CLIMATES OF THE BRITISH ISLES

Present, Past and Future

Edited by
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EXPLAINING THE CLIMATE OF THE BRITISH ISLES

Trevor Davies, P. Mick Kelly and Tim Osborn

It's a warm wind, the west wind, full of birds' cries.
John Masefield, 'The West Wind'

INTRODUCTION

The climate of the British Isles is the cumulative result of each day s weather. The weather on a particular day depends on the character of the atmospheric circulation over the Islands and on the **synoptic system** – **depression** or **anticyclone** – affecting the region: in other words, the overall weather type' (see Chapter 8). The nature of the winds and the character of the synoptic system depend, in turn, on the interactions between the atmosphere, the oceans and the land surface at every other point on the globe.

The weather experienced at a particular location in the British Isles is a function of the precise part of the synoptic system which is overhead at the time and of local influences. The weather associated with a particular synoptic system (which has a typical horizontal scale of 1–2,000 km) can have important variations on scales of 10–100 km. The weather which is 'delivered' to the British Isles by the synoptic systems is, in turn, modified by the underlying surface: high land may enhance the precipitation process (see Chapter 3); different land surfaces will react to incoming solar radiation in different ways, affecting the near-surface air temperature; local circulations (on the scale of one to a few tens of km) can be affected by topography or by the discontinuity across a coastal zone. Local effects such as these can exert a strong influence on the climate experienced by particular places.

The processes which produce the British climate are therefore enormously complex. There are interactions on all space- and time-scales. Consequently, it is over-simplistic to attempt to define a 'beginning' and an 'end' to the various interplays of factors. The picture gets even more complex when one considers that the nature of the interactions in the land–atmosphere–ocean system varies over time. This is either because of natural variability internal to the system, or because of outside 'forcings', such as variations in the receipt of solar radiation at the top of the atmosphere, or changes in the composition of the atmosphere because of volcanic activity or **anthropogenic** gas emissions.

Having made these cautionary remarks, we will consider in this chapter the most important processes affecting the climate of the British Isles sequentially, from the global to the local scale – and thereby run the risk of over-simplification. We start by considering the manner in which global climate is shaped by the planetary-scale atmospheric circulation, or 'general circulation'. The general circulation of the atmosphere takes the form it does because of the way energy from the Sun is distributed and utilised, because the Earth rotates, and because of the particular geographical pattern and orography of the land and ocean basins. We first look at the radiation and consequent heat budgets of the surface-atmosphere system, the driving force behind the wind systems

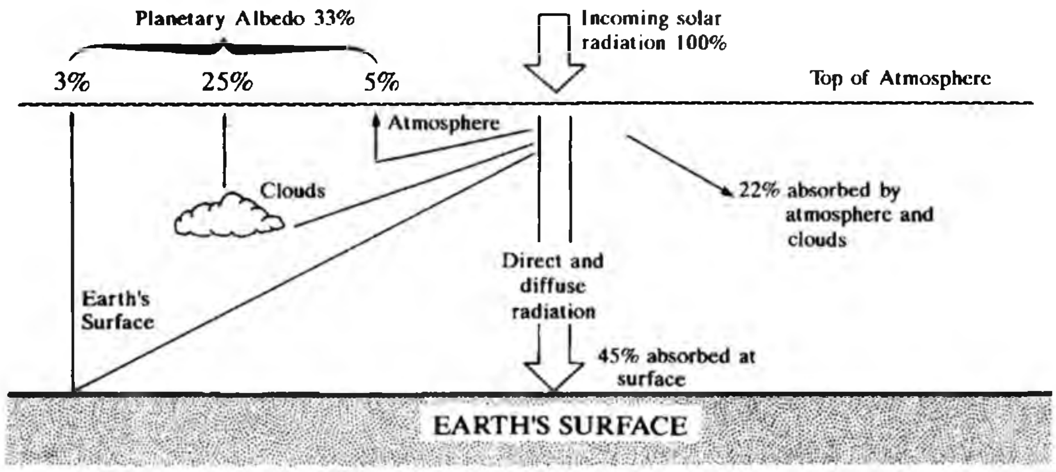


Figure 2.1 A representation of what happens to the incoming solar radiation.

of the Earth.[1] Later sections look at links between climate and the oceans and at local-scale influences on climate.

