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CHAPTER 1

INTRODUCTION

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INTRODUCTION TO THE HANDBOOK OF WEATHER FORECASTING

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CHAPTER 1

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1.1 THE FORECASTING PROBLEM

Weather forecasting provides an important service to the community, enabling many enterprises to be carried out more effectively, more efficiently, or with a greater degree of safety than would otherwise be possible. The basic forecasting problem may be stated very simply: it is to predict the state or behaviour of the atmosphere at a given location at a specified time in the future. This bald statement, however, gives no indication of the enormous complexity of the physical processes which have to be considered, or of the vast organization needed to accomplish the task.

The forecasting process can be divided into two main stages:

(a) The assessment of the actual state of the atmosphere over the appropriate area at a time as near to the present as possible, and

(b) the estimation of how that state is likely to change during the forecast period.

For most forecasting tasks there is a third stage, the formulation of a statement which gives the user the information he requires in a form which he can readily understand.

The first stage requires the speedy assembly of surface and upper-air data from a wide area, and entails a vast and complex observing and communications organization. Before the second stage can be attempted these data must be reduced to a manageable and readily assimilable form. This is the analysis stage: it may be carried out manually or by computer or by a combination of the two – the 'man-machine mix'. The analysis procedure almost invariably incorporates some form of quality control of the data.

The usual form of analysis, in extratropical regions, is the determination of the pressure field at the surface, or the geopotential field of a given isobaric surface aloft. The patterns drawn in this step show a variety of features on a wide range of scales, most of which maintain a degree of continuity from one observation time to another. At one end of the scale large areas of relatively low or high pressure, the major depressions and anticyclones of temperate latitudes, largely determine the general weather type over a wide area and generally persist for several days. Smaller features may exist for a few days, giving rise to the day-to-day variations in the weather of a given place, while the smallest features identifiable on a large-scale chart of a limited area, such as the British Isles, may have lives of only a few hours.

Examination of a series of charts reveals certain patterns of behaviour of the systems and, in particular, that the movement and development of a given feature usually depend a good deal on the general pressure pattern on a larger scale and on the flow at higher levels in the atmosphere. It is this property of the atmospheric pressure systems, and their association with weather types, which leads to the feasibility of the second stage, the prediction of the atmospheric state and the weather at some time in the future. In the early days, when only surface observations were available, forecasting on a given occasion was based mainly on a knowledge of what had happened on similar occasions in the past. As time went on, empirical rules were derived, spurred along by the development of the polar-front model of depressions by Bjerknes and Solberg.[1] The introduction of regular upper-air soundings resulted in a greater understanding of the development and movement of pressure and weather systems, based on physical and mathematical reasoning, and this has led in turn to the objective forecasting of atmospheric pressure fields and some weather elements by the use of electronic computers.

Computer-produced analyses and prognoses now relieve the forecaster of much of the tedious work of upper-air analysis and forecasting, although subjective checking and intervention are still necessary in the quality control, the analysis and, for some purposes, the forecasting processes. Surface analyses and forecasts are still largely carried out by hand, but with valuable guidance being provided by the computer products. Detailed forecasts of the local weather in a given area or at a given place also depend to a great extent on subjective methods, often based on more detailed analyses than at present provided by computer models.

A demand still exists, therefore, for the sort of skills that a trained meteorologist has acquired, and, though the requirements of the job may change, there is little doubt that the human forecaster will be needed just as much in the foreseeable future as today. The aim of this handbook is to provide material which will assist the forecaster to understand and carry out the processes mentioned above. The first eight chapters deal mainly with the physics and dynamics of atmospheric circulation systems and with the weather associated with those systems, information which is required for the analysis and forecasting stages described in the following few chapters. The forecasting of individual weather elements is then discussed, with emphasis on the physical processes at work, and a final chapter mentions ways in which the forecaster can help himself to understand his local weather and to improve his forecasting skill.

The rest of this chapter gives a brief account of the general circulation of the atmosphere and of the part played in this circulation by the westerlies of temperate latitudes.

