CHAPTER NINE

The General Circulation of the Atmosphere and Oceans

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IN THE previous chapter the point was made that in order to make full use of climatic data we have to know how the atmosphere *works*. We shall now consider what is as yet known about the mechanism, for although the general circulation is only one aspect we find that we cannot understand it without knowing much about other, more detailed, aspects. It will be best to begin with a definition. The general circulation is the global aspect of the motion of the atmosphere. We concentrate our attention on broad air-current systems, ignoring for the moment all the complicated and rapidly varying details. Though the emphasis is on wind systems in the first instance, we do not ignore the aspects of weather with which we are usually more directly concerned. It is merely that winds are a convenient basis for discussing associated changes of cloudiness, rain and temperature. We have noted earlier that from the point of view of their *dynamics*, the oceans and the atmosphere must be considered together as components of a single system. Nevertheless, it will be convenient to begin by concentrating our attention on the atmosphere.

THE ATMOSPHERE

The term 'general circulation' is to some extent misleading, in that it suggests a *steady* flow of air. Now, when we were attempting to define climate we found that, no matter what period we averaged over in order to eliminate rapid fluctuations, we still found slow temporal variations of the smoothed elements. Similarly, when we average over *space* in order to smooth out local fluctuations and make clear the broad pattern of the flow, we find that this varies with time, no matter how large the area over which we smooth, and there is still variation if we average over a period as well. At any one instant we may think of the air motion as composed of superposed flow patterns of different 'scales'. Since the motion often appears to be rather like the circular motion in an eddy, we may rather loosely refer to the patterns, or parts of patterns, as 'eddies'. The scale is then defined as the average horizontal size of the eddies.

For present purposes we shall ignore the very little eddies and even those some ten miles across associated with local showers and thunderstorms. The largest eddies, which we shall later have to study in detail, are the weather systems, showing near the ground as depressions or anticyclones some one thousand miles across. The general circulation is disclosed by smoothing out all eddy-motion up to and including the scale of weather systems. Since the depressions, and their opposites the anticyclones, move about, we may smooth them out by averaging over space or time or both. Usually time averaging, over five days or more, is preferred since it gives a more sharply defined picture. The resulting pattern changes with time more slowly than even the largest eddies which have been smoothed out, and in quite a different way. Though the currents vary in shape and intensity they have some features which seem to be always, or almost always present. Not surprisingly there are systematic changes with the seasons but there are also changes, some comparatively slow, which are not simply seasonal. They are evidently related to differences in weather between one year and the next and, on a longer time-scale, to changes in climate.

Let us first consider those features of the general circulation which are always, or nearly always, present. Because they are permanent they show up strongly when we average over a long period—say fifty years. The flow pattern near the ground is familiar to most of us and is shown in many atlases. What concerns us at the moment is not the detail, which is complicated, but the broad features. We may note:

(i) A broad belt of easterly winds between latitudes 30°N and 30°S. The flow is weaker near the boundaries and near the equator.

(ii) A broad belt of *generally* westerly winds between 30°N and 60°N and a corresponding belt between 30°S and 60°S.

There is some seasonal variation, the whole pattern moving a few degrees of latitude towards the summer hemisphere, but this motion is much less than the apparent motion of the Sun ($23\frac{1}{2}°$ lat.). There is also a seasonal change which is evidently related to geography, particularly to the large land mass of Asia. In winter there is an anticyclonic flow (i.e. the winds correspond, in the northern hemisphere, to clockwise rotation of air) over Asia as a whole, while in summer there is a generally cyclonic (anticlockwise) flow over southern Asia and the north Indian Ocean. These flow patterns are called *monsoons*. The broad belts of easterly (*trade*) and westerly winds which otherwise dominate are called *zonal* because they blow along lines of latitude. It is true that the trade winds are north-easterly and south-easterly in the northern and southern hemispheres respectively but we must remember that we are concerned with the flow pattern through the *whole* of the atmosphere. At a height of one mile above the ground the *average* flow is much more nearly zonal. Similarly the westerlies, which are, on the average, south-westerly near the ground in the northern hemisphere (north-westerly in the southern hemisphere) become almost westerly as we rise in the atmosphere. For a reason we shall speak of later, all winds become, on the average, more strongly westerly as we continue to rise, so that at a height of 7 miles the zonal winds are westerly almost everywhere. Somewhat above this level at about latitude 30° (N or S) we encounter the strongest average winds in the lower atmosphere. The maximum *average* wind there is about 100 miles per hour. These two strong wind belts are called *jet-streams*. At higher levels (in what is called the *stratosphere*) the winds decrease before beginning to increase again.

We can understand much about what happens in the lower atmosphere (called the *troposphere*) without knowing much about what happens above it, so we shall postpone a discussion of stratospheric currents.