GLOBAL-SCALE RADIATION AND HEAT BALANCES

The main source of energy available to the planet is the Sun. Radiation emitted by the Sun is short in wavelength. Around 33 per cent of the solar radiation which is received by the planet from the Sun is scattered back to space (see Figure 2.1). The fraction lost in this way is called the planetary **albedo**. Most of the scattering occurs in the atmosphere, predominantly by clouds. The Earth's surface contribution to the planetary albedo is small. There are, however, considerable geographical differences in surface albedo – desert albedos are generally greater than those of forests and snow has a high albedo, as does water when the Sun angle is low. The atmosphere (gases, clouds and dust) absorbs about 22 per cent of the incoming solar radiation, leaving 45 per cent of the original received solar radiation to arrive at the Earth's surface, where it is absorbed.

The Earth's surface also receives radiation from the atmosphere, where the solar radiation has been absorbed and re-emitted as longwave radiation.[2] This longwave radiation is also absorbed at the Earth's surface. At the surface and in the atmosphere there is therefore a complex pattern of longwave radiation absorption, emission, re-absorption and re-emission.

An issue of considerable current concern is the change in concentration of the so-called 'greenhouse gases' in the atmosphere because of human activities. These greenhouse gases are the constituents of the atmosphere which play the important role of absorbing longwave terrestrial radiation (while not interfering significantly with the incoming solar radiation) thereby trapping heat near the Earth's surface. This **greenhouse effect** maintains temperatures which make the planet habitable; without it the Earth's surface would be much colder than it is. An increase in the concentrations of greenhouse gases should lead to significant global warming. This enhanced greenhouse effect is likely to change global climate, and the climate of the British Isles, in important ways. This issue is addressed at some length in Chapter 15, but we will neglect it for now

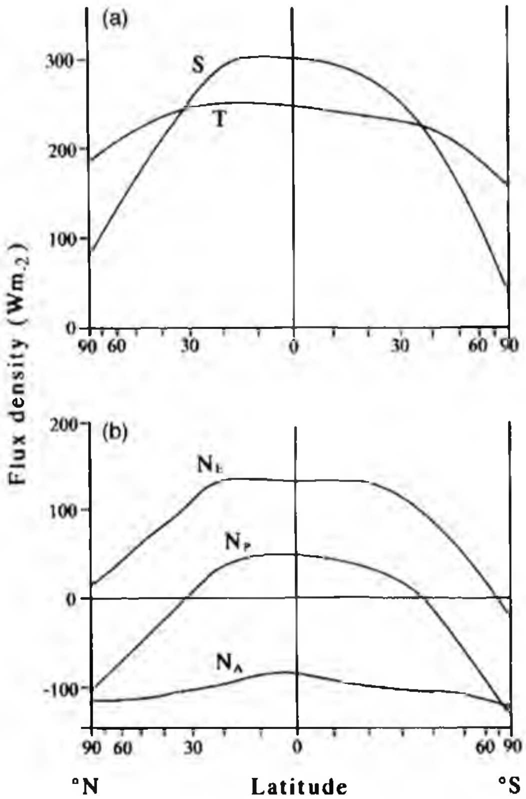


Figure 2.2 (a) Latitudinal averages of solar radiation (S) and longwave (terrestrial) radiation (T), and (b) latitudinal averages of net planetary radiation (N_P) net radiation at the Earth's surface (N_E), and net radiation to the atmosphere (N_A). Adapted from McIlveen (1992) (see p. 32 note 13).

as we attempt to explain the atmospheric processes which give the Islands' 'normal' climate its essential character.

Although the net radiation (the difference between the incoming radiation and the outgoing radiation) for the planet as a whole is zero, this is not the case at all latitudes; neither is it the case for the atmosphere and the Earth's surface separately. For now, we shall ignore geographical variations, and only consider the radiation budget for the planet averaged over all lines of longitude from 90°N to 90°S. This gives us a globally averaged latitudinal distribution. Since the Sun's overhead position traverses between the Tropics of Capricorn (23.45°S) and Cancer (23.45°N), the input of solar radiation is at a maximum in equatorial latitudes. The low Sun angle towards the Poles, and the greater albedo of snow and ice, means that the input of solar radiation at the surface is much smaller at these high latitudes (Figure 2.2a). The emission of longwave radiation shows a smaller latitudinal variation and exceeds the solar energy input outside tropical latitudes. The resulting planetary net radiation distribution (Figure 2.2b) is positive between 30°N and 40°S and negative at higher latitudes. The general circulation of the atmosphere and the oceans is driven by this latitudinal energy imbalance and it serves to transport heat polewards.