1.2 THE GENERAL CIRCULATION

The primary cause of atmospheric motion is differential heating. On the global scale it is the difference between the atmospheric heating at the equator and at the poles that is important, the broad-scale characteristics of the resultant flow being largely determined by the rotation of the earth. Smaller-scale temperature contrasts occur between, for example, land and adjacent ocean, or between two different types of land surface, and these contrasts influence or result in the formation of circulation systems ranging in size from the major depressions and anticyclones of temperate latitudes, through mesoscale features such as sea-breezes, down to the microscale of individual thermals and eddies. The formation and behaviour of many features are markedly influenced by the general flow of the larger systems. The rotation of the earth plays a part which increases with increasing distance from the equator, and which diminishes as the scale of the phenomenon decreases; the effect is negligible near the equator and, in temperate regions, for systems which have lives of an hour or two. The topography of the earth also has an important part to play, in addition to the thermodynamic influences of different surfaces, in providing a braking effect which varies with the type of surface, and in providing obstacles to the movement of the lower atmosphere where there are large mountain ranges or plateaux.

It can be seen, even from the very brief outlines given above, that the atmospheric circulation is the result of a complex set of influences. It is imperfectly understood, and is likely to yield only slowly to theoretical and observational studies, although numerical models can simulate many of the major features in a fairly realistic way. General physical principles may be used to illustrate the main features of the circulation, but many difficulties remain.

1.2.1 The heat balance of the atmosphere

The prime source of atmospheric heating is the sun. The rate of receipt of solar energy on a horizontal 'surface' of unit area at the upper limits of the atmosphere depends upon the angular distance of the sun from the zenith, which in turn depends upon latitude, season and time of day. (There is also an annual variation of a few per cent as a result of the varying distance of the earth from the sun, but this will be neglected in the following discussion.) On its way through the atmosphere, the radiation undergoes absorption, scattering and reflection. The main absorption of the solar radiation occurs in the ultra-violet and infra-red regions of the spectrum, being responsible for the high temperatures of the thermosphere and around the stratopause (see section 8.3 on page 11 of Chapter 8). Scattering occurs when the radiation encounters molecules, groups of molecules or aerosol: some of the scattered radiation is returned to space, while the rest, perhaps after multiple scattering, reaches the surface of the earth as diffuse solar radiation. Clouds reflect a major portion of the radiation falling upon them, i.e. they have a high albedo, while the earth's surface reflects a fraction which varies from about 3 per cent for a calm water surface at high solar elevations to 85 per cent for fresh snow. What is left, on average rather less than 50 per cent of the initial radiation, is absorbed by the surface. The surface re-radiates some of the energy as long-wave radiation, and in turn receives long-wave radiation from the atmosphere (see Chapter 17 – Temperature). The total energy received by the surface, as short-wave and long-wave radiation, minus the energy lost by the surface as long-wave radiation, is called the 'net radiation'.

Although the net radiation varies widely with latitude, time of year, time of day and the nature and topography of the surface, it is useful in general circulation studies to look at the latitudinal mean values and their annual variation. Figure 1 shows how the daily net radiation varies with latitude and time of year: it is based on data by Rasool and Prabhakara,[2] re-analysed by Palmén and Newton.[3] The data were derived from observations of albedo and terrestrial radiation by the TIROS VII satellite. The main features are the belt of positive values over most of the tropical region throughout the year, extending to temperate latitudes during the summer months, and the negative values over most of the temperate latitudes for the rest of the year. The pattern immediately suggests that there must be some means of transporting heat from the tropics to temperate and polar regions in order to maintain the observed temperature distribution.

Estimates of the heat fluxes, based on measurements of atmospheric temperatures and winds, and of similar quantities in the oceans, are shown in Figure 2. They are based on data presented by Sellers,[4] and republished in Palmén and Newton.[3] It can be seen that the major part of the energy transport is accomplished by the atmosphere, mainly as 'realized energy' (sensible heat, $\int c_p \, dT$, plus potential energy) together with latent heat. Transport of latent heat occurs when evaporation in one region is followed by condensation and precipitation, with release of the latent energy, elsewhere. The oceanic heat transport, though smaller than that in the atmosphere, provides an important contribution, particularly in tropical and lower temperate latitudes: the ocean acts as a large heat store, energy being supplied by the absorption of solar radiation, or from the air when it is warmer than the sea surface, and being given up to the atmosphere if the water is warmer than the air. The ocean currents themselves are largely wind driven, so that even the oceanic heat transport is mainly determined by atmospheric motions.

