in the limestones of the Pennines. The reason appears to be that the jointing in chalk is much less marked, and the rock itself is much more porous than the harder and more compact limestones. The water therefore percolates more readily through the mass of the rock and its action is not so much confined to the planes of the joints.

Special characters of limestone districts. Since a large proportion of the rainfall sinks immediately into the rocks and runs in channels underground, limestone districts usually have a special character of their own.

The higher land is always dry. Few streams flow upon its surface and even these may suddenly sink into the ground and disappear. Lower down the slopes the stream may reappear as suddenly from an opening in the rocks.

The valleys are often narrow, and their sides are usually steep, frequently forming precipitous cliffs ; for limestone is a firm and compact rock, and, though it is readily dissolved, it resists mechanical erosion by the weather. Moreover, since the rain sinks so quickly into the rock, it has but little time to wear away the surface. Most limestones, too, have well-marked joints and the faces of the cliffs are therefore abrupt and cleanly cut (Pl. XI, fig. 1).

The carbonate of lime is carried away in solution and only the insoluble impurities are left to form a covering to the rock. Therefore the soil is usually thin, and limestone districts are often very bare, excepting in the valleys, where the soil may collect to a greater thickness.

These are the general features of the "Karst" type of country, so named from the Karst district of Austria where they are developed to an exaggerated extent. They may also be seen in the Pennines and other parts of England, but they are not there so strongly marked.

Chalk districts show some of these characters, but not all. Chalk is made of carbonate of lime, but it differs from an ordinary limestone in several respects. It is softer and more porous, and generally it is not so strongly jointed. Chalk hills are as dry as limestone hills, and the soil upon them is often as thin ; but because the rock is soft and easily worn away, the valleys are wide and open and show no crags upon their gently sloping sides. Except on the coast, where the sea eats continually into its base, chalk seldom forms precipitous cliffs.

CHAPTER XXIV

SNOW AND ICE

Frost and snow. When water enters a narrow cleft in a rock and afterwards freezes, its expansion on changing into ice tends to widen the cleft. If this happens often, a part of the rock may be broken off. The consequence is that in mountain districts projecting crags are gradually broken up, and the fragments are scattered on the mountain side, often forming screes (Pl. IX).

This kind of action goes on only in places where the temperature is sometimes below the freezing point and sometimes above, and the more rapid the alternation the more rapid is the disintegration of the rock. Where, on the other hand, the temperature is always above freezing point the water never becomes ice, and where it is always below the freezing point the ice never becomes water. It is accordingly on the borders of the Polar regions and towards the tops of mountains in other parts of the world that the effect is greatest, and it is there so marked that in some places it is almost impossible to find a projecting point of rock firm enough to afford a hold in climbing.

Snow itself as it lies upon the ground is neither a disintegrating nor a corrasive agent. It is in fact a protection to the rocks beneath. It is a bad conductor of heat, and acts as a blanket, preserving the ground from changes of temperature, and especially from the destructive effects of alternate freezing and thawing. To a certain extent, also, it protects the rocks from the action of rain. The rain may melt the upper layers of snow, and running water may flow over the surface ; but if the covering of snow is thick enough, the water will be frozen before it can sink to the rocks beneath.

Resting snow is thus a preservative agent. But in certain circumstances it begins to move and it then becomes an agent of erosion.

In the summer the snow melts even on the top of a high mountain, but above the snow-line the total loss by melting and evaporation during the year is less than the amount that falls. If there were no other means of escape the thickness of the snow would accordingly increase indefinitely. But the snow is got rid of in two other ways. If the slope of the ground is sufficiently steep, when the snow becomes deep its own weight makes it slide down rapidly as an avalanche.

Glaciers. If the slope is less the sliding still takes place, but slowly and in a different way. A thicker accumulation of snow can be formed on such a slope. The weight of the upper layers compresses the lower layers into a kind of granular ice and presses them outwards so that they begin to flow slowly down the hill. As in the case of water, the flow is naturally concentrated in the valleys; and from the snow-field on the top of a ridge tongues of moving ice flow slowly down the valleys as glaciers.

A glacier may extend far below the snow-line. At its origin the supply of ice is comparatively rapid and the melting slow, and the glacier is broad and deep. As it flows downwards it encounters higher and higher temperatures, and the supply of ice diminishes because some of it has melted higher up. The glacier therefore dwindles in size until it reaches the limit where the supply of ice from above is just equalled by the melting, and there the glacier ends. Glaciers accordingly are usually tongue-shaped, broadest above and narrowest below. A glacier moves most rapidly in the middle, and in the middle therefore the supply of ice is greatest. Consequently the termination is convex.

Rate of movement. The movement of a glacier is most rapid in the middle because at the sides and bottom it is retarded by friction against its bed. If a row of stakes is driven in a straight line across a glacier, they will travel downwards with the glacier; but in the course of a year or two the

T. D. La Touche, *photo.*

A Himalayan glacier, showing crevasses and ice-falls.

row will become curved, with the convexity facing downwards (Fig. 147).

In the Mer de Glace J. D. Forbes found that in summer and autumn the rate of movement was 20 to 27 inches per day near the centre, 13 to 19½ inches near the side. In Greenland the motion is often much more rapid and in one place a rate of 100 feet per day has been observed.

FIG. 147. Movement of a glacier. The small circles indicate the original position of the stakes; the black dots indicate their position after the lapse of a few years.

Crevasses and ice-falls. If the valley of the glacier is uniform in width and has a smooth and regular floor, the surface of the ice is usually even and unbroken. But if the slope of the floor changes, or the valley becomes contracted, the rate of movement is no longer uniform and the surface becomes broken and irregular, just as a stream becomes rough in similar circumstances. If, for example, the slope increases as in Fig. 148 the glacier moves more rapidly on the steeper part, and

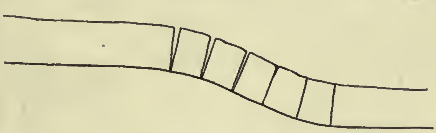


FIG. 148. Formation of crevasses and ice-falls.

where the change of slope begins, a vertical crack or crevasse forms across it. Down the steep slope the ice is greatly broken, forming what is known as an ice-fall (Pl. XII). At the foot, where the slope again decreases, the rate of movement diminishes and the cracks begin to close up.

Crevasses formed in this way are transverse to the glacier, and since they move more rapidly in the middle than at the sides, they become curved like the row of stakes in Fig. 147.

But crevasses are not always transverse. If the valley suddenly widens, the glacier in spreading out may develop cracks which are more or less longitudinal in direction.

Even the difference between the velocity in the middle and the velocity at the sides tends to produce cracks. It is evident from Fig. 147 that as the row of stakes moves downwards and becomes curved the distance between the stakes increases, especially at the sides. The ice is stretched along the line of stakes, and cracks may be formed across the line obliquely to the course of the glacier.

Moraines. The crags which overlook the glacier are broken up by frost and other agencies and blocks of rock fall from them on the ice below, and are slowly carried downwards. From every projecting bluff accordingly a stream of blocks stretches down the side of the glacier, and these streams of broken rock uniting with one another form the lateral moraines, one upon each side (Fig. 149 and Pl. XIII, fig. 1). When two glaciers meet, a lateral moraine of one glacier unites with a lateral moraine of the other forming a medial moraine. If other tributary glaciers join them subsequently, there may be several medial moraines (Fig. 150).

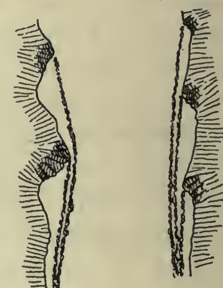


FIG. 149. Lateral moraines.

A block of rock resting on a glacier protects the ice beneath it, while all around the surface may be melted by the sun. The block will then stand on a pedestal of ice; but the pedestal itself will slowly melt and sooner or later the block will fall to one side or another. In this way the material of the moraines is slowly spread out and sometimes covers almost the whole width of the glacier.


FIG. 150. Lateral, medial and terminal moraines.

T. D. La Touche, photo.

Fig. 1. A Himalayan glacier, showing moraines.
Umasi La, Kashmir.

S. H. Reynolds, photo.

Fig. 2. End of the Suphelle glacier, with terminal moraine.
Fjaerland, Norway.

When the glacier melts at its termination all this material is dropped and a mound is formed, called the terminal moraine (Pl. XIII, fig. 2). It stretches across the valley like a dam, and since the end of the glacier is rounded it is usually crescentic in shape.

The boulders which fall upon the glacier do not all travel downwards on its surface. Often they are engulfed in the crevasses which open beneath them, and they finish their journey either in the middle of the ice or at its base. The streams formed by the melting of the surface also frequently fall into a crevasse, carrying with them the finer material of the superficial moraines. Moreover, the moving ice itself breaks off fragments from its bed. Thus a glacier carries a great deal of material in its midst and at its base and sides as well as on its surface.

The material frozen into the sides and base of the glacier is dragged along by the movement of the ice and rasps the sides and floor of the valley. The finer material acts like sand-paper, smoothing and rounding the surface of the rocks over which it rubs; but the larger fragments cut deep grooves. The fragments themselves are ground down in the process and consequently the boulders carried in the foot of the ice show smoothed or even polished faces, marked by scratches or striations. Sometimes only one side is affected, but in the course of its journey downwards a boulder will usually turn over more or less, and several surfaces will be ground down. A glacial boulder, however, is not rounded like a pebble in a river. The ice into which it is frozen prevents it from turning freely. Usually it is dragged along with its longer axis parallel to the direction of movement; the ends are left comparatively rough, the angles are rounded off, and the sides are smoothed, but with scratches more or less parallel to its length.

At the end of the glacier all this material is set free by the melting of the ice. Some of it is added to the terminal moraine, some is carried away by the glacier stream. There is always a stream flowing from the end of the melting glacier. Generally it emerges from a tunnel beneath the ice. Pressure lowers the freezing point of water, and thus the weight of the ice above assists in melting the ice at the bottom of the glacier.

The glacier stream is abundantly charged with fine mud, produced by the grinding action of the boulders at the bottom and sides of the glacier.

Piedmont glaciers. In the Alps the glaciers terminate before they reach the foot of the mountains. But in colder climates they may flow out over the plains beneath. Sometimes several glaciers unite at the base of a mountain range and form an extensive sheet of ice over the low-lying ground. Such a sheet is called a piedmont glacier.

The width of a piedmont glacier is usually greater than the united widths of the glaciers which form it, and the rate of movement is therefore slow, for the same reason that a river slackens when its channel widens. Sometimes indeed the movement practically ceases; and trees have time to grow upon the moraines that cover the surface of the motionless ice.

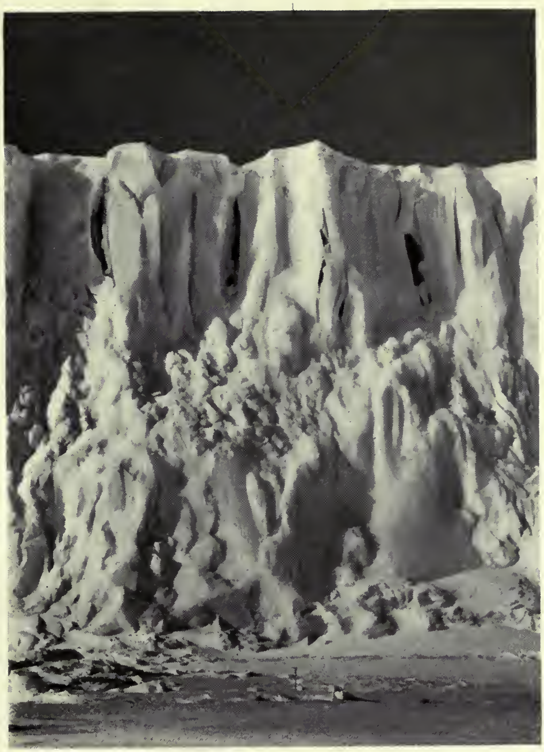
The Malaspina glacier in Alaska is one of the best-known examples of a piedmont glacier.

Ice-sheets. When the winter snowfall is sufficiently in excess of the loss by melting during the summer, the snow may accumulate to so great an extent as to bury plains and mountains indiscriminately. There will not then be separate glaciers, each in its own valley, but one continuous sheet of snow and ice, which will flow outwards from its highest point.

This is the case in Greenland, which is almost completely covered by such a sheet. Near the edge the peaks of some of the buried mountains pierce the covering, and close to the sea the ice divides into separate streams with rocky ridges between. But in the interior there is nothing visible but snow and ice.

The greater part of the ice-sheet is free from moraines because no rock rises above its surface, but towards its margins, where projecting peaks appear, moraines are formed in the usual way.

The Antarctic land-mass is also covered by an ice-sheet; but owing to the greater irregularity of the surface and probably in part to the smaller snow-fall, the higher hills are not completely buried (Pl. XIV).



C. S. Wright, photo.

Fig. 1. Antarctic Ice.
Terminal face of the Barne Glacier, Ross Island.

C. S. Wright, photo.

Fig. 2. Antarctic Ice.
The Tryggve Glacier, Ross Island.

Ice-bergs. When a glacier enters the sea the ice tends to float upon the water, and the end of the glacier is buoyed up. It is accordingly easily broken by the waves, and the mass of ice floats off. Occasionally the glacier reaches a cliff overlooking the sea. The ice will still move on, and at intervals the overhanging part will break and fall.

It is in these ways that ice-bergs are formed. Since the specific gravity of ice is about eight-ninths of that of sea-water, about eight-ninths of the ice-berg is below the water and only one-ninth above.