It is instructive to see how associated features of the general circulation, pressure and temperature, fit into the pattern we have described. We know that the Earth is an oblate spheroid because the centrifugal force associated with its spinning motion acts most powerfully at the equator, tending to force it away from the axis of rotation. Now a westerly wind is air rotating more rapidly than the Earth and the *extra* centrifugal force would push it towards the equator were there not an equal and opposite force acting to keep it in position at its original latitude. This additional force is supplied by a difference in the pressure of the air, the pressure being greater in the equatorward direction. Similarly, an easterly wind can persist only if pressure increases towards the poles. Our description of the zonal winds is therefore *consistent* with the observation that there exist two belts of high pressure, at latitudes 30°N and 30°S respectively. We have *not*, of course, explained *why* the pressure belts and zonal winds exist. Near each of the poles the pressure rises again, implying the existence of generally easterly winds at about latitude 70° (N or S). Rather careful averaging is needed to verify, from the wind data, that these rather feeble *polar easterlies* actually exist. A wind, *calculated* from the distribution of pressure, is called a *geostrophic* wind. It is a fairly good approximation to the true wind only if the wind is a large-scale average and it is not changing too rapidly. There are rather subtle reasons, which it would take too long to discuss, why the formula used is somewhat inaccurate (especially in low latitudes) when applied to *time-averaged* charts. We may apply similar ideas to winds which are not zonal. An immediate deduction from the geostrophic formula is that cyclonic winds blow round regions of low pressure and anticyclonic winds round regions of high pressure in *both* hemispheres. We may check this against the monsoon circulations we have described. It is worth remembering that cyclonically rotating air is spinning in the same direction as, and therefore faster than, the Earth, whereas anticyclonic rotation is in the opposite direction and is therefore slower than the Earth. The rotation of the Earth we are referring to is its *vertical component*, that is the rate at which the ground beneath us is *spinning* about the *local* vertical. Except at the poles this is *less* than the rate of spinning of the Earth as a whole and depends on latitude. It is zero at the equator: there the ground is turning about a *horizontal* axis.

Temperature and Other Elements

The temperature pattern, like the pressure pattern, is closely related to the wind pattern. In discussing the most obvious general characteristic, the decrease of temperature from the equator to the poles, we start with the advantage that we can account for it quite simply. The poles are colder because sunshine comes in at a lower angle and is therefore, for unit area, less intense.* The pressure decreases more rapidly with height when the air is cold than when it is warm (Chapter 7). Considering the effects of temperature alone (i.e. assuming uniform pressure at the ground) the pressure at a given height (say 7 miles) will be greater above the equator than above the poles. This corresponds to a geostrophic westerly wind at 7 miles in both hemispheres. Such a wind change (due to horizontal variation in temperature) is called a *thermal wind*. The true wind is obtained by adding (algebraically in this case, more generally, for winds in any direction, vectorially) the winds near the ground, which in this case are comparatively small. Hence the winds consistent with the temperature pattern agree broadly with the observed winds. To the extent therefore that we have explained the *temperature* pattern we have therefore explained the zonal westerly winds aloft. We have tacitly assumed that the temperature difference persists as we go up and this is found to be true throughout the troposphere. The second generalization we can make about the temperature pattern is that, in the troposphere, temperature decreases with height. The division between troposphere and stratosphere is *defined* in terms of the way temperature varies with height: in the lower stratosphere it varies little with altitude. Consistent with the dividing point (called the *tropopause*) being lower in high latitudes we find the lower polar stratosphere (except in winter) to be a little *warmer* than the equatorial stratosphere. This agrees with the westerlies decreasing with height in the lower stratosphere.

As we should expect from the change with latitude of the annual variation in the length of the sunlit period, the annual variation of temperature is much larger in high latitudes than in low latitudes. A consequence is that the westerly winds aloft are much stronger in winter than in summer. We cannot however be satisfied with this as a complete explanation, because if we were concerned only with direct solar heating, then during, and especially towards the end of the polar night, the polar regions should become very much colder than they are observed to be. Their comparatively mild climates arise from warm winds bringing heat from lower latitudes. We are immediately confronted with an important characteristic of the general circulation. Narrowly interpreted as a

* When we try to calculate *how much* colder we find that this problem is just as complicated as, and closely related to, all the other problems we shall discuss.

system of long-term average currents it is an incomplete mechanism. For zonal winds cannot transfer heat between latitudes*, and this heat transfer is not only one of the most important things we want to calculate, but it affects, through the thermal wind relationship, the zonal winds themselves.

Leaving this problem on one side, let us glance at the non-zonal annual temperature change. This is much larger over land than over sea and we may relate it to the changes in monsoon winds. In *this* case high summer temperature is fairly distinctly related to low surface pressure and low winter temperature to high surface pressure. From the thermal wind relationship it may be seen that the monsoon flow patterns die out as we ascend and become quite different in the upper troposphere. This does not explain the *existence* of monsoon patterns nor has the relation found between surface pressure and temperature any *general* validity.

Finally let us examine the patterns of cloud, rainfall and evaporation. To a considerable extent these vary in the same way. We may note a marked correlation with the pressure at the ground (strictly, the pressure 'reduced' to mean sea-level), heavy rainfall in particular being associated usually with low pressure, and, therefore, cyclonic mean winds. Thus there is a belt of heavy rainfall near the equator, a comparatively dry region including many deserts in the sub-tropical high pressure belt near latitude 30°, and another rainy belt near and on the equatorial side of the low pressure belt near latitude 60°. Since cold air cannot take up so much water vapour as warm air the annual rainfall and the annual evaporation decrease *on the whole* with increasing latitude, but cloudiness is not affected in the same way.