When we consider the net radiation distribution for the atmosphere (Figure 2.2b), we see that the values are negative at all latitudes. In terms of radiant energy, the atmosphere is cooling at a rate equivalent to about 0.8°C per day. This radiational cooling is offset by the transfer of energy from the Earth's surface which experiences positive net radiation at practically all latitudes (Figure 2.2b). This transfer is effected through the **convection of sensible heat** and the release of **latent heat** through the condensation of water vapour which was evaporated from the surface. Globally, latent heat transfer from the surface to the atmosphere is around four times more important than the convection of sensible heat, although there are large latitudinal variations. For example, near the Equator sensible heat accounts for only around 5 per cent of the total vertical heat transfer from surface to atmosphere, and yet at 70°N it accounts for about one-half of the total transfer.

In terms of polewards **heat advection** in the atmosphere, however, sensible heat transport is more important than that of latent heat. This indicates that most water vapour re-condenses in the atmosphere in much the same latitudinal zone as it was evaporated from the Earth's surface. More than half of the atmosphere's sensible heat originates in the global atmospheric 'engine': the tropical rain belt (between around 0–10°N). Polewards sensible heat

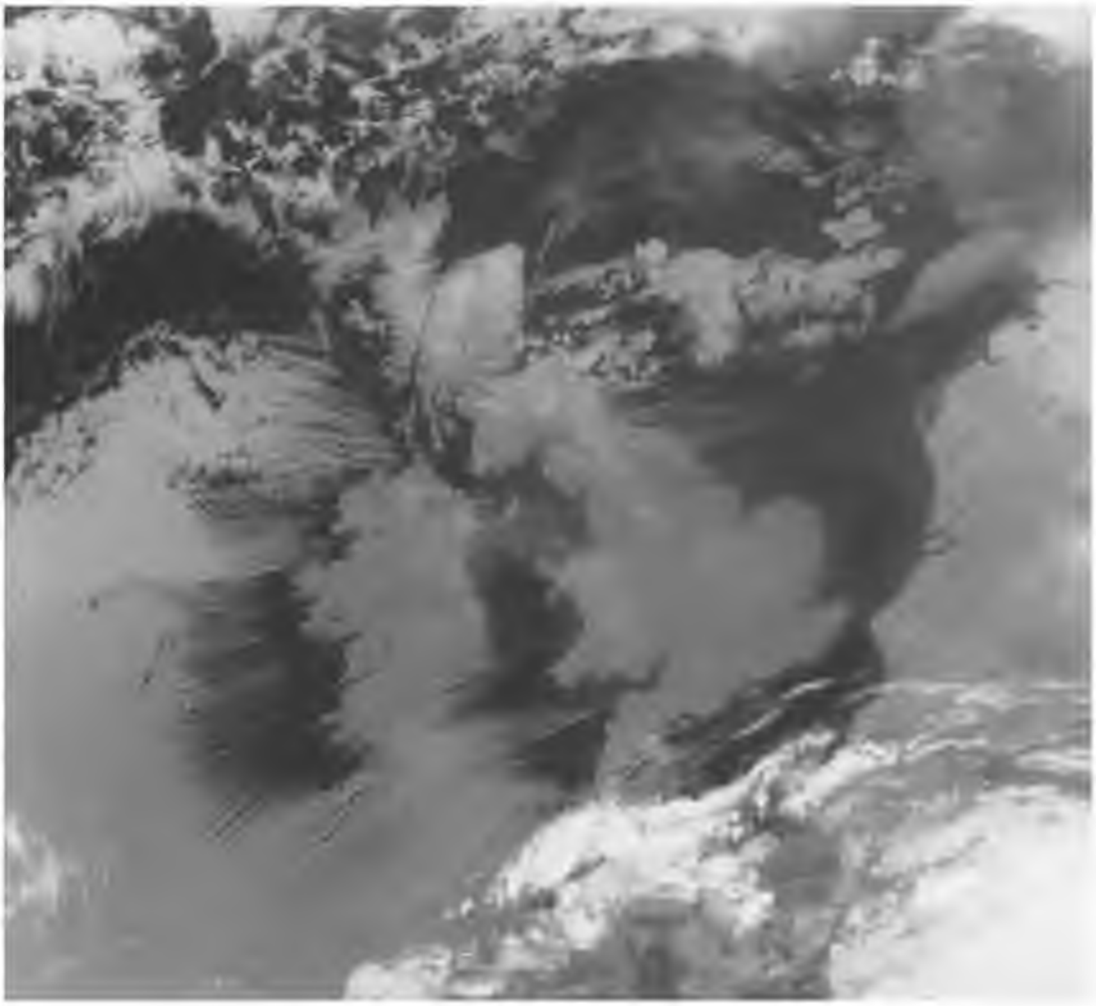


Figure 2.3 Two examples of infra-red satellite images of the British Isles showing the contrast between (above) winter, 27 February 1986, when the cold land shows light against a warm sea, and (p.15) summer, 16 May 1980, when the warm land shows dark against the colder sea. The land surface cools and warms much more rapidly than the ocean surface because of its lower heat capacity. Images courtesy of the University of Dundee.

transfer exhibits a double maximum in each hemisphere; polewards of the tropical rain belt at around 20°N and S, and again at around 50–60°N and S in response to condensation of water vapour in the mid-latitude cyclonic storm belts. Ocean currents account for around one-third of the polewards sensible heat transport.

Thus far, we have considered annual energy distributions averaged over latitude. There are important seasonal and geographical variations. We shall come back to these later but, at this point, some important features will be introduced. The specific heat capacity of land surfaces is much less than that of the oceans (Figure 2.3), so the adjustment time of the oceans to