Ice-bergs bear away with them the boulders and mud carried by the glacier, and when they melt they deposit the material upon the bed of the sea. The Newfoundland banks have probably been formed partly in this way; and it is possible that the great width of the continental shelf in the North Atlantic may be due to the same cause.

Characteristic features of a glaciated region. The Alpine glaciers were once more extensive than they are at present, and even in the last hundred years they have sometimes advanced and sometimes retreated. The parts of the valleys which have been beneath the moving ice show certain peculiar features which are easily recognised and which are evidently due to the action of the glacier. Similar features are to be seen in our mountain districts, proving that glaciers once occupied the valleys.

The floor and sides of the valley are smoothed, and prominent crags are rounded, but the smoothed and rounded surfaces are marked by grooves and striations in the direction in which the glacier flowed (Pl. XV, fig. 2). Projecting spurs are sometimes sharply truncated and the valley is straightened (Fig. 151).

When a little hill of rock rises up in the path of the glacier it is not usually worn away completely. The side which faces up the valley is smoothed and striated, while the opposite side remains rough and rugged. The side against which the glacier flows is ground down by the stones embedded in the ice, but on the downstream side the glacier pulls away any blocks which may be sufficiently loosened by the development of joints.

Such hillocks of rock, rounded on the upstream side and rough on the other, are called *roches moutonnées* and are to be seen in almost any glaciated valley (Pl. XV, fig. 1). Owing to their form the appearance of the valley sometimes differs greatly according to the point of view. Looking down the valley we see the smoothed surfaces, and all the little inequalities are rounded and convex. Looking up the valley it is the rough sides that we see and the general effect is one of ruggedness.

An ordinary glacier does not fill the valley in which it flows, and above the surface of the ice the sides of the valley are worn by other agents. Frost usually plays an important part in the erosion, and rain and streams assist. Consequently, when the glacier disappears, the characteristic smoothing and

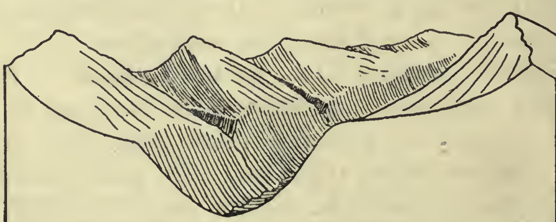


FIG. 151. Truncated spurs and hanging valleys.

striation reach only to a certain height. Above the former level of the ice the sides of the valley are often very rough, with rugged crags and deep gullies carved by mountain torrents.

Loose blocks of rock, often of enormous size, are scattered at intervals along the floor and sides of the valley, sometimes in very insecure positions. Some of these may have fallen from the crags above; but others are made of a different kind of rock, and have been brought down by the glacier. Occasionally they form a definite line upon the hill-side, corresponding with a former lateral moraine. But since the glacier has usually dwindled away very gradually, the position of its lateral moraine varies, and the blocks are therefore more often irregularly scattered.

The terminal moraine is often visible, forming a crescent-shaped mound across the valley. Sometimes there are several terminal moraines, each of them marking a pause in the gradual retreat of the glacier.

In places the bottom of the valley may be filled with boulder-clay, a stiff clay with striated boulders of various sizes scattered through it. This appears to have been the material at the bottom of the glacier, deposited in places where the movement of the ice was arrested, in somewhat the same way as the sand or mud brought down by a river may be deposited in a sheltered hollow.

Lakes are a common and characteristic feature of glaciated valleys. Sometimes they have been formed by the blocking of the valley by a mass of boulder clay. Occasionally a terminal moraine serves as a dam to hold the water up; but such a dam, being made up largely of loose blocks, is not always

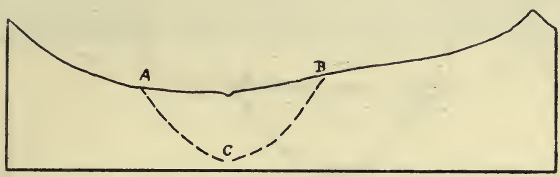


FIG. 152. Formation of hanging valleys by glacial erosion.

watertight, and is, moreover, liable to be breached by the overflow from the lake. In some cases at least, the glacier has scooped out a hollow in the solid floor of the valley, and the lake lies in a true rock-basin, and is not simply a part of the valley dammed up by glacial deposits.

In course of time the lake may be filled up by peat and silt, or the outflowing river may cut its channel deep enough to drain off the water. There will then be left a peaty flat, through which a stream meanders to the former outlet.

In mountain districts the valley of a tributary stream often opens high up the side of the main valley, and below the opening the tributary forms a series of rapids and waterfalls down the slope. Such tributary valleys as these are known as "hanging valleys" (Fig. 151).

Hanging valleys are especially abundant in glaciated regions and apparently therefore glaciers have assisted in their formation. The obvious explanation is that the main valley has been eroded so rapidly that the tributary was unable to keep pace with it. Most authorities attribute the extra erosion to the glacier that formerly occupied the valley. Originally the bed of the main river was at the level shown in Fig. 152. The glacier has eroded the part *ABC*, and the tributaries now enter the main valley at the level of *A* and *B*.

But some observers believe that glaciers are very ineffective as agents of erosion, and if this be true the formation of hanging valleys must be explained in a different way. When a valley is filled by a glacier up to the level *AB* (Fig. 153) the sides of the valley beneath the glacier are protected from water erosion, but above the ice they are exposed. Streams will be formed

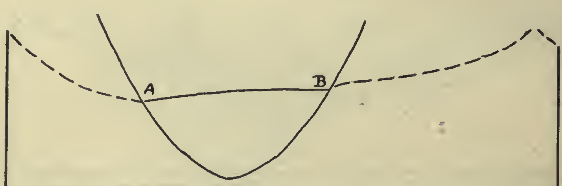


FIG. 153. Formation of hanging valleys by glacial protection.

on the sides of the valley and will cut backwards producing valleys (as indicated by the broken lines) in the ordinary way. But these tributary valleys cannot be cut below the level of the ice in the main valley, and therefore they open out at that level. When the glacier disappears they will be hanging valleys.

Hanging valleys, however, are not confined to glaciated districts and do not necessarily require glaciers for their formation. If for any reason the main valley is eroded more rapidly than the tributary valleys, the latter will "hang" to a greater or less extent. If for example the main stream runs along a line of fault its erosion may be very rapid. Moreover, when a waterfall cuts its way backwards and leaves a gorge, the tributaries below the waterfall will enter the gorge at a high level, and for a time at least their valleys will be hanging valleys.

P. Lake, photo.

Fig. 1. Roche moutonnée.
Nant Ffrancon, North Wales.

P. Lake, photo.

Fig. 2. Glaciated rock-surface.
Nant Ffrancon, North Wales.

Glaciated lowlands. In the Scottish Highlands, in the Lake District and in Wales all these characteristic features of a glaciated region are visible, and there can be no doubt that most of the valleys were formerly occupied by glaciers. But even in the lowlands there is evidence of the action of ice. Wide tracts of country are covered with boulder-clay. Here and there large blocks of rock lie on the ground, and these blocks are often entirely different from any of the rocks in the neighbourhood and have evidently come from a distance. In Yorkshire, for example, there are many boulders of granite, precisely similar to the granite of Shap in Cumberland. Such far-travelled masses of rock are called erratic blocks.

According to the view which is most widely held, the whole of Northern Europe, down to the latitude of the Bristol Channel, was at one time covered by an almost continuous ice-sheet like that of Greenland, and it was this ice-sheet that spread the boulder-clay over the plains and scattered the erratic blocks. According to other authorities the region was submerged and only the mountain districts rose above the water. Amongst the mountains there were glaciers which reached the sea, and ice-bergs carried off the glacial material and, melting, dropped it on the submerged plains.

Whichever view is correct the boulder-clay was formed in some way by the agency of ice; and a region covered by boulder-clay has a special character of its own. Often the boulder-clay has blocked up pre-existing valleys and diverted the courses of the rivers. The Vale of Pickering in Yorkshire is a wide valley running from west to east and opening out upon the coast. It seems to offer a direct and natural route for a river flowing eastwards into the North Sea. But the Derwent rises at its mouth, within a short distance of the sea, flows westwards in the Vale and breaks through the hills that form its southern bank, and finally flows into the Ouse. The mouth of the Vale is blocked by boulder-clay. It was probably dammed up still more by ice, and a lake was formed. The lake overflowed on its southern side, and the outflowing stream cut the channel by which the Derwent now reaches the Ouse.

The surface of the boulder-clay is very uneven, with low rounded hills and shallow irregular depressions. The depressions are often occupied by lakes. In England the lakes have usually been silted up or else drained by the outflowing streams; but in eastern Prussia the original character has not yet been lost, and the boulder-clay forms a low plateau bearing scores of shallow and irregular lakes upon its surface.

The unevenness of the boulder-clay forces the streams upon it to take tortuous and unexpected courses. Sometimes, too, a river bifurcates and the two branches flow in very different directions. In time these irregularities tend to be removed by the erosion of the rivers themselves, and in England this has happened to a very considerable extent. In Sweden it is not so long since the ice retreated, and bifurcations and other irregularities are more frequent.

CHAPTER XXV

WIND

In the British Isles the wind is often violent, and, upon the coast, by producing waves and currents in the sea, it is indirectly the cause of considerable changes in the shore-line; but inland its effect upon the form of the surface is small compared with that of rain and rivers. Nevertheless even here it transports dust and the finer particles of soil from place to place. The remains of Roman and other ancient buildings are often buried beneath the soil, and the burial is due in part to the action of the wind.

In England, however, the surface is usually covered with grass or other vegetation and it is seldom perfectly dry. The moisture causes the loose particles to adhere to one another, the roots of plants bind the soil together and their stems and leaves protect it from the wind. Only on some of the barer heaths and on sandy shores above high water mark are the effects of the wind at all conspicuous.

In drier regions it is different. There is no dampness in the surface soil and but little vegetation to interfere with the action of wind ; and in an actual desert the wind becomes one of the most important of the agents of land sculpture.

Wind, like running water, transports, corrades and deposits ; but it is different from running water both in its mode of action and in its effects. It has not the power of a rapid river and cannot move such heavy fragments as a river often does ; but its strength is independent of the slope over which it blows, and it can carry material upwards as well as downwards. Here and there a turbulent stream may throw sand and mud upon its banks, above the general level of the water, but the upward movement is never more than a few feet. Wind on the other hand whirls the finer particles to great heights, and in desert regions it often blows clouds of dust across a mountain range.

The material carried. The wind can carry with it any loose material which lies upon the surface and which is sufficiently light and fine. In a temperate climate such material may be produced by rivers, by the waves of the sea or by the ordinary processes of weathering. In a dry climate rivers are generally absent, and the weathering is due mainly to changes of temperature and to dew. The surface of the rock is broken up by the changes of temperature and the wind itself may then assist by causing the smaller pieces to knock against one another. Dew causes decomposition in many kinds of rock and the products of the decomposition are usually soft and powdery.

The fineness of the loose material produced by the weathering will depend partly upon the nature of the rock ; but also to some extent upon the wind itself, for as the particles are blown along they hit against one another and against the ground and are therefore broken up still further.

Transport. The power of the wind as an agent of transport is shown quite clearly even upon our own shores. During a gale the wind will often blow the smaller pebbles along the beach, and in gusts it may even move fragments of considerable size. The sand is blown upwards into the air and may travel far inland.

The distance to which the wind can carry the smaller particles is very great. Dust from the Sahara is often blown into Southern Europe and occasionally it even reaches our own islands. The "red rain" which fell in February 1903 at several places in Great Britain seems to have owed its colour to particles of dust brought from Africa by the wind.

The dust from volcanic eruptions is carried still greater distances, partly because it is very fine, partly because it is thrown high up into the air, where the winds are stronger than on the ground. Volcanic ash from Iceland has fallen in the north of Scotland, and the finest of the dust from the Krakatoa eruption was carried several times round the globe before it settled.

Erosion. The material transported by the wind, like the sand and pebbles carried by a river, becomes an instrument of corrasion wearing away the surface with which it comes in contact. In desert regions accordingly the exposed rocks are fretted, the harder bands standing out and the softer bands being more deeply worn. The effect is very well shown in the statue of the Sphinx in Egypt.

Since the heavier material is swept along the ground and only the smaller fragments are carried up into the air, the corrasion is greatest near the ground and decreases rapidly upwards. A projecting crag will therefore be gradually cut away at the base. If the direction of the wind is constant, only the windward side will be worn away; but if the direction is variable, the crag will be undercut all round (Fig. 154).

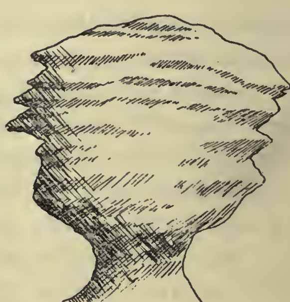


FIG. 154. Rock worn by the wind.

The erosion of a horizontal surface of rock is beautifully shown in the Sinai peninsula. A large area is covered by a layer of sandstone, called the Nubian Sandstone, in which there are sometimes many manganese concretions. The sandstone is gradually worn away by the wind, but the concretions are harder and more resistant. They accordingly stand out above the surface, and on the leeward side of each is a little mass of the sandstone protected by the concretion from the wind (Fig. 155).

Wherever the erosive action of the wind is great and deposition small, the solid rock will be laid bare and a rock desert

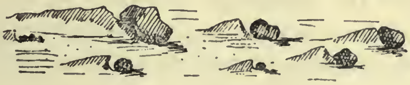


FIG. 155. Wind erosion in the Sinai peninsula. (After Walther.)

will be formed. If the beds are horizontal, the softer strata will be worn away until a hard layer is exposed.