The effect of geography is even more marked than in the case of temperature so that the above picture is very crude indeed. Nevertheless, though crude, it is of value, for it indicates a relation between rainfall and the more mechanical aspects of the general circulation we considered earlier. Further evidence is the North-South annual motion of the rain belts with the pressure and wind patterns (though to a slightly greater extent). The relationship is actually much closer than has so far appeared because cloud and rainfall are directly related to *vertical* motion of the air.

* The actual deviation of mean currents from latitude circles complicates but does not basically affect the conclusion.

Clouds, from which rainfall descends if they are sufficiently thick, are formed when the air expands and cools as a *consequence* of the vertical motion of the air. Unfortunately it is not possible to measure the upward drift over a large area because it is both irregular and very small—less than one hundredth, on the average, of the horizontal flow. It is possible to *calculate* it however, though calculations are not yet sufficiently detailed or accurate to enable us to present a complete picture. We can make these calculations from a knowledge of the horizontal wind alone, so that to this extent rainfall appears as a by-product of the general circulation. However, the production of cloud and rain entails a release of latent heat which alters the temperature of the air so that we cannot obtain an accurate idea of the mechanism of the general circulation, without including rainfall and evaporation as an integral part. Evaporation affects the mechanism mainly through the absorption of latent heat associated with it. In so far as it is dependent on wind and temperature, evaporation, in turn, depends on the general circulation.

THE OCEANS

We are concerned here only with those aspects of the motion and thermal effects of the oceans which are intimately connected with the atmosphere. Waves are raised by the wind, but we shall not discuss them as they are dealt with elsewhere (Chapter 5). The tides are not appreciably connected with weather but there are large-scale currents and drifts which we must consider in some detail.

Not only are they caused by wind, cooling by the air or evaporation into the air, but they react back on the atmosphere by moving large quantities of heat, which ultimately enter the air at a place determined partly by the strength of the ocean currents. These currents constitute the general circulation of the oceans.

Though the details are different, the circulations of the oceans and atmosphere are processes of the *same kind* and much of what we have said about the atmosphere applies equally well to the oceans. In particular, the relationship between ocean currents and pressure and temperature at different depths in the sea is exactly similar to the corresponding relationships (geostrophic wind and thermal wind) in the atmosphere. The only difference is that density of sea water depends not only on temperature but also on the quantity of salts dissolved in it—that is on its *salinity*. Hence the thermal wind corresponds to the *thermohaline* ocean current, the decrease of current with depth associated with horizontal variation of density. It may also be noted that the long-term average currents in the oceans are no more a satisfactorily complete description of long-term behaviour than are the average winds in the atmosphere. Heat, salt and other things are transported *across* the mean currents by eddies which smooth out in the long-period mean, just as heat, water vapour, etc., are transported by large atmospheric eddies (weather systems) across the climatic-average winds. We can in fact assume ocean current behaviour to be *similar* to that of air currents unless there are definite reasons why behaviour should be different.

The most important factors, which cause the ocean circulation to be different from that of the atmosphere are as follows:

(i) Neglecting the very small heat supply from radio-activity in the Earth, the oceans are heated, by sunshine and by exchange of heat with the atmosphere (evaporation, 'convection'—to be explained later) at the *top*. The atmosphere, as we shall see, is heated mainly at the *bottom*.

(ii) Friction, associated with the setting up of surface gravity waves on the sea, acts as a *brake* on the atmospheric circulation. It is the primary *driving force* of ocean currents.

(iii) The oceans are bounded, right up to the top, by continents. Mountain ranges are more limited and extend only part of the way up in the atmosphere.

(iv) There is no significant analogue of cloud formation and consequent release of latent heat inside the oceans. In fact, all the processes which affect the density of sea water (including rainfall, run-off and the melting of ice which dilute the salt-content and evaporation and freezing which concentrate it) occur at the top.

Other factors, such as the much greater density and heat capacity of water as compared with air, and its much smaller motion and compressibility are merely quantitative. It is curious and significant that the relative magnitudes of many quantities (such as horizontal and vertical variations of density), which are determined by mutual adjustment as the mechanism operates, attain values such that corresponding processes in the ocean and in the atmosphere are often of equal importance.

Let us now look at the observations, concentrating our attention on the broad features. Most obvious are the surface currents which consist mainly of *gyrals*, the water circulating anti-cyclonically about centres situated in the sub-tropical atmospheric high-pressure belts. (Fig. 1). Typical is the North Atlantic gyral consisting of the North Atlantic drift, towards the East, most

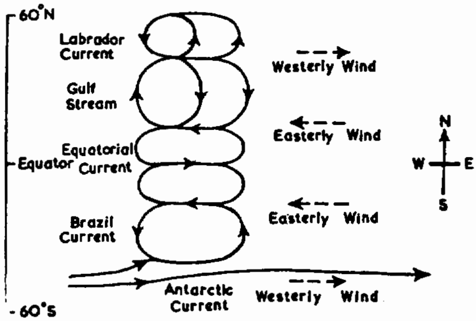

FIG. 1. Surface gyrals in Atlantic and Antarctic current in relation to wind stress.