Differential heating in a fluid leads to temperature, and hence density, differences between one part of the fluid and another. The inhomogeneities set up motions which tend to destroy the resultant temperature differences. In the atmosphere the heating in tropical regions and cooling at high latitudes lead to the motions which bring about the energy fluxes shown in Figure 2: the heat transport cannot take place instantaneously, so the temperature difference between tropics and poles is never reduced to zero. The precise nature of the circulation set up depends upon a number of factors, in particular the difference between the net radiation values at the pole and tropics, the rotation of the earth, the distribution of land and sea, and the topography of the land. Taking a broad overall view, the general circulation can be divided into two main zones: the tropical circulation and the meandering westerly flow of temperate latitudes.

1.2.2 The tropical circulation

In a fluid, subject to no external forces other than gravity, differential heating at the lower boundary would lead to a 'direct' circulation, with warm fluid rising above the heated area and cold fluid descending elsewhere. To complete the circulation there would be horizontal branches, with flow from warm to cold aloft and from the cold area to the warmed area at the lower level. On a uniform non-rotating earth, heated at the equator and cooled at the poles, this type of circulation would be set up, with warm air rising in tropical latitudes, poleward motion aloft, descent in polar regions and an equatorward flow near the surface. This is the simple cell first proposed by Hadley.[5] However, the earth is rotating. The component of rotation about the local vertical is the important factor, and this increases with increasing latitude, effectively limiting the meridional cell to the tropics while other forms of atmospheric motion predominate at higher latitudes (see 1.2.3, page7). The poleward boundary of the tropical cell is marked by a belt of strong winds near the tropopause, the subtropical jet (STJ), a result of the marked contrast of temperature set up in the upper troposphere between the poleward-drifting tropical air and the air of temperate latitudes. It is of interest to consider the angular momentum of the air as it moves polewards from the equator. The air will conserve its absolute angular momentum, i.e. relative to axes fixed in space. As it moves towards the subtropics it moves nearer to the earth's axis and the eastward component of motion must increase as it does so. Figure 3 shows the calculated zonal component, assuming conservation of absolute angular momentum, of an element of air, initially stationary relative to the earth's surface, as it moves from the equator to 30° latitude. Observed speeds in the STJ are about half the calculated values, the difference being the result of loss of angular momentum by eddy* transfer. The effect of the eddies becomes greater with increasing distance from the equator and as the meridional cell becomes weaker.

* 'eddies' in this context being used to indicate disturbances on scales ranging up to that of synoptic systems.