In the hard rock erosion is assisted by dew, and takes place chiefly along the joints, where the dew is sheltered from the sun. The hard rock is cut through at the joints, down to the

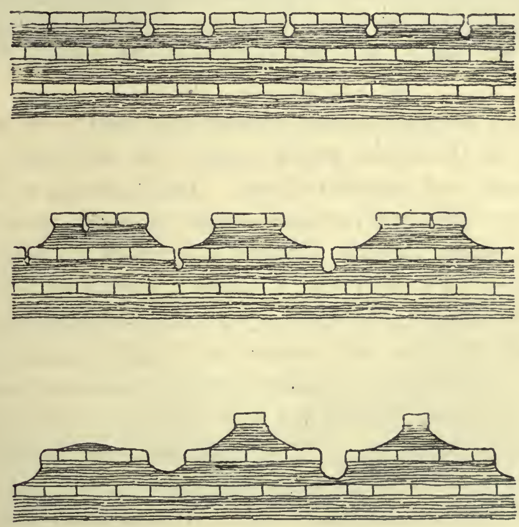


FIG. 156. Stages in the erosion of horizontal strata by the wind.

next soft bed of rock beneath. By the combined action of dew and wind the hard layer is gradually undermined and falls, until only a few caps are left resting upon pedestals of the softer rock. Finally even these are completely removed, and the next hard bed is exposed and is eroded in its turn (Fig. 156).

Sometimes the erosion proceeds in a different way. The surface of the rock is broken up by changes of temperature and the ground is covered by a layer of fragments, which serves as a protection to the rock beneath. Erosion is not concentrated at the joints, but the fragments continue to break up and the smaller particles are carried away by the wind. Deserts of stones formed in this way occupy large areas in the Sahara and Arabia.

Deposition. Sooner or later the material carried by the wind must be deposited. It may be dropped when the wind dies down, and fall either on the land or into the sea ; or it may be drifted into a sheltered hollow or banked against an obstacle.

To a certain extent the material is sifted according to its coarseness. The fine dust can be carried by lighter winds than the sand and travels farther. A mountain range is a barrier to the sand, but the dust may be blown completely over it.

Partly on account of this sifting and partly on account of the nature of the rocks which supply the material, there are deserts of sand and deserts of loam. Both are due to deposition and thus are different in their mode of origin from the rock deserts and stony deserts already described, in which erosion predominates. Pl. XVI, fig. 1 shows a valley in South Africa which has been partially filled up by wind-blown sand.

In sandy deserts the surface is usually undulating, consisting of a succession of dunes with depressions between. A sand-dune is formed wherever any obstacle, such as a cactus bush, interferes with the free movement of the wind (Pl. XVI, fig. 2). The sand is heaped against the obstacle until it overtops it and falls on the other side. The heap of sand is itself an obstacle and the same process goes on until the heap becomes a little hill. The windward side, up which the sand is blown, slopes gently, the leeward side, formed by the falling sand, is steep.

The wind blows the sand not only over the hill but also round it. Therefore the sides of an isolated dune are prolonged as horns in the direction towards which the wind is

A. W. Rogers, photo.

Fig. 1. Sand-filled valley.

Henkries Valley, Little Namaqualand.

W. R. Rickmers, photo.

Fig. 2. Sand hills.

Turkestan.

blowing, and in plan the dune is crescent shaped. A simple dune of this kind is called a barkhan (Fig. 157).

If the winds are fairly constant in direction the sand on the windward slope is continually blown up to the top and falls

FIG. 157. A barkhan.

down on the leeward side, and the sand-dune travels slowly forward. If the winds are variable the dune becomes a shapeless heap and its movements are irregular.

When the dunes are crowded, two or more may unite and may lose their crescent shape. In the Indian desert two distinct types of sand-hills are conspicuous. In one type (Fig. 158 *a*) there is a ridge of sand at right angles to the prevalent wind, with a steep slope on the leeward and a gentle slope on the windward side. But the windward slope is not

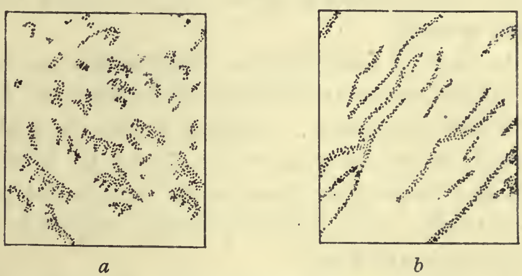


FIG. 158. Sand-hills in the Indian desert.

uniform ; it is divided by little valleys into a series of spurs at right angles to the main ridge. Probably these hills are formed by the coalescence of several barkhans, and possibly each spur may originally have been the windward slope of a separate dune[1].

[1] It is worthy of notice that many of the smaller hills are horse-shoe shaped and look like barkhans formed by the north-east monsoon, though it is the south-west monsoon which is the stronger here.

In the other type (Fig. 158 *b*) the shape is entirely different. Each hill is a long and narrow ridge running in the direction of the prevalent wind. Both sides are steep and the leeward end is also steep. The crest of the ridge rises gently from the windward to the leeward end. It is in fact like an ordinary dune enormously elongated in the direction of the wind. Here and there two neighbouring ridges are connected by a short transverse ridge.

The formation of these elongated sand-hills appears to be connected in some way with the strength of the wind; for they are confined to the western and seaward margin of the desert, where the force of the south-west monsoon is presumably at its maximum.

On the edge of a sandy desert the forward movement of the dunes has often devastated areas which once were fertile, and in Egypt and Syria ancient buildings and cities have been buried beneath the sand. Even in our own islands the dunes of a sandy shore sometimes encroach upon the cultivated fields; and on the southern shores of the Moray Firth the sand-hills have penetrated a forest, overwhelming the trees on their forward march.

In the Landes of Gascony the destruction caused by the advancing dunes became so serious that measures were taken to prevent their movement. Certain herbaceous plants with spreading roots were sown upon the dunes to bind the sand temporarily and to prepare the way for permanent plantations of coniferous trees. The method proved completely successful and the dunes are now fixed.

Even in actual deserts, however, the movement of the sand is often entirely superficial and the greater dunes do not alter their positions. In some cases it may be because the winds are variable; in others it is because the dunes serve as reservoirs of water. Rain is rare, but when it falls it soaks at once into the sand. Evaporation is confined to the surface layers and lower down the sand is more or less moist. Damp sand is not easily disturbed by the wind, and accordingly it is only the superficial sand that moves. The larger the dune the more completely is the water in its depth protected; and thus in

Plate XVII

W. R. Rickmers, photo.

Valley in loess.

Near Guzar, Turkestan.

some regions the smaller sand-dunes move while the larger ones are practically stable.

Loess. The deposits of loam formed by the wind are often known as loess (Pl. XVII). They spread far beyond the limits of the desert regions and cover extensive areas in Central Europe, China, and America, where the rainfall is fairly abundant. But the remains of animals found in the loess suggest that it was formed when the climate was drier than it is at present.

In China, where the deposit is of enormous extent and thickness, the loess is a fine calcareous loam, yellowish or buff in colour. It is penetrated by innumerable vertical and very narrow tubes, probably produced by the rootlets of the grass which grew upon the surface while the deposit was forming. The loess is still soft, and deep valleys have been cut in it by streams. Since it is very porous the rain sinks quickly into it, and the surface is dry. Partly for this reason and partly because the loess breaks most readily along the vertical tubes, the sides of the valleys are often almost vertical or rise in a series of precipitous steps.

Somewhat similar deposits occupy many of the depressions between the mountain ranges of Central Asia. In general where there are no rivers the surface is nearly level, but towards the mountains it rises gently, frequently extending high up the flanks of the ranges.

CHAPTER XXVI

INFLUENCE OF CLIMATE UPON TOPOGRAPHICAL FEATURES

Climatic zones. The shapes of hills and valleys are determined to a very large extent by the agents of denudation which are at work upon them, and the nature of these agents depends mainly on the climate. Where, for instance, rain is frequent, denudation is due chiefly to water; where rain is rare, wind and changes of temperature are of far greater importance.

Consequently the topographical features of a country are greatly influenced by its climate.

It will be useful therefore to consider briefly the general processes of earth sculpture from the geographical point of view; and for this purpose the globe may be divided into regions according to temperature and rainfall, as follows:

Equatorial zone and the Monsoon regions. Temperature high; rainfall heavy.

Tropical dry regions. Temperature high; rainfall small.

Temperate zones. Temperature moderate; rainfall moderate.

Polar regions. Temperature low; precipitation mainly in the form of snow.

But no sharp boundary can be drawn between these regions. Moreover, the temperate zones, in particular, are far from uniform in character. They include areas of very low rainfall, while in the mountainous districts the rainfall may be excessive; the average temperature is always moderate, but the range of temperature may be very small or very great.

Earth sculpture in temperate regions. In temperate regions wherever there is at least a moderate rainfall the greater part of the work of erosion is carried on by rivers. Weathering is due partly to the chemical action of water, with carbon dioxide and other substances in solution, partly to the mechanical action of frost. Where the former predominates the rocks decay and rounded surfaces are produced; where the latter is more important, angular fragments are broken off and rugged crags are common. In either case the loose material is carried downwards by running water and finds no permanent resting place until it reaches the sea. The valleys are river valleys and, even on a hill-slope where there is no actual stream, it is chiefly running water, during showers of rain, that washes down the soil and weathered rock. Thus not only the valleys but also the sides of the hills usually show the curve of water erosion; and this is the most characteristic feature of a temperate climate. The hills rise gently from the plains and the slope increases gradually towards their summits. In the lowlands, where chemical weathering prevails, the hill features are smoothed and rounded off; but in the mountains, where frost plays a more important part, jagged summits and rugged slopes are frequent.

Earth sculpture in tropical dry regions. In a tropical desert water is generally absent. There is no vegetation to hold the soil together and wind accordingly becomes the principal agent of transport and erosion. The breaking up of the rocks is due mainly to changes of temperature and there is but little decay. The broken fragments are therefore angular, and the hills, unless they are made of some soft rock like shale, are rugged. The loose material is carried downwards by gravity, by the wind and by occasional storms of rain ; but there are no rivers to bear it to the sea, and it collects in the depressions and gradually fills them up (Pl. XVI, fig. 1). The lower hills may be completely buried, and only the tops of the higher ranges may be left uncovered. The loose material is still further broken up by changes of temperature and distributed by the winds, and forms extensive plains, diversified by sand-dunes. The curve of water-erosion is seldom seen, and the characteristic features are wide plains of broken rock or sand or loam, with hill ranges rising abruptly like rocky islands in a sea.

Earth sculpture in the equatorial zone. In equatorial regions the rainfall is excessive and vegetation is luxuriant. The temperature is high and uniform. On account of the warmth and moisture, together with the acids produced by the decomposition of dead leaves and plants, decay of the rocks is rapid, and the depth of the soil, or covering of weathered rock, is usually great. The most prominent parts, because they are the most exposed, decay most quickly and all angularities are rounded off. Owing to the dense growth of vegetation the loose material is not easily removed ; and in spite of the presence of running water, the hills very commonly show the convex shape characteristic of chemical weathering.

Earth sculpture in polar regions. In polar regions there is no running water except in the height of summer. There is no decay of the rocks, but the rocks may be broken up by changes of temperature or at certain seasons by frost. Owing to the absence of vegetation the wind may become an important agent of denudation. Therefore in many respects the topographical features are often not unlike those of a tropical desert. The hills are equally abrupt, without any well-marked curve of water-erosion, and on the lower-lying ground there may be the same accumulation of loose material.

Much, however, depends upon the amount of snow. Where the snow-fall is heavy, as in Greenland and Spitzbergen, the whole region may be covered by an ice-sheet, or glaciers may descend from the mountains to the coast. Much of the loose material is carried away to sea, and the topographical features, concealed beneath the ice, must be those of a glaciated region. Any peaks, however, which may pierce the covering of snow and ice, show the jagged summits and angular outlines characteristic of the action of frost.

General observations on earth sculpture. Thus each of the four chief types of climate has its own characteristic topographical features; but the typical forms are liable to modification. A resistant rock will always tend to project, and may form steep slopes and rugged crags in the wettest of regions. A rock that allows the rain to sink rapidly into it may show little indication of the curve of water-erosion even in a temperate climate, and may present some of the features of a desert region. The flat-topped hills of Egypt are very different from the rugged ranges of Baluchistan though both show the abrupt slopes of a region destitute of water. In the former the strata are horizontal, in the latter they are strongly folded. Glaciers and glaciated valleys occur in the higher mountain ranges of low latitudes, and in the north of Europe and America even the lowlands still retain the features impressed upon them by their former Arctic climate.

It should perhaps be added that much of the Temperate Zone, as the term is usually understood, has a far from temperate climate. It includes, in Central Asia, some of the largest deserts in the world and in North-eastern Siberia the most extreme climate known. In the former the topographical features are naturally those of a desert; in the latter they are to a large extent those of a polar climate with a small snow-fall, but modified by the presence of rivers.

CHAPTER XXVII

VOLCANOES

Condition of the earth's interior. In mine-shafts and well-borings the temperature usually increases with the depth. The rate of increase is far from uniform but a large series of observations gave an average rise of about 1° F. for each 64 feet of descent. How far downwards the increase continues is still uncertain, but if it is maintained unchanged the temperature of the interior of the earth must be above the melting point of any of the rocks of the earth's surface.

These and other considerations led the earlier geologists to conclude that the interior must be liquid and that the solid crust is a comparatively thin layer floating upon the molten mass. But in the interior of the earth the pressure must be great, owing to the weight of rock above, whether this rock is solid or liquid; and increase of pressure raises the melting point of most kinds of material. It was therefore suggested that in spite of its high temperature the interior is probably kept solid by the pressure.