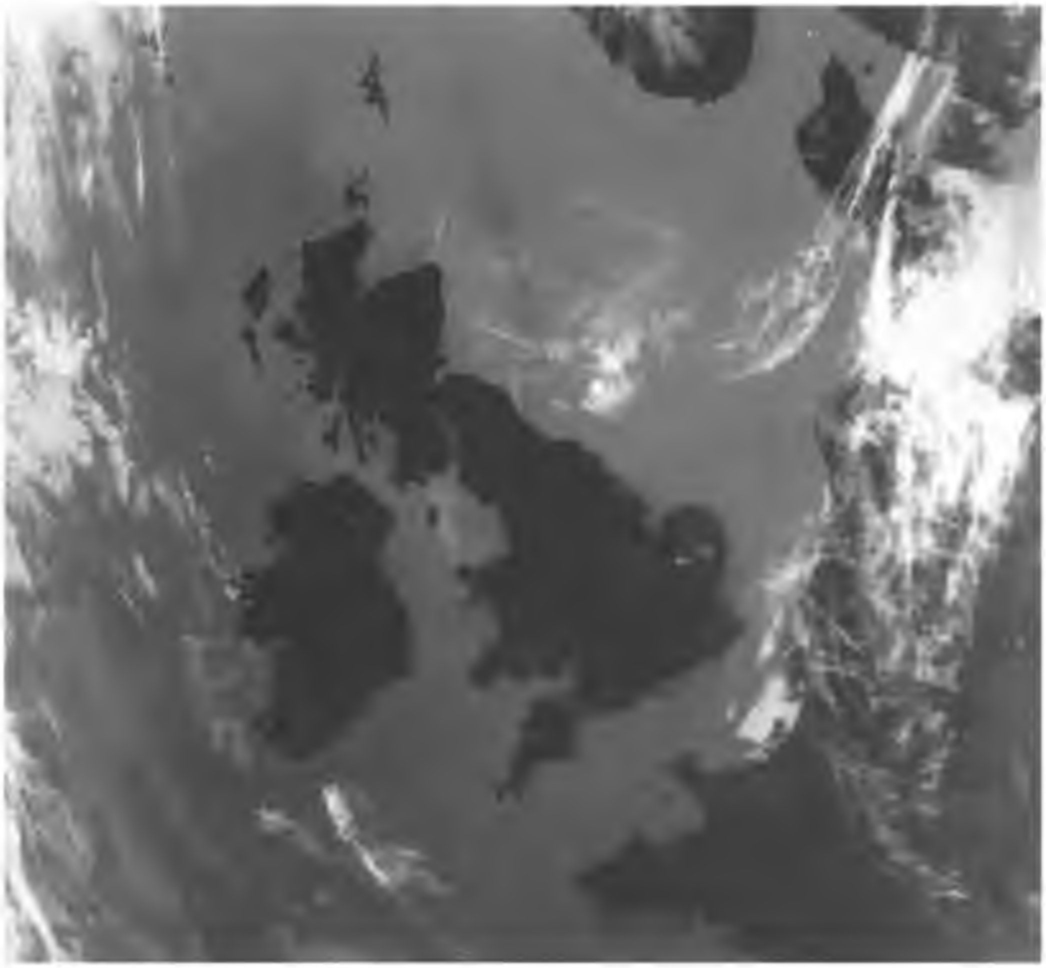


Figure 2.3 Continued

any change in energy input is greater than of the land. Moreover, unlike the land, the oceans can transport heat, so **sea-surface temperatures** are not related in a simple way to the energy balance. Figure 2.4, which illustrates the annual range in temperature at the Earth's surface, clearly reflects the large-scale land/sea distribution. The annual temperature range in the interior of the high northern continents approaches, or exceeds, 50°C. The sea-surface temperature range is much smaller than the range over land and reflects the ocean circulations. The moderating influence of the North Atlantic Drift stretching over to the British Isles, for example, is most pronounced. The greater adjustment time of the oceans means that maximum and minimum sea-surface temperatures lag the solstices by around six weeks. Bathing in the sea around the British Isles is most pleasurable late in the summer.

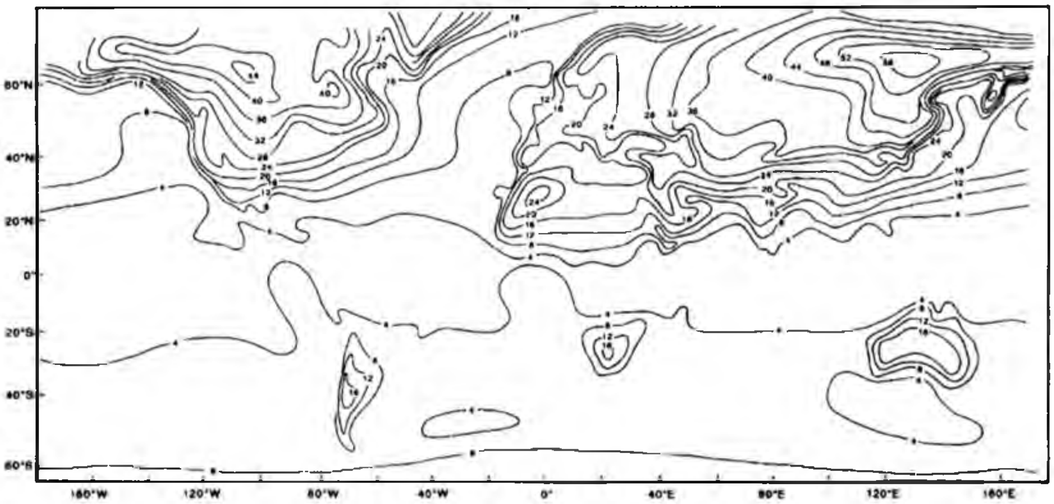


Figure 2.4 Average annual temperature range at the Earth's surface (°C). The annual temperature range is the difference in mean surface air temperature between the warmest and coldest months. Adapted from Wallace and Hobbs (1977).

THE GENERAL CIRCULATION

Early conceptualisations of the general circulation of the atmosphere took the energy imbalance between equator and pole as a starting point. They then focused on variants of a single, large-scale thermally driven circulation cell in both hemispheres – with a rising limb in tropical latitudes, descent in higher latitudes, and a return flow near the surface – as a means of redressing the imbalance. Such a large convective cell, with one dimension much more extensive than the other, is not stable. Improved observations, coupled with better understanding of the workings of a very thin atmospheric skin on a rotating Earth, produced a latitudinal conceptualisation which looks something like the scheme shown on the right-hand side of the hemisphere in Figure 2.5.