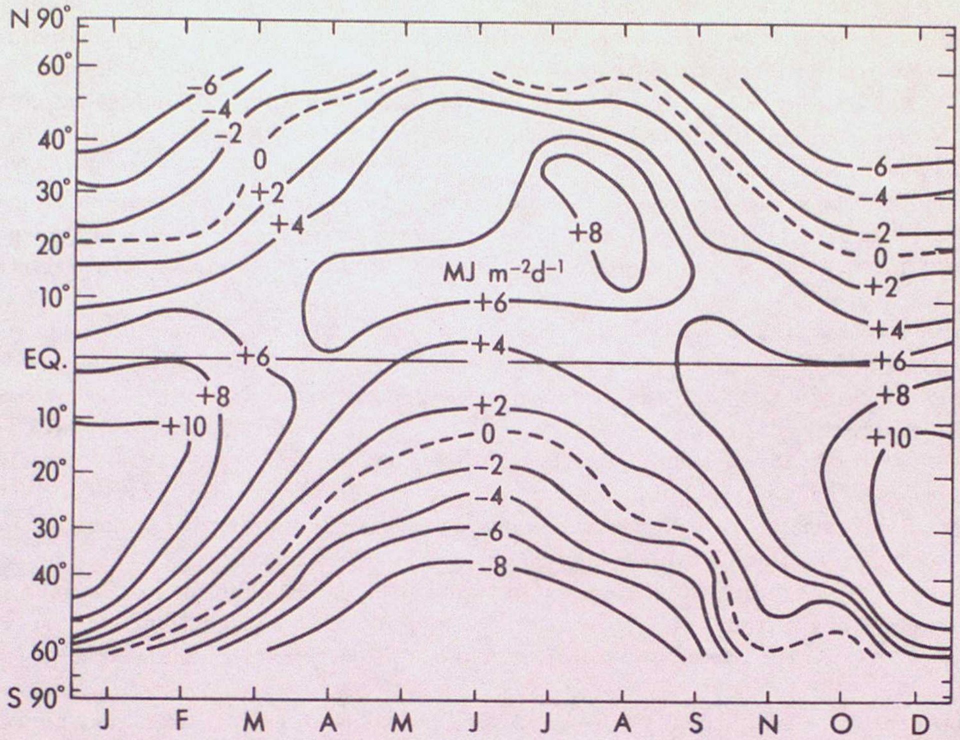

FIGURE 1. Isopleths of net daily radiation at the surface in megajoules per square metre, determined by Rasool and Prabhakara[2] from a year of TIROS satellite measurements

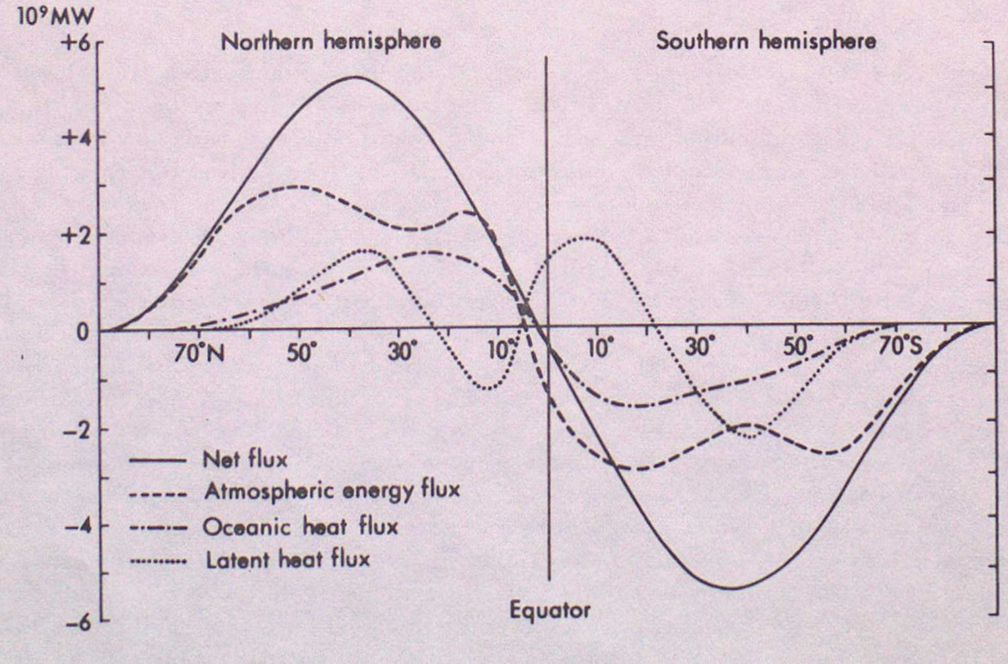

FIGURE 2. Mean annual meridional flux of energy

Solid curves show, in megawatts, the mean annual northward flux of net energy in the atmosphere/ocean system. Other curves show atmospheric flux of realized energy (enthalpy plus potential energy), atmospheric flux of latent heat and oceanic flux of heat.