The strongest argument in favour of the solidity of the earth is derived from the tides. If the interior is liquid it will be affected by the moon in the same way as the waters of the ocean, and the outer crust will yield to the movements of the molten material beneath. If it yielded as much as the water does, the tides would be imperceptible; and if to a less extent, the apparent height of the tides would be reduced by the amount of the movement of the crust. From the heights of the tides as actually observed, it has been calculated that the amount of yielding is very small, and that the earth as a whole is as rigid as a solid ball of steel of the same size. The natural inference seems to be that by far the greater part of the interior is solid.

It is, however, certain that molten rock or lava rises at times from below and is poured out over the ground; and beneath the surface there must be reservoirs of molten rock or else of rock which melts when the pressure upon it is removed.

Formation of volcanoes. Volcanoes are formed by the escape or attempted escape of this material. It is possible that water percolating through the crust comes into contact with the molten or potentially molten rock, and the sudden conversion of the water into steam forces the lava upwards and causes the eruption. In support of this view it has been pointed out that most volcanoes are situated near the sea, and that volcanic eruptions are generally accompanied by the emission of enormous clouds of steam.

Another suggestion is that pressure produced by earth movements is the primary cause of the eruption of the molten material, and certainly volcanoes are most abundant in regions where earth movements are proceeding. But the whole question is one of great complexity and cannot be discussed more fully here.

A typical volcano is a conical hill, from the top or sides of which eruptions take place at intervals. The hill is the product of the eruptions. It is formed of the material thrown out, which is naturally deposited most thickly near the outlet and less thickly at a distance.

In most volcanoes there is a funnel-shaped hollow at the top of the cone, and this hollow is called the crater. The bottom of the funnel opens into the channel or pipe through which the erupted material finds its way to the surface. When the volcano is not in action, the pipe is usually plugged by solidified lava or by fragments which have been thrown up into the air and have fallen back into the crater.

Sometimes an eruption takes place on the sides of the volcano and forms a secondary or parasitic cone with a crater of its own.

Products of eruption. The material thrown out during an eruption may be solid, liquid, or gaseous.

It has generally been believed that the great masses of cloud which rise from an erupting volcano, consist mainly of condensing steam; but recent observations have shown that it is not always so, and according to M. Brun the clouds emitted during eruptions in Java and elsewhere contain no more water vapour than the surrounding air. He believes that they consist chiefly of ammonium chloride.

There can be little doubt, however, that water vapour is sometimes present in enormous quantities. Volcanic eruptions have frequently been followed by deluges of rain, and the rain appears to be formed by the condensation of the vapour emitted from the volcano; but a thorough examination of the clouds sent out during an eruption is impossible.

Amongst other gases which have been detected in considerable quantities hydrochloric acid, sulphuretted hydrogen, sulphur dioxide, hydrogen and carbon dioxide are perhaps the most important. Inflammable gases such as hydrogen and some of its compounds are occasionally in sufficient quantity to produce actual flames; but in most eruptions the appearance of flaming at the top of the cone is due chiefly to the red-hot fragments thrown up into the air and the reflection of the glowing lava upon the rising column of cloud.

The gases issuing from a volcano are for the most part dissipated into the air and have little effect upon the volcano itself. They may, however, act chemically upon the rocks in the neighbourhood of the channel through which they escape, or they may produce by their reactions with each other deposits of sulphur and other minerals in and near the crater.

The liquid products of the eruption are of much greater importance. They consist of molten rock and form the streams of lava (Pl. XVIII, fig. 1) which flow out of the crater or out of fissures in the side of the volcano. Some lavas contain a high proportion of silica and are said to be "acid"; in others the percentage of silica is comparatively low and these are known as "basic." Acid lavas have a high melting point and are usually very viscous, and therefore they flow slowly and do not travel far. Basic lavas melt at a lower temperature and are generally very liquid; and a basic lava stream moves rapidly and may flow for many miles before it becomes solidified.

Sometimes the surface of a solidified lava is smooth; but more often it is very rough (Pl. XVIII, fig. 2). If the lava is viscous, the surface becomes ropy like that of a stream of flowing pitch. Frequently, too, both in acid and basic lavas, escaping gases make the upper layers of the stream vesicular, or full of little holes, like a piece of bread. Moreover, the outer surface cools and solidifies first, forming a crust, which is continually being broken up and carried forward by the moving stream beneath. The fragments of the crust, if they are vesicular, are known as scoriae.

Consequently a lava-flow of recent date has usually a very rough and irregular surface, and its upper layers are full of holes and crevices into which the rain can penetrate. Partly for this reason weathering is rapid, and often produces a covering of rich and fertile soil.

In some volcanoes the liquid lava wells up quietly, filling the crater and overflowing without any great disturbance. But most eruptions are accompanied by explosions, which are often very violent. In such cases vast quantities of broken rock are thrown into the air and fall back into the vent or cover the country round. At the birth of a volcano the fragments are pieces of the rock through which the pipe of the volcano is drilled; but at a later stage they are for the most part fragments of lava. The solidified lava that commonly plugs the channel of a quiescent volcano must be blown out before the molten rock beneath can rise into the crater. Sometimes the whole top of the hill is blown off by the explosion and sometimes an entirely new channel is opened on the sides. No doubt, too, in many cases there has been partial solidification in the molten reservoirs beneath, and the solidified rock is thrown out during the explosion.

In most eruptions, therefore, a large amount of fragmental solid material is ejected, and showers of dust and stones fall over the surrounding area. The larger pieces form "breccia," a name applied to deposits of large and angular fragments; the smaller pieces are known as cinders or volcanic ash or, if the material is very fine, volcanic dust.

The liquid lava may also be blown into the air and may solidify in drops before it reaches the ground. Volcanic bombs are masses of lava which have been thrown out in this way in

Tempest Anderson, photo.

Fig. 1. Lava flow.
Kilauea.

Tempest Anderson, photo.

Fig. 2. Scoriaceous lava.
Etna.

a more or less liquid condition, and have assumed a rounded or pear-shaped form as they fell.

Owing to the torrential rains which sometimes accompany an eruption the volcanic dust may be washed down as a stream of mud, causing as much destruction as a flow of lava. In the great eruption of Vesuvius in 79 A.D., Herculaneum was overwhelmed by a stream of volcanic mud which has since become compacted into solid rock.

Forms of volcanoes. The form of a volcano depends largely upon the material of which it is composed.

Crater rings. In some cases there is nothing but a cavity produced by the explosion and around it a ring-shaped mound consisting of the fragments of rock thrown out. Crater rings of this type may be seen in the Eifel district with a small lake within the hollow.

Cinder and ash cones (Fig. 159). More often the fragments of rock and lava are piled up till they form a hill of considerable size. Naturally the accumulation is greatest immediately round the vent and gradually decreases outwards, and a volcano formed in this way is conical in shape. Successive eruptions deposit layer after layer of ash on the sides of the cone, and thus on the flanks of the volcano the beds dip outwards. But in the crater itself the ash falls inwards, forming beds which dip towards the centre.

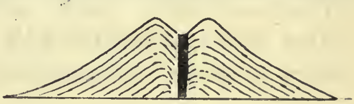


FIG. 159. Ash cone.

Volcanoes of this type, consisting entirely, or almost entirely, of fragmental material, are known as ash or cinder cones. In shape they are often almost perfect cones, but the cone spreads out towards the base so that in profile the sides are curved instead of straight (Pl. XIX, fig. 1). The curve is concave upwards and is remarkably like the form produced by water erosion. Near the vent the fragments are piled so thickly that the cone slopes at the angle of rest, *i.e.* at the steepest angle at which such loose material can stand; and fragments falling at the top or on the sides roll down the slope. Farther away from the vent the accumulation decreases and the hill is formed, not of fragments which have rolled down the sides, but of the material which has been thrown outwards from the volcano and has fallen from the air.

If a strong wind blows during an eruption, the cinders or ashes will fall more thickly on the leeward side of the vent and the cone will be unevenly developed. Unsymmetrical volcanoes may be formed in this way in regions where the winds are constant; or even where the winds are variable, if the cone is due to a single eruption.

Lava volcanoes (Figs. 160, 161). Volcanic eruptions, however, are not always explosive. In some volcanoes the lava rises slowly and is poured out quietly. Puffs of gas and vapour may escape from the molten rock, but they cause very little disturbance.

When this is the habitual type of eruption there are no fragmental deposits and the volcano is formed entirely of streams of lava. Such volcanoes are usually dome-shaped rather than conical (Pl. XIX, fig. 2), but the form of the dome varies according to the nature of the lava.

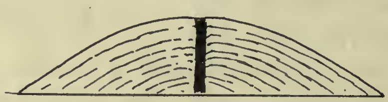


FIG. 160. Acid lava volcano.

If the lava is acid and viscous it behaves like a mass of paste squeezed through an opening, and the dome is very convex, with steep sides and often without a crater (Fig. 160).

If the lava is basic and fluid, it flows readily and for great distances. The diameter of the volcano is therefore great but

FIG. 161. Basic lava volcano.

the slope of its sides is gentle, and the form is that of a very flat dome. The crater is simply the hollow left when the lava in the pipe subsides (Fig. 161).

Mauna Loa and Kilauea in the Sandwich Islands are excellent examples of lava volcanoes. Mauna Loa is 13,675 feet

Fig. 1. An ash volcano.
Misti, near Arequipa, Peru.

Tempest Anderson, photo.

Fig. 2. An acid lava volcano.
Grand Sarcoui, Auvergne.

high, but the angle of slope is only about 6°. Kilauea is about 4040 feet and lies upon the flanks of Mauna Loa. It is not, however, a subsidiary cone receiving its lava from the same source, for eruptions in Mauna Loa cause no disturbance of the liquid lava which is always visible in the crater of Kilauea, nearly 10,000 feet below. Kilauea must be an independent volcano which has been partly overwhelmed by the flows from Mauna Loa.

In both volcanoes the crater is a broad and relatively shallow pit with vertical sides and a terraced floor. On the lowest platform in the crater of Kilauea there are lakes of very fluid lava, constantly boiling. Now and then a crust forms on the surface for a time, but soon breaks up and is engulfed. The floor of the crater is itself a thicker crust resting upon the lava in the pipe, and the lakes are really holes in this crust through which the lava appears. The level of the floor is not always the same, but rises and falls with the lava in the pipe.

Outflows of lava may take place either from the crater itself or from the sides of the volcano. But even in the latter case there is as a rule no violent explosion. The lava appears to melt its way quietly through the rock, until it reaches the open air. In this respect lava volcanoes of this type are very different from such volcanoes as Vesuvius and Etna.

Composite volcanoes. The great volcanoes generally consist partly of ash and partly of lava, in irregularly alternating beds. In the early stages of their growth, the form is conical, as in a simple ash volcano ; but the shape is usually not so perfect, because the streams of lava flow out unevenly. As the height of the volcano increases, the column of lava in the central channel is compelled to rise higher and higher to reach the crater. The pressure that it exerts upon the walls of the channel increases until at length it is easier for the lava to force its way through the sides of the hill than to rise up to the crater. In large volcanoes, therefore, the streams of lava usually issue from fissures in the flanks, and subsidiary cones are often formed.

Volcanic eruptions. If a volcano has remained quiescent for a long period, the channels through which the lava escaped become choked either by solid lava or by fallen fragments. When the next eruption occurs, the old channels must be cleared or new ones must be opened. Usually the pressure inside increases until at length there is a violent explosion. The top of the hill may be blown off, or the explosion may occur upon the flanks. In either case a gigantic cavity is formed, and in the middle of this a new cone may be piled up (Pl. XX, fig. 3). The large cavity is sometimes known as a caldera.

The history of Vesuvius illustrates these changes. In the time of the Romans the mountain had the form of a truncated cone with the top hollowed out into a great amphitheatre (Fig. 162 *a*). There were then no records of any eruptions; but presumably the top of the cone had been blown off in some

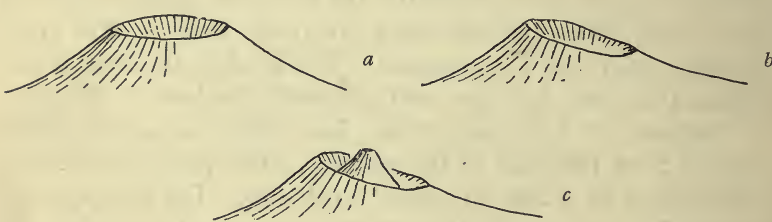


FIG. 162. Stages in the history of Vesuvius.
a. Probable shape before 79 A.D. *b.* Shape after the eruption of 79 A.D. *c.* Commencement of the new cone.

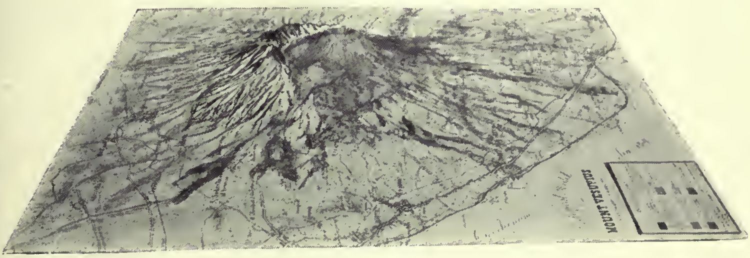
prehistoric explosion. In 79 A.D. the volcano again became active and a great explosion destroyed half the wall of the amphitheatre (Fig. 162 *b*), and buried Pompeii and Herculaneum in volcanic ash and mud. The other half of the wall still stands as a semicircular ridge, while in the midst of the broken hollow a new cone has since been built up with a crater at its summit (Fig. 162 *c*). The semicircular ridge round one side of the new cone is called Monte Somma (Pl. XX, figs. 1, 2).