clearly indicated near latitude 40°N on the western side, and a general drift towards the West below latitude 30°N. The circulation is completed by a very narrow and strong current towards the North, called the Gulf Stream, just off the American coast (at the edge of the continental shelf) and a general slow drift towards the South in other longitudes. There are similar circulations in other oceans, that in the North Pacific being particularly strong. They are evidently caused by the 'twist' of the wind stress, which acts towards the East poleward of the sub-tropical (Azores) anti-cyclone and towards the West in the trade wind belt. A calculation shows that currents should be produced in roughly the regions where they are observed and it also accounts for the strong concentration on the West side of the oceans (east coast of continents). Poleward of the latitude of strongest westerly winds (about 50°N) there are, in the northern hemisphere, cyclonic gyrals. That in the North Atlantic includes the Greenland currents and the Labrador current. The trade wind stress piles up water on east continental coasts. This yields not only the 'head' of water which drives the Gulf Stream and similar currents but it also forces water back towards the East near the equator (the equatorial currents) where the trade wind stress is weakest. The southern hemisphere is peculiar in that in a range of latitudes, between the tip of South America and the tip of Graham Land (Antarctica), there is no land against which the westerly wind stress can pile up water. No gyral is set up, the wind stress accelerating the water until an equal and opposite stress is set up, due to the motion of water over the bottom of the ocean. This Antarctic current flows, roughly along lines of latitude right round the Antarctic continent, though at a considerable distance from it. All the currents we have described decrease as we go down. In *most* places the motion of water near the bottom of the ocean is probably very sluggish.

Less well known, and less well understood, is a different kind of drift called the 'vertical circulation' which is most marked in the Atlantic Ocean. Although the most evident motions are the gyrals, there is superposed on these in the Atlantic a general drift near the top, from South to North and a compensating drift at greater depths from North to South (Fig. 2). The circulation is completed by water sinking in the northern part of the North Atlantic (somewhere near Iceland), and rising in the Antarctic, somewhere near the Antarctic current. Because the drift is very slow*, compared with other motions it has not been measured directly. Sections in vertical planes along parallels of latitude (the planes in which the water particles move) indicate quite clearly, however, the existence of this circulation. At lower depths temperature and salinity act as 'markers' indicating the origin of the water mass and there is a well marked bulge of *comparatively* warm saline water at great depths—greater than a mile—pointing towards the South. At the surface there is displacement of the region of maximum salinity, caused by evaporation in the sub-tropics, towards the North. A circulation of this type, corresponding to rotation about an E-W axis, is called *meridional*.

There exist meridional circulations in the atmosphere, the most marked being the Hadley *cells* in which air rises near the equator and comes down on either side near latitudes 30°N or S (Fig. 3). Near the ground there is a general drift towards the equator consistent with the very marked equatorward component of the trade

* A water particle would probably take hundreds of years to complete the circuit.

winds. Between 30° and 60° latitude in either hemisphere there are other meridional atmospheric circulations, sometimes called the Ferrel cells, which are reversed in direction as compared with the Hadley cell of the same hemisphere.

We may note two ways in which the oceanic meridional circulation differs from the atmospheric ones. Firstly, the atmospheric circulations are nearly symmetrical about the equator, whereas the Atlantic circulation is not—perhaps connected with the fact

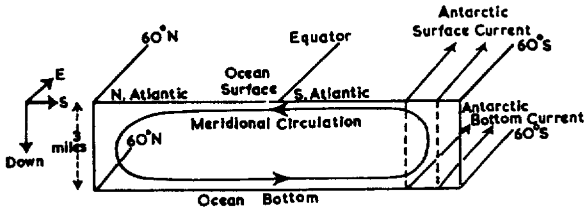

FIG. 2. Meridional circulation in Atlantic ocean in relation to Antarctic current. The figure is not to scale and Earth's curvature is neglected.

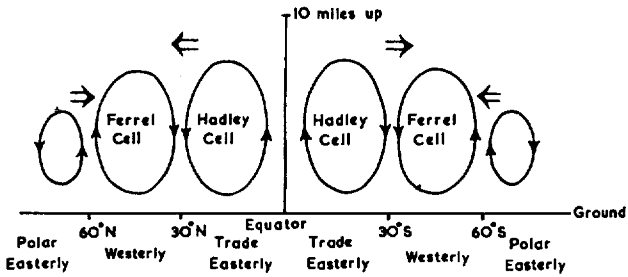

FIG. 3. Meridional circulations in the troposphere. Broad arrows indicate eddy transport of angular momentum. Surface zonal winds indicated below figure.

that the Antarctic current has no analogue in the northern hemisphere. Secondly, the atmospheric drifts cannot be geostrophic since there are no boundaries against which a pressure difference could be built up. The Atlantic circulation (north of the tip of South America) *could* be geostrophic, so that the thermo-haline current, yielding the *difference* between the upper northward current and the lower southward current, may be used, together with the fact that the *total* northward flow is small (and can be estimated), to estimate the strength of the circulation. The results are not very reliable because of complications near coast lines but they are in general agreement with estimates obtained by other means. The interesting conclusion emerges that the meridional oceanic circulation may, because the large temperature contrast between top and bottom compensates for the smallness of the current, transfer as much heat to the northern end of the Atlantic as does the Gulf Stream.