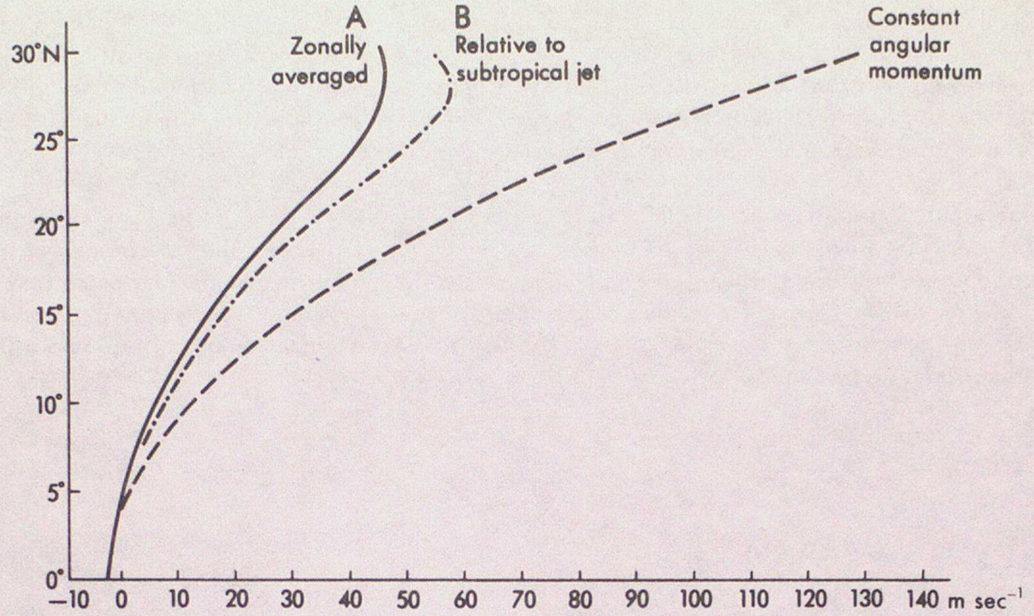

FIGURE 3. Profiles of the 200-mb mean zonal wind between the equator and 30°N in winter, compared with the profile corresponding to constant angular momentum.

Curve A represents the zonal wind averaged around latitude circles, while in the case of profile B the winds were averaged relative to the core of strongest wind in the subtropical jet stream which meanders north and south about a winter mean latitude of 27° 30´N. (From Palmen[6]).

The picture of a simple Hadley circulation has then to be modified, and the modifications have an important bearing on the relationships between the tropical atmosphere and that of temperate and polar latitudes. In order to see this, we must start with the equatorward-moving air near the surface, the north-east and south-east trade winds. The easterly component of the air is reduced by friction at the surface; by this process the air acquires some of the earth's westerly absolute angular momentum, and carries it to the high troposphere in the ascending branch of the circulation. As the air then moves polewards and the westerly relative angular momentum increases, eddies begin to take over the major portion of the transport of the angular momentum. The horizontal eddies bring about the transport of angular momentum polewards across the latitude of the subtropical jet stream into temperate regions. Here, in the westerly winds near the surface, the atmosphere must lose to the earth by friction an amount of westerly angular momentum equal to the amount gained by the atmosphere in tropical regions, if the earth's rate of rotation is to remain constant.

In the poleward parts of the meridional cell the air sinks slowly, losing momentum by eddy transfer as it does so. On reaching the lower troposphere, in the latitudes of the subtropical anticyclones, some of the air moves polewards and takes part in the formation of the mid-latitude depressions and anticyclones, while some moves equatorwards as a branch of the trade-wind system.

The average speeds of the horizontal meridional motions are about 2-3 metres per second in the winter, when the cell is at its strongest. The average vertical velocities are probably somewhat less than 1 centimetre per second, and while this is reasonable for the subsiding air over the subtropical highs there is no gentle overall rising of air near the equator since the atmosphere is, on the whole, stable. Instead, the upward motion appears to occur in bursts as a result of intense convection in large thunderstorms or cloud clusters,[7] covering only a small fraction of the equatorial belt at any one time.

The transport of momentum in the poleward halves of the tropical cells and across the subtropics at high levels is accomplished largely by disturbances or eddies. Poleward transport of angular momentum across the subtropics in this way implies also the transport of kinetic energy: calculations by Kao[8] and Pisharoty[9] suggest that about $0.20 - 0.24 \times 10^9$ megawatts is transported by eddies across the 30° latitude circle in winter. This amounts to a small, but not negligible, fraction of the kinetic energy required to maintain the mean westerlies of extratropical latitudes.

Climatologically, the interaction between the tropics and mid latitudes can be seen in the mean positions of the ridges and troughs in the subtropical jet stream (STJ) and the jet streams of temperate latitudes (the polar-front jet streams (PFJ)). Ridges in the STJ occur in roughly the same longitude as troughs in the mean PFJ; these are favoured locations for the injection of kinetic energy into the middle latitudes. Figure 4 shows the mean positions of the STJ and the boundaries of the region within which the PFJ is normally to be found.