In Etna a similar explosion must have occurred at some very early date, probably before the appearance of man upon the globe. It was not the top of the hill, however, that was blown off, but a portion of the side, and an immense hollow, now known as the Val del Bove, was formed on the flanks of the volcano.

Plate XX

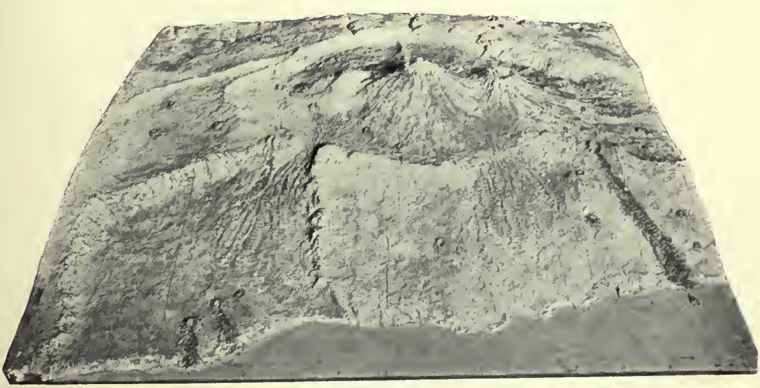
W. Tams, photo.

Fig. 1. Model of Vesuvius, from the south.



W. Tams, photo.

Fig. 2. Model of Vesuvius, from the west.



W. Tams, photo.

Fig. 3. Model of the Peak of Teneriffe, from the north

One of the greatest eruptions of modern times was that of Krakatoa in 1883. Krakatoa is the principal island of a little group in the straits of Sunda, between Java and Sumatra. The whole group is the remains of a great caldera formed by some ancient explosion and partially submerged. Upon the rim and in the midst of the caldera subsidiary cones had arisen, but the whole group had been quiescent for about 200 years. Premonitory earthquakes occurred in 1878 and succeeding years, and in May, 1883, eruptions began and gradually increased in intensity. On August 26th a succession of most violent explosions took place and continued till the morning of the 27th. The ashes and cinders thrown up darkened the sky for miles around and, falling in the sea, obstructed the navigation of the straits for days. The noise of the explosions was heard almost all round for a distance of about 2000 miles and it was even audible at Rodriguez, nearly 3000 miles away. About two-thirds of the island disappeared.

The group of islands was not permanently inhabited and the actual eruption caused little loss of life. But the disturbance produced great waves in the sea which swept along the coasts of Java and Sumatra and drowned many thousands of the inhabitants.

The fine ash was shot many miles upwards and reaching the region of the upper winds was carried several times round the globe. It extended even as far as our own islands and was the cause of the extraordinarily brilliant sunsets observed in the autumn of 1883.

Another disastrous eruption of modern times was that of Mont Pelée, in the island of Martinique, in the year 1902. About the same time a similar eruption took place in the neighbouring island of St Vincent. The prominent feature in both eruptions was the sudden outburst of a dark and heavy cloud, consisting of hot gases and incandescent dust, which rolled down the mountain side with great rapidity. Everything that lay in its path was burned and destroyed and the town of St Pierre at the foot of Mont Pelée was overwhelmed in a few minutes. Less violent eruptions broke out at intervals for several months and when they had begun to subside a remarkable spine-like projection rose slowly from the crater of Mont Pelée to a height of 700 feet or more above the cone. It was formed by the partially solidified and pasty lava in the neck which was gradually forced outwards by the pressure within. The spine, however, was unable to resist the action of the weather and rapidly crumbled away.

Fissure eruptions. Volcanic eruptions sometimes take place not from a single vent but along a line of considerable length. In such cases the channel communicating with the interior is evidently not a pipe but a fissure, and the eruption may occur simultaneously throughout the whole length of the fissure or at numerous points along it. The eruption of Laki in Iceland in the year 1783 was of this type, the fissure being about 20 miles in length. The eruption of Tarawera, in New Zealand, in 1886, appears to have taken place from a fissure about nine miles long.

The Tarawera eruption was violently explosive and large quantities of ash were ejected, but fissure eruptions generally seem to be characterised by the absence of explosive action. The principal feature is the quiet welling out of molten lava which may spread over many miles of country.

Fissure eruptions are rare at the present day, but in the past they have taken place on a gigantic scale. The basaltic lavas of the Deccan in India, covering an area of about 200,000 square miles, and the lava-flows of the Snake River plains in the United States, about 200,000 to 250,000 square miles in extent, appear to have been poured out from fissures. In our own islands, the basalt flows of the north-east of Ireland and the Hebrides are of a similar type, and are merely the remains of an enormous lava-field which probably extended as far as Greenland.

Solfataras. Long after a volcano has ceased to eject lava and ashes and has become practically extinct it may continue to emit steam and gases of various kinds. It is then said to have reached the solfatara stage, the term being derived from the volcano called Solfatara, near Naples, the last recorded eruption of which occurred in 1198 A.D.

Volcanoes in the solfatara stage are common in all the great volcanic districts of the globe.

Geysers. In some parts of the world where volcanic action is still going on or has taken place in recent geological times, hot water and steam are thrown out at intervals in the form of a fountain, sometimes rising to a height of one or two hundred feet. These intermittent fountains are called geysers and have long been known in Iceland. There are excellent examples also in the Yellowstone Park in the United States, and in New Zealand.

In the case of the Great Geyser in Iceland the water rises through a cylindrical pipe-like channel which opens in the middle of a basin-shaped hollow. When the Geyser is not erupting, the basin is filled with water at a temperature of 170° to 190°, while at a depth of a hundred feet in the pipe the temperature is about 260°. At intervals the water in the basin and the upper part of the pipe is thrown upwards into the air along with clouds of steam.

The cause of the eruptions is that the column of water in the pipe is heated down below, and the channel is so long and narrow that convection does not take place freely. Consequently the temperature below continues to rise while the water at the surface is still comparatively cool. Since pressure raises the boiling point, the water towards the bottom of the column must be heated far above 212° before it begins to be converted into vapour. But when the necessary temperature is reached, the change takes place and the water above is forced upwards and begins to flow away. The escape, even of a little water only, at once reduces the pressure at the bottom of the column and accordingly the water there is rapidly converted into steam, forcing the whole column above it into the air.

The phenomenon is sometimes illustrated accidentally in the chemical laboratory when a narrow test-tube full of water is heated at or near the bottom. If the tube is clean and the water has already been boiled so as to free it from dissolved gases, it not uncommonly happens that the water is ejected violently, owing to the sudden formation of steam at the bottom of the tube in the manner described.

Hot springs. In other cases the heated water flows out continuously, without any explosive action whatever. Hot springs of this type are common in volcanic districts, but they are found also in many places where there are no signs of volcanic activity. The springs of Bath, for example, have a temperature of 120°.

Mud volcanoes. If the erupted waters are muddy instead of clear, a conical mound of mud may be formed, with a crater at the top. Mud volcanoes are found in Sicily, New Zealand, and other volcanic regions, and here they probably represent the last phase of volcanic activity. But they occur also in the Crimea, at Baku on the Caspian, in southern Baluchistan and other districts, where no true volcanoes exist. In these cases the water is forced outwards, not by any kind of volcanic action, but by the production of gases beneath the surface in other ways. At Baku the volatile hydrocarbons given off from the petroleum-bearing beds beneath are probably the primary cause of the mud-eruptions. In other places gases are produced by the decomposition of organic matter or by other chemical changes.

Distribution of volcanoes. Volcanoes are not scattered irregularly over the globe. Most of those that are now active lie within certain well-defined belts, and by far the greater part of the earth has been free from volcanic action since man appeared upon its surface.

These belts coincide to a large extent with the belts of crumpling which have formed the great mountain ranges of the present day, but the coincidence is not complete. There are, for instance, no volcanoes in the Himalayas, and on the other hand there is no sign of recent folding in Iceland.

At the present day it is upon the borders of the Pacific that volcanic activity reaches its maximum development. A line of great volcanoes may be traced up the Andes and through Central America and Mexico. In the United States and Canada there are now no active vents, but in the ranges of the west there are many which have not been long extinct. Living volcanoes re-appear in Alaska and the line is continued through the Aleutian Islands, Kamchatka, the Kurile Islands, Japan, Formosa and the Philippines to the Moluccas group.

Another belt of volcanic activity, meeting the line already described about the Moluccas, runs through Sumatra, Java and the Sunda Islands generally. Barren Island, in the Andamans, which is still occasionally active, and some extinct volcanoes in Burma, mark the north-westerly termination of the belt, while towards the east it is continued, with several interruptions, through New Guinea, the Solomon Islands, the New Hebrides and New Zealand to Mount Erebus on the Antarctic continent. It should, however, be observed that owing to the large gaps, it is by no means clear that there is any real connection between the different volcanic regions in the portion of this belt east of the Moluccas.

The islands in the midst of the Pacific are all either volcanic or made of coral, and in many of the groups eruptions still take place. There are active volcanoes, for instance, in the Sandwich Islands, the Tonga Islands, and the Samoa group. The Fiji Islands are an example of a group which is of volcanic origin, but in which volcanic action has now ceased.

The great belt of folding which runs from west to east across Europe and Asia, like that surrounding the Pacific, is also associated with volcanic activity, but not to the same extent. In Italy and the neighbouring islands, Vesuvius, Etna, Stromboli and Vulcano are still active, and several other vents have erupted in recent times. Santorin in the Grecian Archipelago has been the scene of many outbursts. Farther east there are numerous volcanoes of gigantic size, but they are either extinct or in the solfatara stage. Ararat, for example, is volcanic and many other mountains in Armenia and Asia Minor. One of them, Nimrud, near Lake Van, is said to have erupted in 1441. Elbruz and Kazbek in the Caucasus and Demavend south of the Caspian are also old volcanoes, and the last still sends out sulphurous gases. In the region where the boundaries of Persia, Afghanistan, and Baluchistan meet, there are several volcanic cones of considerable size and one or two of them emit steam and other vapours. But these appear to mark the present easternmost limit of volcanic action in this belt of folding. In the time of Humboldt active volcanoes were said to exist in the great mountain chains of Central Asia, but all recent exploration tends to show that the reports were erroneous. There seem, however, to be a few extinct vents.

A shorter line of volcanoes, also associated with recent folding of the earth's crust, occurs in the West Indies, where the Lesser Antilles are largely volcanic. Most of the vents are extinct, but several of them still show signs of activity. Their nearness to the Andes and to Central America suggests that they may form a branch of the Pacific belt, but it is not clear that there is any real connection.

The remaining volcanoes of the globe appear to have no relation to the belts of folding, and in general show no definite linear arrangement. Iceland is the last surviving remnant of a great volcanic area which in earlier times extended from Greenland to the North of Ireland. The Azores, Madeira, Cape Verde Islands and Canary Islands are all volcanic, but the volcanoes for the most part are now extinct. In the Azores, however, there have been several eruptions in historic times, and Teneriffe in the Canary Islands has erupted even during the last few years. Ascension Island, St Helena and Tristan d'Acunha are all volcanic but have long been extinct.

In Africa there are a few volcanic centres. In the Cameroons there was an eruption in 1909. A considerable number of volcanoes lie in or near the great rift valley which extends from the Jordan down the Red Sea and through the east of Africa. Kenia and Kilimanjaro are volcanoes, though probably they are now extinct. A small cone south of Lake Rudolf has recently been in eruption and also one or two south of Albert Edward Nyanza. There are also records of an eruption in Arabia near Medina in the year 1256 A.D.

There are many volcanic cones in Madagascar, but no record or definite tradition of eruption. In the Comoro Islands, however, the Grand Comoro has been active several times since the islands were discovered. In Réunion the Ptone de la Foumaise is still frequently in eruption. Mauritius and many other islands in the Indian Ocean are volcanic but now extinct. Far to the south in Kerguelen Island there are still signs of activity.

Extinct volcanoes. It would take up too much space to enumerate the volcanic districts of the past; but leaving out of consideration the more remote geological periods, some of the European areas may be noticed here.

In the north-east of Ireland, in Skye, Mull and other islands of the Hebrides, there are enormous flows of basaltic lava. They belong to the volcanic area which, as already mentioned, once stretched as far as Greenland. The flows, apparently, came from fissures and not from distinct volcanoes.

In the Eifel district, west of the Rhine, and in the Auvergne, a part of the Central Plateau of France, there are many volcanic cones which are almost as perfect as when they were in eruption. Others have suffered from subsequent denudation.

Several extinct volcanoes, which are geologically of modern date, lie on the eastern side of the Rhine; and there are others also in Bohemia and on the inner border of the Carpathians.

CHAPTER XXVIII

LAKES

General conditions necessary. During a heavy shower the little hollows in the surface of the ground are usually filled with water and form temporary pools. When the rain ceases, the pools disappear; some of the water evaporates, some of it sinks into the earth. But as long as more water enters the hollow than can escape in this way, the hollow will remain full.

In a climate such as ours the annual rainfall exceeds the annual evaporation. Therefore, in general, every hollow, unless it is in porous rock, or unless there is some underground escape, is filled with water till it overflows. If the hollow is shallow, it cannot hold much water in proportion to its area, and it may dry up completely in the summer. If it is deep, the water will not all evaporate even in the driest seasons, and a permanent lake will be formed.