To some extent the distribution of temperature and salinity in the oceans is implicit in our description of the currents. Surface temperature decreases from the equator towards the poles for the same reason as it does in the atmosphere. Since sea and air are in contact their surface temperatures cannot remain very different for long. Surface salinity is greatest in the dry belts of the atmosphere, where evaporation minus rainfall is a maximum, and least in the wet regions. At great depths the oceans are everywhere very cold being indeed only slightly above freezing point. Although a decrease in temperature is in many regions partly compensated by a decrease in salinity we find everywhere the heaviest water at the bottom.

Variations due to all causes are largest near the top; at great depths the oceans are more nearly uniform. Surface actions, (solar heating, evaporation, wind stress) affect *directly* only a comparatively thin top layer. Elsewhere the distribution of temperature, salinity etc., is determined in a process of mutual adjustment in which water movements are involved. Some of these movements are the direct consequence of surface actions. For example, as equilibrium is approached, a surface stress sets up circulations in the water beneath which are in planes *at right angles* to the stress. The effect of the stress is thereby communicated to great depths, setting up currents there *in the direction of* the stress.

The mechanism can be understood by considering the particular case when the circulations are meridional. For motions of *this* kind there is a theorem which tells us that the angular momentum (mass times velocity of rotation about the Earth's axis times distance from the axis) does not change with time. Now when a particle moves North or South its distance from the Earth's axis changes. If it moves to a lower latitude the final result is that its velocity *relative to the Earth* must decrease if we regard motion towards the East as positive. Hence with a stress towards the East (due to a westerly wind) we can have a steady state if the increase in speed due to the stress is just balanced by the decrease due to equatorward motion. Now in the lower portion of the meridional circulation the water is moving poleward, and this necessitates an increase in the current towards the East, i.e. the effect is as if the stress acted at this lower level. Admittedly this merely shows how the effect of a surface stress *could* be transmitted to great depths. A detailed calculation shows that such a circulation *must* be set up. It also shows that a similar effect is obtained whatever the direction of the stress. An immediate consequence is that the anticyclonic stress which produces for example the Gulf Stream gyral, sets up circulations in which warm saline surface water is *forced* down thousands of feet in the middle of the gyral. The exact shape of the circulation and the currents set up at great depths is determined by the fact that at each stage the thermohaline current relationship must be satisfied. Warm saline water at great depths in the region mentioned is in fact observed. It could not have got there by any other process than the one we have described.

The forcing down of light surface water as a result of the wind stress increases the potential energy of the oceans, i.e. it generates energy of position which could be transformed into energy of motion (currents) by a suitable displacement of the particles. Actually this transformation takes place automatically. By trial and error, as it were, the ocean finds motions which enable eddy-currents to grow at the expense of the potential energy. Once started, no matter how small they may be initially, the eddies continue to grow until they attain a definite maximum size. Another way of describing the state of affairs is to say that the initial state is *unstable*. The condition for this kind of instability in a rotating fluid is that there should exist a *horizontal* variation of density and the associated thermo-haline current or thermal wind. A fluid in this state is called *baroclinic*. Now the atmosphere is, as we have seen, made baroclinic *directly* as the result of unequal solar heating at different latitudes. Hence the same kind of eddy ought to exist in the atmosphere. It does: the eddies are in fact the weather systems we have referred to before. We shall have more to say about the atmospheric eddies later. A general property of these eddies which can be proved theoretically is that they transport heat (or whatever else is the factor causing low density such as *defect* of salinity) horizontally towards the high density region and upwards against gravity. The result therefore of oceanic eddies developing on the sides of the masses of light water forced down by the wind stress is that, in the steady state, the base of the light water is 'eaten away' as fast as it is forced down.

While we are on the subject of oceanic circulations we may, in the light of what has been said, recall the meridional circulation in the Atlantic. It is clear that the westerly wind stress in the Southern hemisphere due to the Roaring Forties and Fifties must set up a meridional circulation reaching right down to the bottom where the increase in bottom velocity due to poleward motion is cancelled by friction on the ocean floor.* Now at the equatorward edge of the Antarctic current the meridional flow is towards the North at the top and towards the South at the bottom, i.e. in the same direction as the Atlantic meridional circulation. Instead of descending at the edge of the Antarctic current, it appears that the water moves right across the equator to high northern latitudes before sinking. It is possible that the water chooses this path because it is the one of least resistance. Surface water forced down near latitude 40°S would tend to spring back because it is lighter than the water below. On the other hand near Iceland the water is nearly uniform right to the bottom and the water can sink without resistance. The question remains: "Why is the circulation concentrated in the Atlantic and not observed (with certainty) in other oceans?" It is conceivable that the answer has something to do with the Atlantic being situated immediately to the lee of the constriction of the Antarctic current between South America and Antarctica. Some distance downstream from the constriction there is a submarine ridge, of which the South Sandwich islands are peaks, extending right across the Antarctic current. The bottom stresses in this region, and consequently the meridional circulation, must be much greater than average. The whole problem needs further investigation.

The factors we have considered in this brief study of the ocean currents are purely mechanical. From this point of view special substances dissolved in the sea such as phosphates and oxygen, though they are essentials of life, are of direct interest only to the extent that they affect density, or as 'markers' for following water masses. Nevertheless a precise understanding of the mechanical

* This 'friction' is not of the ordinary type. Water flowing over submarine hills and mountains sets up internal waves which carry away energy and very effectively brake the main current.

factors would enable us to predict the concentration of all factors affecting suitability of the sea to particular forms of life—a matter which may be of some interest to fishermen.