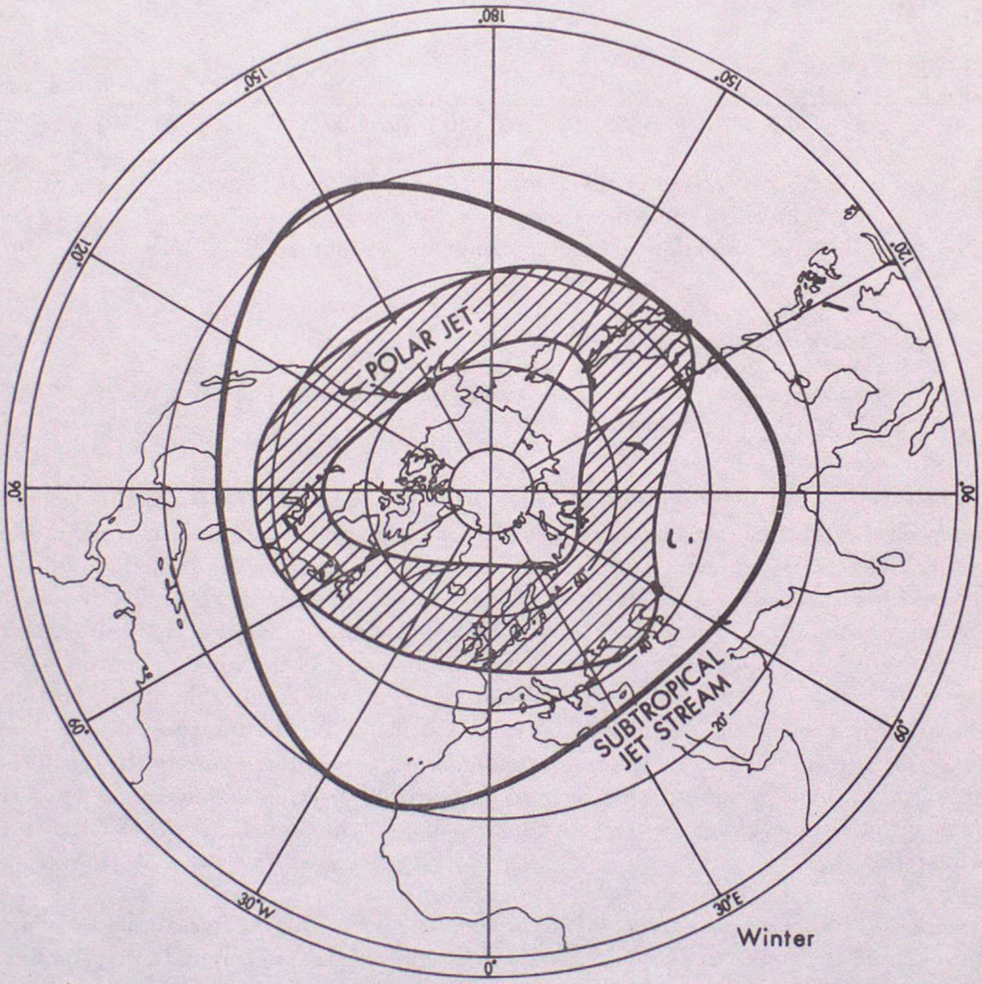

FIGURE 4. Mean axis of subtropical jet stream during winter, and area (shaded) of principal activity of polar-front jet stream.

The tropical circulation cell and eddies also bring about a meridional flux of energy in the form of sensible heat and latent heat. The flux of 'realized energy' (sensible heat plus potential energy) is, on average, poleward at all latitudes. There is little contribution to the energy flux across subtropical latitudes at the upper levels; poleward of 20° latitude, the major portion of the heat transport is accomplished by eddies at relatively low levels, the influence of the eddies becoming more important with increasing distance from the equator.