It is not necessary, however, for the formation of a permanent lake, that the rainfall over the lake should be greater than the evaporation from it; for the water is not all derived from the rain that falls directly into the lake. Much of the rain from the surrounding area drains into the hollow, and provided that the total amount received is equal to the loss, the water will never disappear. Consequently permanent lakes may and do exist even where the evaporation is greater than the rainfall.

In the tropical dry belt the rainfall is small and evaporation rapid. Even the larger depressions in the ground are therefore usually dry. If, however, they receive the drainage of a wide area, and especially if there are mountains round about, they may be covered with water at certain seasons. But even in these cases the annual supply does not usually exceed the loss, by evaporation and otherwise. Most commonly a temporary lake is formed when the snow upon the mountains melts, but in the dry season the lake becomes a swamp or disappears entirely. Even when a permanent lake is formed, the hollows are seldom completely filled and seldom overflow, and the lake accordingly is without an outlet.

Before a lake can be formed there must evidently be an actual hollow, completely surrounded by higher ground. If the supply of water is sufficient the hollow will be filled to the level of the lowest part of its rim and there the surplus water will overflow.

A hollow may be formed, whether the original surface was even or uneven, by deposition, by erosion (using the term in its widest sense), or by earth movements ; and lakes accordingly may be produced in any of these three ways.

Lakes due to Deposition.

Deposition does not always take place ~~uniformly,~~ and the surface of a newly formed deposit may therefore be uneven. If it is exposed to the air the hollows may be filled with water, and lakes will be formed which are completely surrounded by the deposited material. In cases such as these the lakes are due entirely to deposition.

More often, however, the origin of the hollow is more complex. Artificial reservoirs are usually constructed by building a dam across a river-valley ; and a great many natural lakes have been formed in the same way. Glaciers or other agents have deposited material in a valley, in such a fashion as to make a natural dam. The hollow is then due partly to the erosion of the valley by the river, partly to the formation of a dam by deposition. Since it is the dam that completes the hollow, such lakes may be classed as due to deposition.

In a few cases the material deposited is solid rock. It may, for instance, be a stream of lava now solidified, or a deposit of sinter or travertine from a spring. But usually the material is derived from the denudation of pre-existing rocks, and the deposit is an accumulation of fragments of various sizes. It may consist of angular blocks, boulders, pebbles, sand or clay. If it is made up entirely of large fragments the water can escape between them and no lake will be formed. But even a small proportion of clay may be sufficient to block the interstices and to make the whole accumulation impermeable. Moreover, percolating water may deposit silt between the fragments and may occasionally convert a leaky barrier into one that is water-tight. When the fragments are small, capillary attraction tends to hold the water between the grains, and thus even a bank of sand offers considerable resistance to the flow of water through it.

Marine deposits. The sea often throws up a bank of shingle about high water mark, and sometimes there is a small lagoon or salt water lake between the bank and the sea-cliff. Owing to the permeability of the shingle the lagoon is usually dry at low tide, but occasionally silt or decaying sea-weed makes its bed sufficiently impermeable to hold water permanently.

Lagoons on a larger scale are produced in the manner described in Chap. XVIII, by the formation of spits of sand or shingle across bays or across the mouths of rivers. The Norfolk Broads, as already explained, represent an advanced stage of the process.

Sand-hills also may cut off arms of the sea and convert them into lagoons, or may block the mouths of rivers. In the Landes of Gascony, in Holland, and in other low-lying districts many lakes and marshes behind the line of sand-dunes on the coast have been formed in this way.

Alluvial deposits. The lagoons of deltas and the ox-bow lakes of meandering rivers have already been described; but occasionally a stream may form a lake in another way.

In a hilly district the torrents running down the valley sides often carry with them, especially in flood time, enormous quantities of soil and broken rock. As the torrent flows out upon the floor of the valley, its velocity is checked and it deposits its burden as a delta. Very often the delta turns the main river in the valley out of its natural course. Sometimes the amount of material is so great that the river is unable to carry it away, and in time the delta is built completely across the valley. It then becomes a dam. The water above is ponded back and a natural reservoir or lake is formed. Sty Head Tarn, for example, owes its existence to a delta from the Gables.

Screes. In precisely the same way a valley may be more or less completely blocked by the screes descending from the crags above. Hard Tarn, on the flanks of Helvellyn, and Ffynnon Frech, on Snowdon, are dammed by screes.

Landslips. Much larger lakes than these have often been formed by landslips blocking up a river valley, but not in our own country. Such lakes are usually only temporary, for the dam consists of loose fragments piled haphazard, and has no great strength. As the water rises behind it the pressure upon it increases. When the water overflows, the dam is rapidly weakened by erosion and suddenly bursts, letting out a flood of water into the valley below. The unexpected floods which not uncommonly occur in the upper part of the Indus are believed to be due to this cause.

Glacial accumulations. More lakes probably are due to the action of glaciers than to any other cause. In our own islands it is where the signs of former glaciation are most conspicuous, that lakes are most abundant; but not even Cumberland or the Highlands can compare with north-western Russia for the number and size of its lakes. In southern Finland they cover nearly half the country; in the Government of Novgorod, it is estimated that there are 3,200 lakes. It is in this part of Russia that moraines and other indications of a former ice-sheet are still most evident.

The lakes in a glaciated region are due in some cases to deposition and in others to erosion. It is with the former that we are now concerned.

The original surface of a moraine or of a spread of boulder-clay is usually very uneven, and when it is first exposed, water collects in the hollows and forms lakes. As a rule such lakes are very irregular in shape and of no great depth. In our own islands most of them have long ago been silted up; but on the plateau of boulder-clay in north-eastern Germany many hundreds still remain.

These lakes are completely surrounded by glacial deposits, but more often the glacial material forms a lake by blocking up a river-valley. The terminal moraine of a glacier is a natural dam and, when the glacier retreats, may hold the water above it as in an artificial reservoir. For several reasons, however, a terminal moraine is seldom very effective as a dam. The stream which issues from the end of the glacier tends to keep a passage open while the moraine is forming, and the dam accordingly is often incomplete. Moreover, the terminal moraine consists chiefly of angular blocks loosely heaped together. It is not often watertight and it is not solidly enough constructed to resist much pressure or erosion. Consequently few lakes of any great size or depth are dammed by terminal moraines; but a considerable number of small lakes have been formed in this way. Bleawater Tarn in the Lake District, for example, is dammed by a semi-circular moraine at the mouth of the cwm in which it lies.

A far more effective barrier is often formed by the boulder-clay deposited beneath a glacier or ice-sheet. It is not laid down evenly throughout the whole length of the valley but the deposit is thicker in some places than in others. When the ice disappears the valley is partially blocked and water collects above the places where the accumulation is thickest. Windermere and Ullswater, the lake of Llanberis and many others are probably due, in part at least, to barriers of boulder-clay blocking up the valleys in which they lie.

As a rule the stream that leaves the lake, still runs down the original valley; but sometimes the valley is so completely blocked that the water overflows at the sides or even at the head of the valley, and finds some other course to the sea. In this case the actual outlet may be over solid rock, and at first sight it may appear that the lake lies in a hollow eroded in the rock.

The glacier itself may form a barrier which holds the water up. Where two glaciers meet there is sometimes a small triangular lake in the angle between them. In Greenland the inequalities on the surface of the ice due to projecting nunataks often lead to the formation of pools of water which may be of considerable size. Much larger lakes are sometimes found, both in Greenland and elsewhere, where a glacier crosses the opening of a valley that is free from ice. Unless the temperature is above the freezing point for a considerable part of the year no lake will be formed, but many glaciers extend far below the snow-line. The valley of the glacier itself may be filled with ice while the side valleys are occupied by running streams. As a rule these streams flow out upon the glacier and fall down crevasses, but sometimes the drainage of the lateral valley is completely blocked and a lake is formed. The Märjelen See in the Alps is the best-known example. It lies in a lateral valley opening into the valley of the Aletsch glacier. Sometimes the water can escape through crevices in the glacier; but often the dam of ice has been so high and perfect that the lake has overflowed over a col into the neighbouring valley of the Viesch glacier. Since there is always a free outlet over the col, the water can never rise to any greater height, and a beach or terrace is formed at the level of the col.

The parallel Roads of Glen Roy, north-east of Ben Nevis, were probably formed in this way. They are terraces or beaches evidently due to the action of waves and marking the height to which the water reached. Each terrace is on the level of a col over which the water could escape. The mouth of the valley was dammed by ice and a lake was formed. When the ice was at its greatest extension all the cols except the highest were also blocked, and the water stood at the level of the uppermost terrace. As the ice retreated the second col was opened and the water fell to its level and produced the second terrace. Next the third and lowest col was freed from ice and the lowest terrace was formed. Finally the dam at the mouth disappeared and the lake ceased to exist.

Volcanic deposits. In a volcanic district a stream of lava may flow across a valley and dam up a river. Lac d'Aydat near Clermont-Ferrand was formed in this way.

Crater-lakes are much more common. The crater of a volcano is a natural hollow in which water might be expected to accumulate. But in most cases the walls consist to so large an extent of loose volcanic ash that they are not water-tight. Moreover, since the crater is usually at the top of a hill, the only water that reaches it, after the volcano becomes extinct, is the rain that falls directly into it. Consequently the majority of craters are dry. There are, nevertheless, a considerable number of crater-lakes in the Eifel, the Auvergne and elsewhere; but most of them are in craters formed by explosion rather than by accumulations of volcanic ash.

Organic deposits. Even animal or vegetable deposits may occasionally give rise to lakes. The lagoon of an atoll is a kind of lake and owes its origin to the reefs built by coral polyps. In forest-covered districts rivers are sometimes blocked by floating trees washed down during floods. If one of these becomes jammed so that it cannot move, it may cause an accumulation of logs which for a time will form a dam. Above the dam there will be a temporary lake; but sooner or later the dam will give way and the water will escape. More permanent lakes of the same kind are produced by the dams constructed by beavers across some of the rivers of north-western Canada.

Lakes due to Erosion.

Lakes sometimes lie in hollows which are completely surrounded by solid rock. They are not dammed by accumulations of loose material, but are in actual rock-basins. Such lakes may exist in the course of a river; but it is evident that the hollow was not worn by the corrasive action of the river itself. For unless the hollow is very shallow and the river very turbulent the water at the bottom must be practically still, and deposition instead of corrasion must take place. It is only at the foot of a waterfall or where the water swirls round in an eddy that a river can corrade a hollow below the general slope of its bed. Such hollows are always small and it is not too much to say that no rock-basin of any considerable size or depth can be due to the corrasive action of running water.

If, however, the river flows across an outcrop of rock that is soluble in water, it may by solution form a basin. Even if it makes no actual hollow in its bed, the solvent action of the rain and of the river itself upon its banks may widen the channel to so great an extent as to produce a sheet of water which may be called a lake. It is probable, for example, that the expansions of the Shannon known as Loughs Ree and Derg are due to solution of the limestone over which the river flows.

Erosion by wind. Though water ceases to corrade when in a hollow, there are other agents which are not limited to the same extent. Both wind and glaciers can carry fragments upwards and consequently they can corrade when moving up a slope. Either of these agents, therefore, can erode a rock-basin of greater or less extent.

Where there is no vegetation small depressions are often produced by the irregular erosive action of the wind, and during a storm of rain they may be filled with water. This, however, is in desert regions and permanent lakes will not be formed. In moister climates the influence of the wind is less ; but the weathering of rock-surfaces is often very uneven, and since the weathered material is usually soft it may be removed by the wind, leaving hollows which are capable of holding water. These hollows, however, are small and shallow, and it is not possible to point to any rock-basin even of moderate size that has been formed in this way.

Erosion by glaciers. Some writers assert that glaciers are almost ineffective as erosion agents. They allow that a glacier may remove any loose and weathered material over which it flows, but they deny that it is capable of eroding a rock-basin except, perhaps, a very small one. It has, however, been pointed out already that lakes are most numerous in glaciated regions. Many of these lakes owe their origin to the damming of river-valleys by glacial deposits, but many others lie in basins hollowed out in solid rock. Loch Coruisk in Skye may serve as an example. It is entirely rock-bound, and soundings show that it consists of two distinct basins separated by a rocky barrier. No valley worn by water can have this form, for the floor of such a valley must slope continuously to the sea. Glacial striations, *roches moutonnées* and other signs prove that the valley was once occupied by a glacier ; and it is natural to attribute the erosion of the basins to the action of the glacier.

It must not be supposed, in cases such as these, that the valley itself was excavated by ice. There was a river-valley first. In the Glacial Period it was occupied by a glacier. The glacier did not make the valley but modified its form. In particular it hollowed out the two rock-basins, which cannot be due to the action of the original river.

Solution. When a river flows over rock that is soluble in water, a lake, as already explained, may be formed by solution of the rock. But solution does not take place only at the surface. Underground streams dissolve limestone, salt and other soluble material, and often excavate long and spacious caverns. If the roof of a cavern falls in there will be a depression in the ground above and in this depression water may accumulate as a lake.

Many of the meres of Cheshire are probably due to this cause. Beneath the surface there are beds of salt, and the presence of brine-springs shows that the salt is being removed by underground water. Where the process is hastened by pumping up the brine, subsidences frequently occur, and from time to time new lakes are formed or the old ones increase in size.

Deposits of salt are comparatively rare and it is in limestone districts that the effects of underground water are most frequently observed. But although the limestone is dissolved and depressions are produced in the ground above, it is not often that lakes are formed ; for owing to the fissured nature of most limestones the water can usually escape by some underground channel. Occasionally, however, the opening of the channel may be plugged by glacial or other material, and the water will collect in the hollow. Or if there are beds of shale as well as limestone, impervious hollows capable of holding water may be formed.