HEATING

Let us now return to the problems of the atmospheric circulation. The general nature of the mechanism is clear. Unequal heating makes the air baroclinic and therefore unstable. The large eddies, which develop as a result, transport heat upwards and in the direction towards the poles. The heat carried by the eddies reduces the temperature in low latitudes somewhat below that which would be observed if only direct solar heating were involved and increases the temperature in polar regions. This eddy-transport of heat explains the fact that polar temperatures do not fall very low indeed during the polar night. To understand the process more accurately we must study the solar heating process in more detail. Part of the heat from the Sun is reflected back from the tops of clouds and never reaches the Earth. A small part of what remains is absorbed by the air but most of it reaches the surface. Some of this is reflected back (especially in high latitudes from snow or sea) into space. The remainder is absorbed in the top layers of ground or ocean. Thus most of the heat which comes into the atmosphere does so through the land or sea surface.

It is not however enough to consider only the incident heat. This by itself would cause a steady rise in temperature of the Earth as a whole and we know that the surface temperature of the Earth has not changed very greatly over many hundreds of millions of years. Hence on an average the Earth must lose heat at almost exactly the same rate as it gains heat. It does so in the same way as the Sun does, by radiation, the only difference being that Earth-radiation is long-wave (infra-red) and therefore invisible. Because the atmosphere is partly opaque to long-wave radiation (primarily because it contains water vapour) the effective cooling surface, that is the surface from which radiation can escape into space without being re-absorbed, is not on the ground but some distance up in the air, at an average height of about 4 miles. The exact position of the cooling surface is the level above which the total opacity of the air has a definite value, and this is closely related to the temperature at the cooling level itself, because the amount of water vapour the air can contain depends only on its temperature.

It follows that the cooling level is rather higher in low latitudes and also, since heat radiated by a given substance depends on temperature alone, that the outgoing Earth-radiation varies much less with latitude than does the incoming solar radiation. Hence although the *total* incoming and outgoing radiations balance they do not do so at every latitude: in low latitudes there is a net gain, in high latitudes a net loss, the change-over occuring at about 35° latitude (N or S). There must be a long term balance everywhere, so that the excess heat in low latitudes must in some way find its way to high latitudes.

The only mechanism which can carry the heat fast enough is transport by a moving fluid. In fact the larger part of the heat is carried in large atmospheric eddies—the weather systems (Fig. 4).

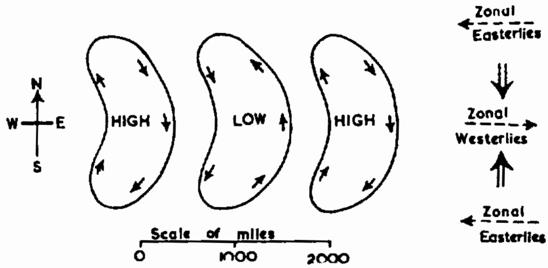

FIG. 4. Flow pattern in train of weather-system eddies (schematic). Full lines denote isobars (or stream lines). Small arrows denote winds. Actual eddies are more complicated and variable, both in space and time. The pattern drifts, irregularly, towards the East. Broad arrows indicate transport of angular momentum due to flow pattern.

The eddies operate a shuttle-service, the air in one part of the eddy being warmed, moving to high latitudes and there radiating away some of its heat. The other part brings back cold air to be warmed in its turn. The inequality in the heat-balance at different latitudes is all the time tending to increase the temperature contrast, the eddy transport to smooth it out. Neither process is quite steady: the heat balance, for example, depends on cloud amount which depends on eddy activity, which depends on temperature contrast, which depends on heat balance at an earlier time, and so on. Hence there exists a fluctuating balance between these two processes. Any semi-permanent change in the equilibrium between them corresponds to a change in world climate.

The smaller part of the heat transport is by ocean currents. Since these are driven primarily by wind-stress, which depends on the atmospheric circulation, we must expect these also to fluctuate. Ocean currents have much more inertia both mechanical and thermal—the density and heat capacity of the oceans is much greater than that of the atmosphere. There is therefore a greater preponderence of slow variations in the sea than in the air. Hence the effect of ocean currents on changes of climate is probably much greater than their effect on short-period changes.

Let us consider now the dependence of heat-balance on height above the ground. We have seen that the radiative cooling occurs *effectively* at about the middle of the troposphere. In fact, due to the complexity of the long-wave radiation spectrum of water-vapour, this cooling is spread over the whole of the troposphere. Even after allowing for the heating due to direct absorption of solar radiation, there is net cooling of the whole of the troposphere at an average rate of about 2°F per day. Most of the solar heat is, as we have seen, absorbed in the top layers of sea and land. For long-term balance this heat must find its way into the upper air, some of it as far as the tropopause. Once again eddy-transport is the only mechanism fast enough. Some of the heat is taken up by large eddies of the weather system type but these can take heat up only if they take it also a much greater distance sideways (polewards).