The flux of latent heat, in contrast, is not poleward over much of the tropical belt. Air subsiding in the subtropical anticyclones is dry, and evaporation can proceed rapidly when it reaches the surface layers. The result is the transport of latent heat both equatorwards and polewards from the subtropics. Near the equator, in the region where the trade winds of the two hemispheres meet (the intertropical convergence zone), the latent heat is released as a result of condensation followed by precipitation in areas of intense convection and in this way plays an important part in the maintenance of the tropical meridional cell. The latent heat in the poleward-moving air is released mainly as the warm moist air ascends and condensation and precipitation occur in the synoptic systems of temperate latitudes.

The transport of energy polewards from the tropics by ocean currents is of the same order of magnitude as the transports of atmospheric energy and latent heat, but the oceanic heat and latent heat fluxes diminish steadily with increasing latitude beyond about 30-40°.

The total heat transport from tropical to extratropical regions is about 30-40 per cent of that required to offset the loss of heat from the temperate and polar atmosphere to space. The rest of the energy lost by the extratropical atmosphere by radiation is largely balanced by heat exchange (both sensible and latent) with the oceans, which absorb a high proportion of the solar radiation falling on them and release the energy gradually when the overlying atmosphere is colder than the sea surface.

To summarize, the tropical circulation performs a complex and important part in the transport of heat and momentum into extratropical regions. Friction at the earth's surface reduces the easterly component of the trade winds, which means that the air acquires westerly absolute angular momentum from the earth. As the air, having ascended in regions of intense convection, moves poleward in the upper branch of the cell, conservation of absolute angular momentum results in increasing westerly relative angular momentum with latitude and the creation, at the poleward boundary of the cell, of the subtropical jet stream. Some of the kinetic energy (KE) of the STJ is dissipated within the tropical cell, but the rest is transported polewards, mainly by large-scale eddies to provide the mid-latitude westerlies with a significant fraction of the KE needed to maintain them. By a rather different mechanism, the transport of heat by eddies at low levels in the atmosphere and by wind-driven ocean currents, tropical regions supply temperate areas with about 30-40 per cent of the energy needed to offset cooling of the atmosphere by radiation. The evidence suggests that events in the tropical atmosphere are likely to have an important bearing on that of temperate and possibly polar regions, and the converse is probably also true.

It has been possible in this short section to give only a much simplified account of the circulation of the tropics. The circulations are far more complex than indicated above: they include such features as the summer monsoon of southern Asia, the interhemispheric transport of angular momentum from the winter to the summer hemisphere, and the equatorial easterly jet stream. More detailed accounts are given in References 3 and 10.

1.2.3 The mid-latitude westerlies

Air moving over a horizontal surface which is rotating about a vertical axis will appear, to an observer rotating with the surface, to be deflected, because, in the absence of any forces acting on the air, its motion will remain constant relative to axes fixed in space. To the observer, the effect is the same as if the surface were stationary and the air subject to a deflecting force – this is known as the Coriolis effect. The surface of the earth rotates about the local vertical at a rate proportional to the sine of the latitude, and the apparent deflecting force increases with increasing latitude in the same way. It is zero at the equator and small over most of the tropics, but in middle and high latitudes the Coriolis effect becomes a major factor in determining the nature of atmospheric motions. On a fairly broad scale at least, the flow is largely geostrophic, that is, it is controlled mainly by the pressure gradient and Coriolis forces.

On a hemispheric scale, the generally westerly flow in the free atmosphere (above the surface friction layer) is in approximate geostrophic balance with the observed meridional pressure gradient, resulting from the temperature difference between the tropics and polar regions. The greater the temperature contrast, the stronger the circulation (as in winter), but there is a feedback mechanism in that, on the whole, the stronger circulation brings about a greater meridional heat flux which tends to reduce the temperature difference.

The flow in mid latitudes is by no means a uniform westerly current. Wave motions predominate, and it will be shown in the next chapter (section 2.2.8, page 38) that the meteorologically important wave motions can be divided into two broad classes, long and short waves. The long waves, of which there may be typically three or four round a hemisphere, behave in some respects as though the atmosphere were barotropic and non-divergent:* their properties depend upon the principle of the conservation of absolute vorticity in non-divergent flow, and they arise mainly because of thermal and orographic influences. Long waves usually move eastwards or westwards only slowly, or they may remain stationary for a time, the wind speed in any case being much greater than the speed of the wave. Although some divergence does occur in association with these waves, its magnitude is usually small.