Volcanic explosions. In the Eifel district there are many circular cavities in the ground produced by volcanic explosions. Sometimes they are surrounded by volcanic ash, sometimes only by the fragments of rock thrown out by the explosion. They are craters, but craters which have been hollowed out rather than built up; and since their walls are of solid rock they can hold water.

The majority of crater-lakes lie in craters of this type. The Laacher See in the Eifel district and Lake Avernus near Naples are well-known examples.

Lakes due to Earth-Movements.

It is easy to see that a hollow capable of holding water can be produced by bending or fracturing of the earth's crust, and there is no doubt that some lakes owe their origin to this cause.

The Dead Sea is one of the best examples. It is not dammed by loose material but lies in a true rock-basin. The surface of the water is 1292 feet and the bottom 2592 feet below the level of the Mediterranean. No such depression as this can be due solely to subaerial erosion, for no erosive agent could remove material from so deep a hollow. The valley of the Jordan is a narrow strip of country which has sunk down between two parallel faults or fractures. From the Waters of Merom southwards for a distance of more than 150 miles this strip is now below the level of the sea; but it did not sink down evenly. The deepest part of the depression is at the Dead Sea, and it is there that the water collects.

Lakes Nyassa and Tanganyika and many other smaller lakes in Eastern Africa lie in similar faulted strips and have been formed in a similar way.

Probably the Great Lakes of North America are also partly due to earth-movements, but the earth-movements were of a different type. Around the shores there are terraces formed when the water stood at a higher level. Originally these terraces must have been horizontal, but they are not horizontal now. The crust of the earth has bent since they were formed. Such bending must certainly modify the shape of the lakes and may have been the original cause of their formation.

There is some evidence, too, that the larger lakes of the Alps are due to bending of the surface. With few exceptions they lie in river-valleys just where the valley opens out upon the plains. It has been suggested that the valleys, like normal river-valleys, once sloped continuously to the sea, but the weight of the mountain mass caused the earth's crust to sag beneath it. Owing to this sagging the slope of the valleys at the foot of the hills was reversed and consequently the lakes were formed. In support of this view it is said that near the lakes the old river terraces, which must have once sloped towards the sea, now slope towards the mountains.

APPENDIX

(1) The Units employed in the Daily Weather Report.

Until April 30th, 1914, the Meteorological Office in its Daily Weather Report made use of the ordinary English measures. The temperature was given in Fahrenheit degrees, the rainfall in inches, the velocity of the wind in miles per hour, and the barometric pressure in inches of mercury. But on May 1st several changes were introduced. The Fahrenheit scale is still employed for temperature, but the rainfall is given in millimetres and the velocity of the wind in metres per second. It is, however, in the method of recording the barometric pressure that the most important change has been made; it is no longer given in inches, but in "millibars."

In the C.G.S., or centimetre-gramme-second, system, which is now almost universally employed in scientific work, the unit of force is the *dyne*. In order to understand clearly what is meant by a dyne, it will be simplest to imagine a mass of one gramme at rest, entirely apart from all other masses, unaffected by any force, and free to move in any direction. In order to cause this mass to move, force is required; and the rate of movement will depend partly upon the force and partly upon the length of time during which the force acts. If a force of one dyne acts upon the mass for a period of one second, then at the end of that second the mass will be moving, in the direction in which the force acted, at the rate of one centimetre per second.

The unit of pressure is one dyne per square centimetre. This, however, is too small a unit for most practical purposes, and accordingly in meteorology the actual unit employed is not one dyne per square centimetre but 1,000,000 dynes per square centimetre. This is called a "bar," and is equivalent to 29·53 inches of mercury.

A centibar is a hundredth part of a bar, and a millibar is a thousandth part. On the first page of the Daily Weather Report the barometric readings at the observing stations are given in millibars, and on the chart isobars are drawn at intervals of half-a-centibar, *i.e.* of five millibars.

The conversion of millibars into inches of mercury, or *vice versâ*, is a question of simple proportion, and the following table will be sufficient to enable anyone who is accustomed to think of barometric pressure in inches, to make use of the charts.

Millibars	Inches of mercury	Millibars	Inches of mercury
920	27·17	990	29·24
925	27·32	995	29·38
930	27·46	1000	29·53
935	27·61	1005	29·68
940	27·76	1010	29·83
945	27·91	1015	29·97
950	28·05	1020	30·12
955	28·20	1025	30·27
960	28·35	1030	30·42
965	28·50	1035	30·56
970	28·65	1040	30·71
975	28·79	1045	30·86
980	28·94	1050	31·01
985	29·09	1055	31·16

(2) Books.

The following short list is far from exhaustive but may be useful to those who wish to follow up the study of some particular branch of the subject. Most of the references are to works in the English language, but a few of the most important French and German books are also included. Supan's *Grundzüge der physischen Erdkunde* gives an excellent bibliography at the end of each section.

(*a*) *General.*

E. de Martonne. Traité de géographie physique. Armand Colin.
A. Supan. Grundzüge der physischen Erdkunde. Veit and Co.

(*b*) *The Atmosphere.*

The most important aid in studying climate is a series of maps of pressure, temperature, rainfall, etc. Many modern atlases give a useful selection, but by far the most complete collection is to be found in the Atlas of Meteorology by J. G. Bartholomew and A. J. Herbertson, forming Vol. III of Bartholomew's Physical Atlas.

R. Abercromby. Weather. International Scientific Series. Kegan Paul.
W. M. Davis. Elementary Meteorology. Ginn & Co.
J. Hann. Handbuch der Klimatologie. J. Engelhorn.

An English translation of the general portion of this work, by R. de C. Ward, has been published by the Macmillan Co., New York. There is a later German edition, but the translation is still useful.

J. Hann. Lehrbuch der Meteorologie. Tauchnitz.
A. Hildebrandsson and L. Teisserenc de Bort. Les bases de la météorologie dynamique. Gauthiers-Villars.—International Cloud-Atlas. Gauthiers-Villars.
W. L. Moore. Descriptive Meteorology. Appleton & Co.
W. N. Shaw. Forecasting Weather. Constable & Co.
R. de C. Ward. Climate. John Murray.

(*c*) *The Ocean.*

G. H. Darwin. The Tides. John Murray.
G. H. Fowler and others. Science of the Sea. John Murray.
O. Krümmel. Handbuch der Ozeanographie. J. Engelhorn.
J. Murray. The Ocean. Williams and Norgate.
J. Murray and J. Hjort. The Depths of the Ocean. Macmillan & Co.
J. Thoulet. L'Océan : ses lois et ses problèmes. Hachette & Co.

(*d*) *The Land.*

For a proper study of the development of land-forms some knowledge of geology is essential, and most of the text-books of geology deal more or less exhaustively with the subject. The books in the following list do not, for the most part, require anything more than a very limited previous acquaintance with the science of geology.

T. G. Bonney. Volcanoes. John Murray.

W. M. Davis. Geographical Essays. Ginn & Co.

W. H. Hobbs. Earth features and their meaning. The Macmillan Co., New York.

—— Characteristics of Existing Glaciers. The Macmillan Co., New York.

J. W. Judd. Volcanoes. International Scientific Series. Kegan Paul.

A. de Lapparent. Leçons de géographie physique. Masson & Co.

J. E. Marr. The Scientific Study of Scenery. Methuen & Co.

A. Penck. Morphologie der Erdoberfläche. J. Engelhorn.

I. C. Russell. River Development. John Murray.

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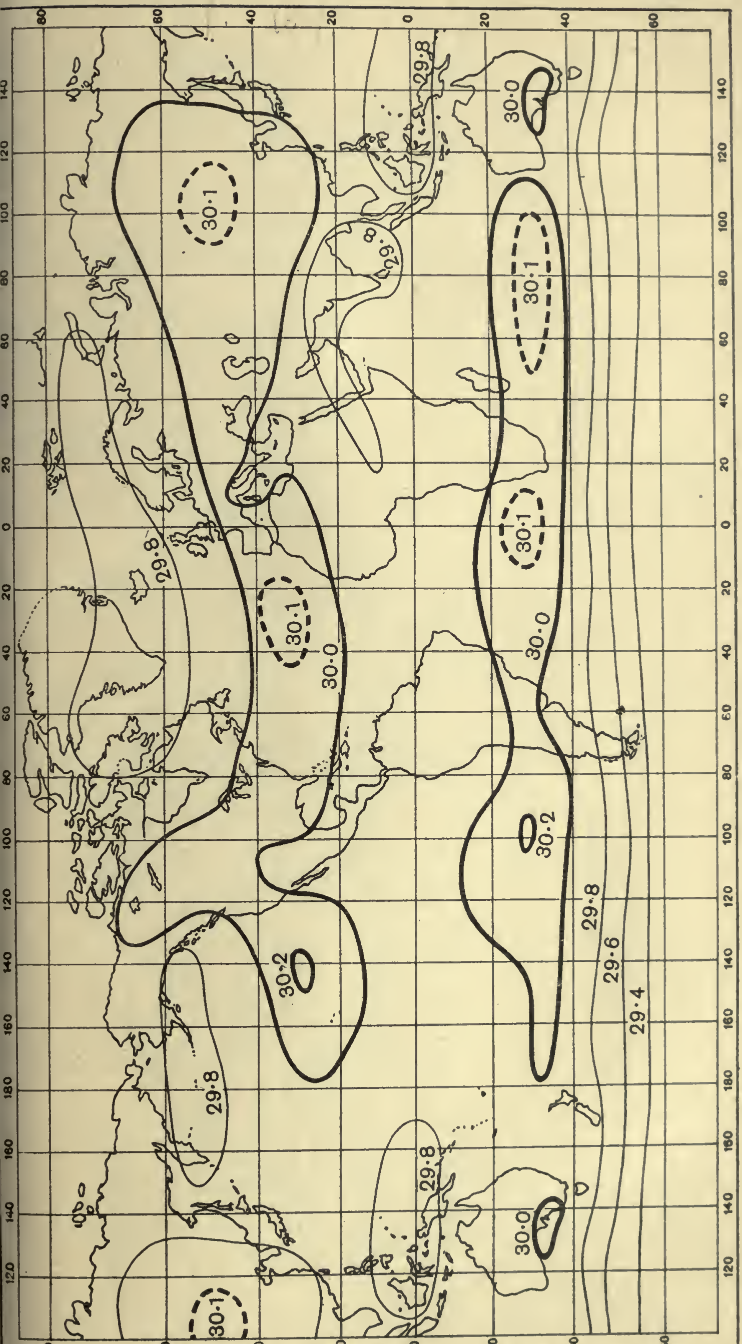
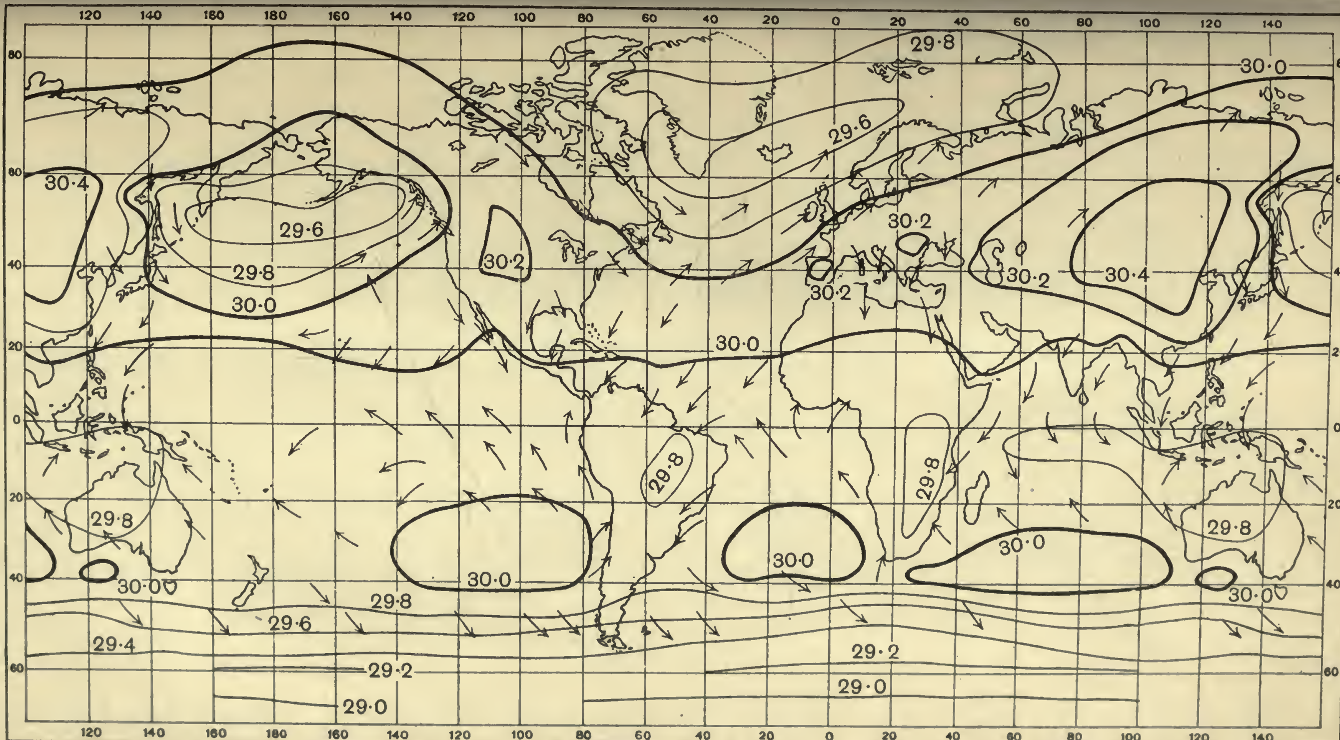