In tropical latitudes there is the so-called Hadley Cell, named after the eighteenth-century English scientist, George Hadley. The Hadley Cell can be viewed as a thermally direct cell, although much of the upwards energy transport near the Equator is concentrated over a relatively small area in vertically extensive convective cloud clusters. A zone of rapidly moving air, the subtropical jet stream, travelling west-to-east, is located near the **tropopause** at around 30°. This results from a need to conserve **angular momentum** as air moves from the Equator to regions closer to the Earth's rotational axis. Just beneath the tropopause in mid-latitudes is another zone of air moving rapidly from west-to-east – the polar front jet stream. This mid-latitude jet stream is a consequence of the strong thermal gradient in the vicinity of the **polar front**, a transitional zone between the relatively cold tropospheric **air masses** of high latitudes and the relatively warm air masses of subtropical latitudes. The latitudes where the polar front occurs represent an area of **slantwise convection**, interleaves of descending cold air moving equatorwards and ascending warm air moving polewards. These are the latitudes of travelling cyclones and anticyclones, synoptic systems which play a crucial part in the maintenance of the atmospheric general circulation, and in shaping the climate of the British Isles.

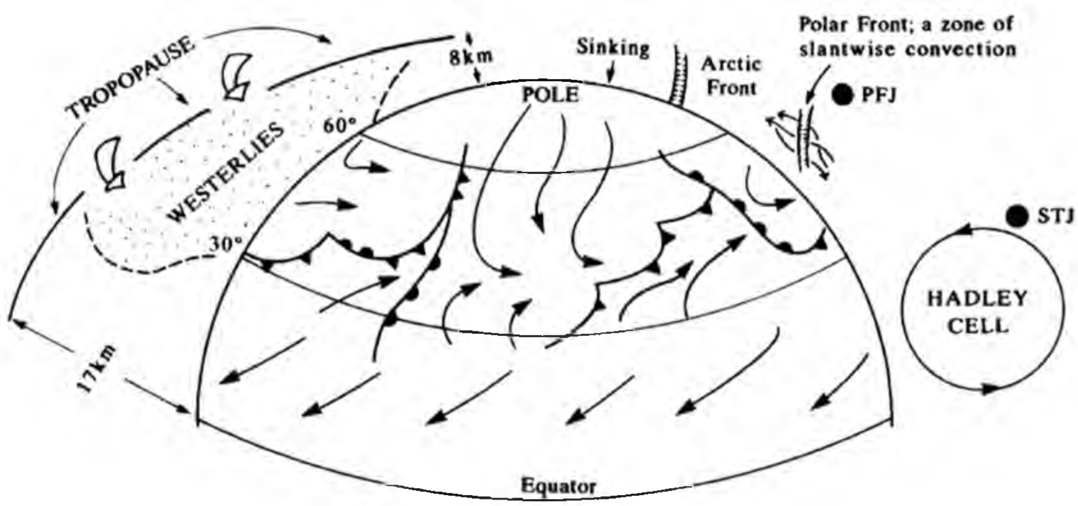


Figure 2.5 A latitudinal cross-section of the general circulation of the atmosphere (at right-hand side). STJ and PFJ are the Sub Tropical and Polar Front Jet streams embedded in the broad zone of westerly flow (see left-hand side). The tropopause (left-hand side) is the top of that part of the atmosphere where weather systems occur, representing a 'lid' on the troposphere, although there is some exchange with the stratosphere above. A schematic of the airflow at the Earth's surface is also shown, indicating that the Polar Front is heavily perturbed on a day-to-day basis, and that even in the broad band of westerlies, airflow with an easterly component does occur. The surface easterlies in tropical latitudes are the trade winds.

The average annual latitudinal distribution of evaporation and precipitation (Figure 2.6) confirms the role of the tropical Hadley Cell and the mid-latitude cyclones, in both hemispheres, in global atmospheric energy transport. The subtropical dry zones (where evaporation is greater than precipitation) correspond with the subsiding and equatorward moving parts of the Hadley Cell circulations. These provide atmospheric moisture transport into the tropical 'engine', where uplift produces condensation (releasing latent heat) and precipitation. The relatively wet mid-latitudes are the zones of high cyclone frequency, where frontal uplift leads to condensation of water vapour evaporated from the ocean surfaces.