However, the atmosphere is not everywhere nearly barotropic, that is to say there are usually variations of temperature (and hence density) along a surface of constant pressure. Where this occurs the atmosphere is said to be 'baroclinic', The main temperature contrasts are very often concentrated into relatively narrow zones, called 'baroclinic zones'. The main baroclinic zones are associated with outbreaks of deep, cold air from polar regions towards, and at times into, the tropics, and with bursts of warm, moist air polewards from tropical regions. Being regions of strong temperature contrast throughout a substantial depth of the atmosphere, baroclinic zones are generally associated with jet streams in the upper troposphere (see section 8.1 of Chapter 8). Equatorward of the main baroclinic zones the temperature contrasts are lower, and towards the poles often very much lower, than in the baroclinic zones themselves. Within the baroclinic zones, the air is usually stable for vertical motion, but in many parts air which is rising and moving polewards at the same time may reach a region of lower potential temperature. Baroclinic instability is then said to exist, and the conditions are favourable for the formation and, possibly, development of short waves. The behaviour of the short waves varies: they may move fairly quickly through the long-wave pattern in the direction of the general flow, or one may develop to become a major feature in its own right, modifying or itself becoming a member of the long-wave system.

The baroclinic waves are generally associated with the mobile depressions and anticyclones of mid latitudes. Upper-level divergence and surface convergence occur, leading to vertical motion and the formation of clouds and precipitation. The compensating downward motion, or subsidence, often results in clear skies and dry weather.

The vertical motion of the short waves transports heat upwards, as both sensible and latent heat, which compensates for the loss of heat by radiation from the upper troposphere. The waves also transport kinetic energy downwards where it is dissipated by friction at the earth's surface. The meridional flows, with cold air moving equatorwards and warm air moving polewards, compensate for the net loss of heat of much of the lower atmosphere of mid latitudes. Wind-driven ocean currents also play a part in transporting heat meridionally.

The long waves are important in the meridional transport of heat and momentum, and also in determining the locations favourable for the formation of short waves. The location of the long waves is influenced by topography - not only by the presence of mountain ranges and plateaux, but also by the temperature contrasts which often exist between land and sea. Particularly in winter, when both the tropical and temperate circulations are much stronger than in summer, the supply of heat to the atmosphere from the oceans is important in determining the flow characteristics in middle latitudes (see, for example, Chapter 15, section 15.3.2.1, page 13, on long-range forecasting). The amplitude of the long waves varies. At times the flow over a sizeable fraction of a hemisphere may be predominantly zonal, that is to say, the amplitude of the long waves is small, but at other times the amplitude is large and the flow has strong meridional components. In mainly zonal flow, temperatures are less variable and generally nearer to normal than when the flow is more meridional. The long waves largely determine the general weather type over a given area for a few days or more. A pattern, once established, may persist for a matter of weeks with relatively minor changes or interruptions.

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* See Chapter 2 for discussion of the meanings and implications of the dynamical terms used in this section.

The short waves, also, vary in amplitude and intensity. Large-amplitude waves are associated with more marked temperature and, usually, weather changes than are small-amplitude waves. The corresponding surface systems bring about weather changes at a given place on a time scale of a day or rather less, as the depressions and other synoptic-scale features progress in the long-wave pattern. A short wave which develops to become a major feature of the circulation may well cause a change of the general weather type over a particular area. At the other extreme, there are smaller-scale features which determine the very short-period changes in weather, such as the mesoscale variations of rainfall intensity within a rain belt, the variations of cloudiness within an anticyclone, and so on. These may be caused by local topography, or by inhomogeneities in the atmosphere.

Finally it should be noted that the presence of water vapour in the atmosphere is an essential factor in the circulations which have been discussed. Neither the tropical meridional cell nor the baroclinic disturbances of the mid-latitude westerlies would reach their present vigour if they did not bring about the condensation of water vapour, with the corresponding release of the latent heat of evaporation.

1.2.4 Concluding remarks

It is important that the forecaster should know, at least in broad terms, the main features of the general circulation of the atmosphere and how it transforms and transports energy in various forms. In this chapter it has been possible to give only a very brief outline of the processes at work; for further details the forecaster should consult References 3 and 10 of the Bibliography.

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