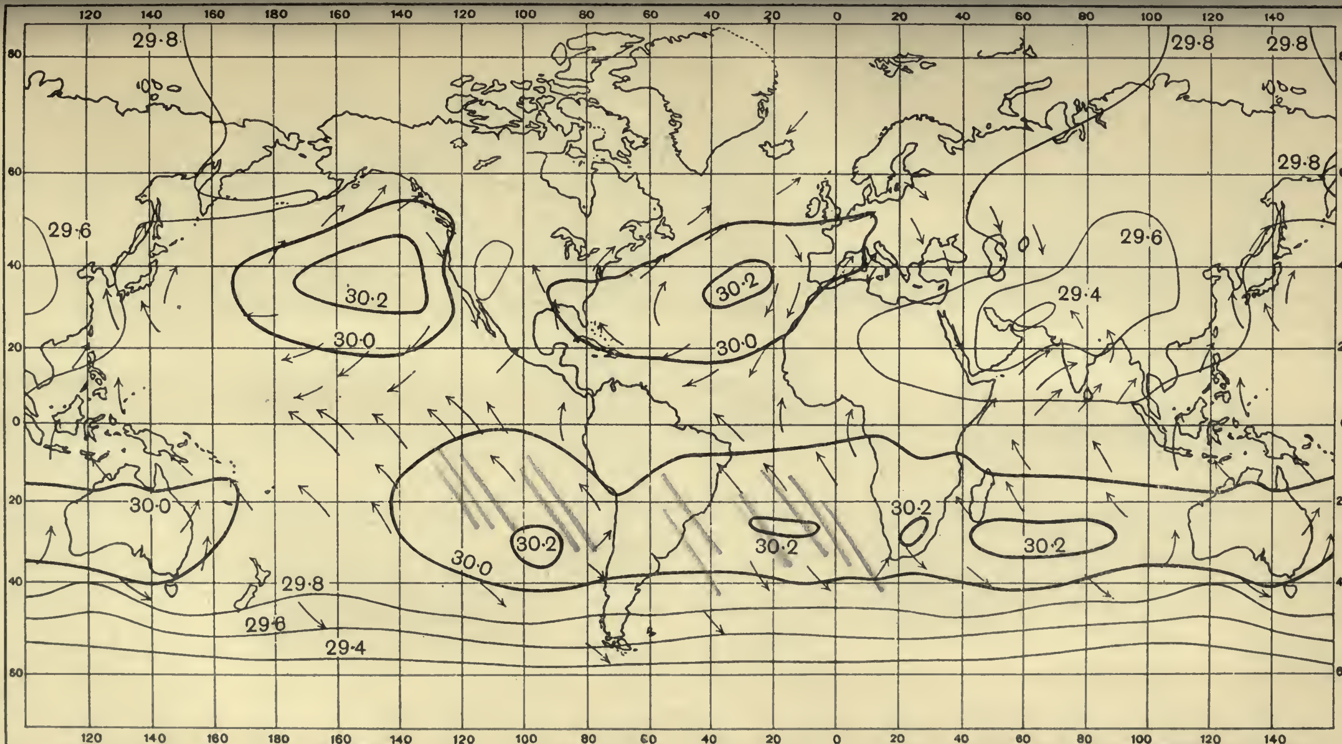
Mean annual isobars.

(After Buchan and others.)



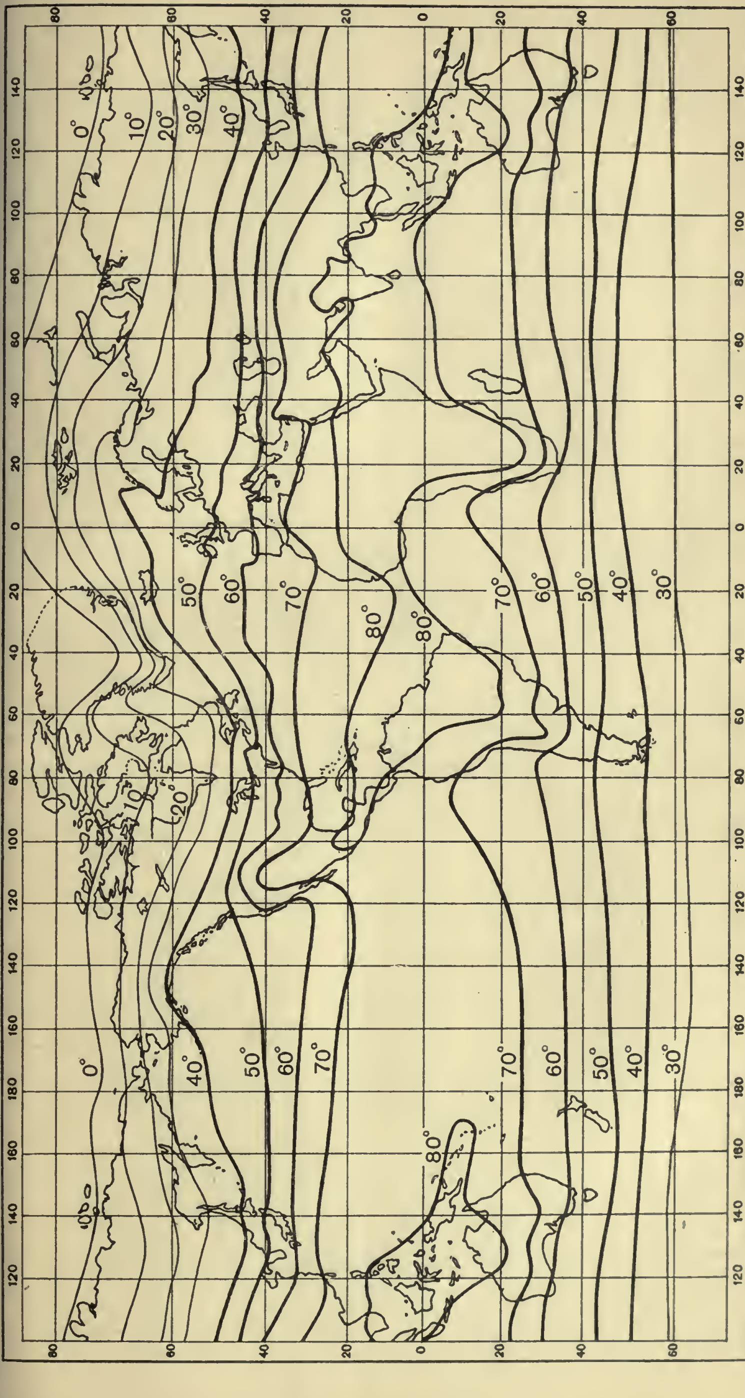
January isobars and winds.

(After Buchan, Mohn and others.)



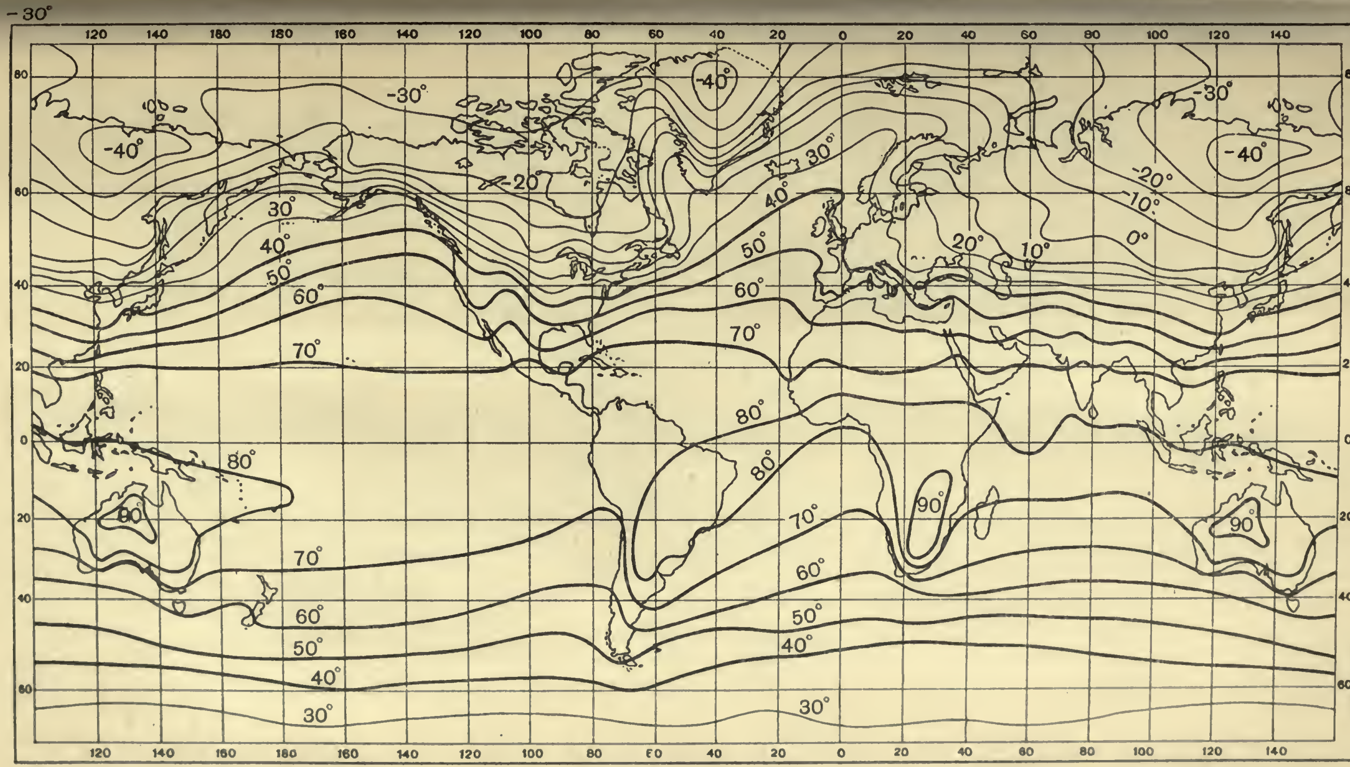
July isobars and winds.

(After Buchan, Mohn and others.)



Mean annual isotherms.

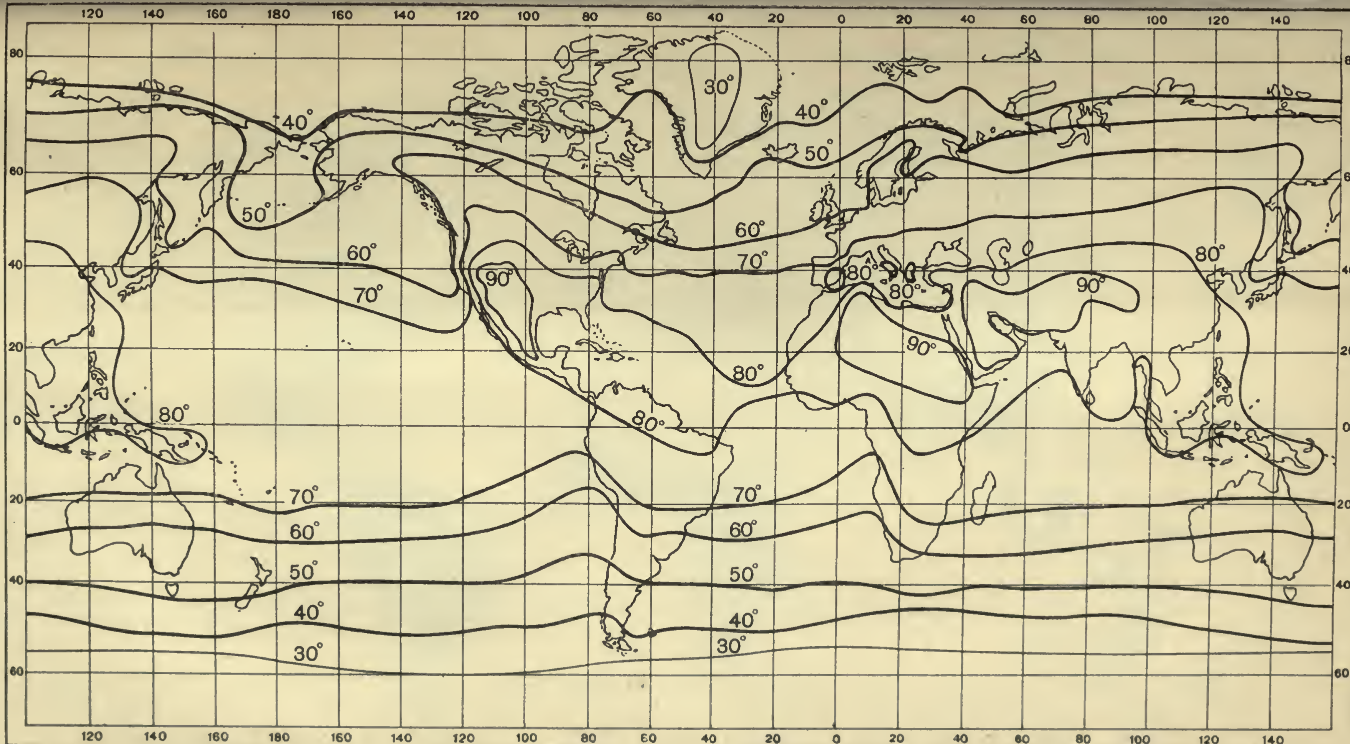
(After Buchan, Mohn and others.)



January isotherms.

(After Buchan, Mohn and others.)

Map 6



July isotherms.

(After Buchan, Mohn and others.)

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