To distribute the heat correctly the assistance of smaller eddies of a rather different type is needed. Near the ground suitable eddies are produced by the flow of air over rough ground or over ocean waves. Elsewhere, or alternatively, heat is taken up by eddies—up to ten miles across— of *convection*. Convective eddies are like weather systems in that they arise from density differences, this time in the vertical. Convection occurs in a nearly incompressible fluid like sea water when, due to cooling or evaporation, the upper part becomes denser than the lower part. The fluid is then unstable (top-heavy), and eddies of convection grow automatically because the fluid is never *exactly* balanced. In air there is a complication due to its compressibility. It is the distribution of *potential* density that matters i.e., we must compare air only when compressed at the same pressure. If *potential* density increases with height* which is the case if temperature falls with height at more than a certain rate (called the adiabatic lapse-rate) the air is unstable and convection develops. As in the case of weather systems, heat is transported by growing eddies towards the region

* The *actual* density always decreases with height.

of greater potential density, that is, upwards. If now the upper troposphere cools until there is a slight increase of potential density upwards throughout the troposphere a state of balance can be set up between eddy transport and radiative heating and cooling. Actually we observe *on the average* a decrease in potential density upwards. The weather system eddies can grow in these conditions, when convective eddies cannot, and they are continually taking heat up to reduce the effectiveness of radiative cooling. Nevertheless, conditions are not uniform so that in the equatorward-moving parts of weather-system eddies, where the air aloft is cold but the air below is heated by contact with the ground, they are usually convectively unstable. Convection through deep layers produces a characteristic type of weather (cumulus clouds, local showers or thunderstorms, with clear bright intervals).

Convection can and does occur in the oceans, for the reasons we have mentioned, at certain places and times.* In the top layers it is particularly important in late autumn and winter when the air is usually colder than the sea and cools the surface. This convection extracts heat from considerable depths so that the sea surface temperature is only slightly reduced. It is because eddy motions in the sea spread heating and cooling over great masses of water that the annual variation of surface temperature over the sea is much less than over land. The temperature of the air in contact with it is also stabilized so that annual temperature variation near coasts, especially those with prevailing onshore winds, is less than in the middle of continents.

We have noticed some similarities in mode of origin and in behaviour between convective eddies and weather-systems. We may in fact regard the weather-system eddies as convective-type eddies greatly modified by the rotation of the Earth. Convective eddies feed on energy of position which is reduced when (potentially) heavy fluid moves down in the growing eddy and light fluid moves up to take its place. Weather-system eddies do the same but the exchange takes place along *sloping* surfaces. The most effective exchange direction is at an angle to the horizontal of just one half

* Not all the movements in the seas are wind-driven or of eddy type. Increase of density at the top can sometimes result in more or less steady currents. The flow through the Straits of Gibraltar, light water in at the top, heavy saline water of smaller amount out at the bottom, is due simply to evaporation exceeding rainfall and run-off from rivers in the Mediterranean. Drifts due to similar causes also exist in the oceans.

that of the surfaces of constant (potential) density existing prior to eddy development. In the atmosphere and oceans the slopes involved are very small, corresponding to displacements of one unit or less in the vertical for one hundred units horizontally. The slope is usually upward in the poleward direction. Eddies of this kind can and do grow even when, as is normally the case, potential density decreases upwards, because this is consistent with an upward *increase* along the gently sloping exchange-surface. The mechanism by which eddy-patterns sort themselves out is of some interest. It is rather like natural selection in the theory of evolution. Patterns corresponding to most efficient transformation

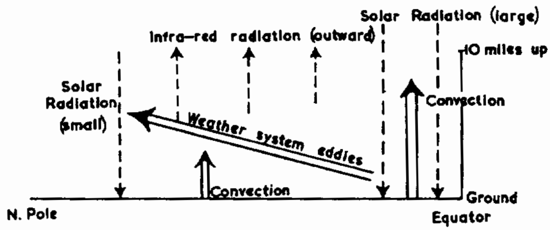

FIG. 5. Transfer of heat by radiation and eddy-transport in northern hemisphere. Eddy-transport by convection and weather systems shown by broad arrows. Solar and terrestrial radiation shown by dashed lines with arrow. Meridional circulations (especially Hadley cells) also transfer some heat.

of energy of position into energy of motion grow fastest and biggest and these dominant eddies are the ones we observe. The transfer of heat is the same for the two classes of eddy, i.e. in the exchange-direction towards the (potentially) cold fluid. This heat transfer is in fact only another way of describing the reduction in energy of position. The eddies can grow only *because* they transfer heat. Another way of looking at the process is to regard it as an irregularly functioning and inefficient natural *engine*. For the weather-system eddies the warm 'source' of heat is at low levels in low latitudes, the cold 'sink' at high levels in high latitudes. The 'sources' and 'sinks' are not *given* however—it is a case of *solvitur ambulando.*

Besides their primary property of transporting heat the weather-system eddies have a secondary (in a sense accidental) property of transporting angular momentum. The cause of this is rather subtle, though the fact is evident enough from the meteorologists' charts. In some latitudes southerly components of wind are more often associated with westerly components than they are with easterly components: there is a distinct preference for either south-westerly or north-easterly winds as compared with other directions. This means that air with more than average angular momentum is moving North, air with less than average angular momentum is moving South. The effect is that angular momentum is transported towards the North. To the North the zonal winds are accelerated, to the South they are retarded. In other latitudes, where the preference is for south-easterly or north-westerly winds, there is transfer of angular momentum towards the South. Now we *observe* that the weather-system eddies are not symmetrical about North-South lines. The flow patterns are swept back at the edges towards the West so that the high and low pressure areas look somewhat like fat boomerangs. Application of the geostrophic wind formula then shows that these eddies must be transporting angular momentum from the edges towards the middle i.e. towards the latitude at which eddies are most intense. Since eddies develop most easily in middle latitudes this means that the zonal winds are accelerated there. The acceleration continues until the surface winds become westerly and of sufficient strength for frictional forces to balance the acceleration. At the same time easterly winds are generated at both edges—the trade winds and the polar easterlies. A rather complicated theoretical analysis shows that large eddies *ought* to behave in this manner.

The transport of angular momentum by the weather-system eddies is concentrated where the winds are strongest, near the top of the troposphere. Now when we were discussing the Antarctic circumpolar current in the oceans we discovered what must happen when a zonal stress is applied at a level where it cannot be destroyed by friction. A meridional circulation is set up which effectively transfers the action of the stress to the bottom layer. Now transport of angular momentum *from the sides* at the top of the troposphere has exactly the same effect as a zonal stress acting at the same level. Consequently meridional circulations are set up in the troposphere. Below the region where the angular momentum at the top is being increased we have the Ferrel cell. Below the region in low latitude where it is being decreased we have the Hadley cell. Because the effective stresses are opposite the meridional circulations are in opposite senses. There should also be a very feeble meridional circulation near the poles in the same sense as the Hadley cell of the same hemisphere. There are some indications that the convective eddies also transport angular momentum, though they can do so effectively only in the vertical because of their small *horizontal* size compared with weather systems. They probably transport angular momentum downwards. Because convection is more intense in low latitudes the effect should be most noticeable there. If this is so, then for equilibrium the Hadley cell must rotate faster, the Ferrel cell a little slower.

Though not much energy is expended in generating the surface zonal currents they are of vital importance in climatology. As the air is forced to flow over large mountain ranges and plateaux, over surfaces of different roughness, over surfaces heated at different rates, it sets up flow-patterns on a scale slightly larger than that of the travelling weather-system eddies. These are nearly stationary for long periods.* They govern the place of development and movement of the weather-system eddies, thereby affecting the average weather. The difference in weather between one year and another is associated with differences in large-scale flow pattern. Nevertheless there are some regularities in these flow-patterns which show up in the long period averages. Particularly striking are the upper-troposphere flow-patterns over North America and eastern Asia: they are attributable mainly to the effects of large elevated regions. It is easy to 'explain' the monsoon patterns in a very general way. For example heating in summer is most effective over land in low latitudes. If we apply the thermal wind relation and also use the fact that the amount of air sucked in at the bottom as the heated air rises is roughly equal to that forced out at the top, we obtain something like the observed solution—a depression at the ground and an anticyclone at some higher level. But to obtain the true flow pattern and also to understand why the monsoon is different in different years (the practially important problem) we need to ask what keeps the flow pattern nearly stationary and enables it to fit in with patterns due to other causes.

THE STRATOSPHERE

We must deal very briefly with conditions in the stratosphere. This does not matter too much because as far up as the base of the ionosphere the *principles* involved are exactly the same. Some new

* The setting-up of these patterns constitutes part of the friction which brakes the zonal winds. (Compare the remarks on bottom-friction below the Antarctic current.)

features appear, however. Of considerable interest is the presence of ozone (especially at a height of 20-25 miles) in small but very important quantities. (cf. Chapter 7). The amount above a given place varies from day to day, the variation being closely related to the tropospheric weather systems. The upward extension of large weather-system eddies into the stratosphere, predicted on theoretical grounds, accounts broadly for these changes, though more detailed calculations are desirable. The most striking feature is, however, the distribution with latitude. Although more ozone is formed in low latitudes, because sunshine is more intense, the greatest concentration is in high latitudes. This effect could not be produced simply by 'mixing'. Nor is ozone the only 'marker' substance affected in this way. The 'rain' of very small radioactive dust particles, originally thrown up into the stratosphere by hydrogen bomb explosions in rather low latitudes, is concentrated apparently in rather high latitudes. Resisting the temptation to regard this simply as divine retribution we may be inclined to suspect that this and the ozone distribution result from a peculiar feature of the general circulation in the stratosphere. As in earlier attempts to explain the trade winds, meridional circulations have been invoked, but, warned by our experience with the zonal wind problem, we may be chary of accepting this as the whole story—or even the main part of the story. It is possible that the true explanation is quite complicated, involving, perhaps, not only eddy-transport of angular momentum with associated meridional circulations but also yet another accidental eddy transport phenomenon associated with the increase in relative concentration with height of both ozone and radioactive dust. The subtleties of eddy-transport phenomena are not yet fully understood.

If, from this incomplete survey, the reader has gained the impression that general circulation problems are complicated, this is as it should be. The point is that mere complication does not prevent their being solved. Much of the complication shows itself when we attempt to give precise answers instead of vague ones. Precision is important, not only because for practical reasons we need numerical answers, but because it is the surest way of distinguishing the true from the seemingly plausible. To answer problems in any branch of geophysics we need vast quantities of observations but we also need precise, consistent, mathematical theory to make proper use of them.

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