The Earth rotates west-to-east and, because the atmosphere clings to the Earth, it rotates with it. When viewed from the Earth's surface, however, some winds travel from east-to-west (i.e., they are easterly relative to an observer at the surface). In absolute terms, they are rotating in a west-to-east sense at a rate which is slower than that of the Earth's surface. In broad terms easterlies occur in the latitudes of the lower limbs of the tropical Hadley Cells and in a narrower latitudinal band of their upper limbs; and restricted regions of high latitudes – see Figure 2.5. On the other hand, of course, the westerlies are rotating west-to-east at a rate which is faster than that of the Earth's surface. The drag of the Earth's surface extracts angular momentum from the westerlies. In order to keep blowing, there must be a reliable mechanism to inject angular momentum into the westerlies from the easterlies, which have angular momentum fed into them from the Earth's surface.

In tropical latitudes, the Hadley Cell plays an important role in this poleward angular momentum transport. In mid-latitudes, however, it is the cyclones, and the waves in the westerlies associated with them, which play the important role in the

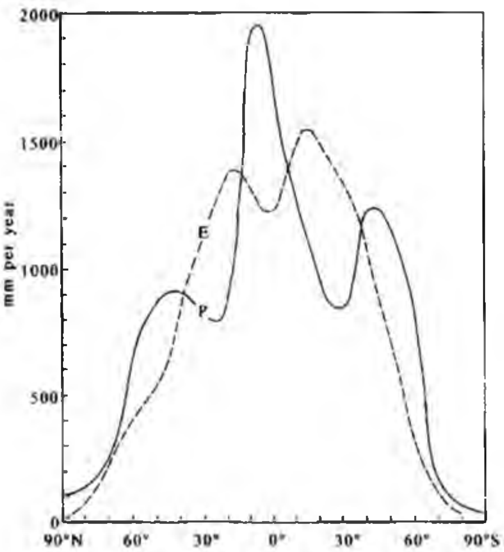


Figure 2.6 The average latitudinal distribution of precipitation (P) and evaporation (E) in mm per year.

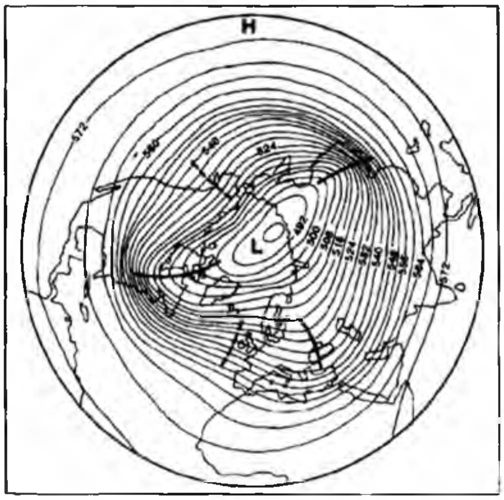
poleward transport of angular momentum. They accomplish this by exchanging air with a stronger westerly component (moving polewards) with air with a weaker westerly component (equatorwards). It has been calculated that five simultaneous well-developed mid-latitude cyclones can provide all the necessary angular momentum transport required in winter. Since similar calculations lead to similar conclusions about poleward heat transport, it is clear that the mid-latitude cyclone belt is a crucial feature of the general circulation of the atmosphere. Since it also determines the character of the climate of the British Isles, it is appropriate that we examine the mid-latitude westerlies a little more closely.

The westerlies

In the following discussion, we shall focus on the Northern Hemisphere. Much is also pertinent to the Southern Hemisphere, although the contrasting land and sea distributions of the two hemispheres do result in some differences. In the Northern Hemisphere, the subtropical jet stream and the polar front jet stream represent the zones of strongest flow within the broad band of mid-latitude westerlies. The westerlies are more vigorous and extensive in winter; their kinetic energy (that energy due to motion) is three times greater in winter than in summer. The flow in the upper westerlies can be particularly strong, with wind speeds up to 140 ms^{-1} near the top of the **troposphere**. This represents a jet stream in the strict sense of the term which says that jet streams should be characterised by wind speeds of 30 ms^{-1} or above. The term is also used more loosely to describe the locally stronger flow of the subtropical and polar front jet streams.

A prominent feature of the westerlies is their wave-like form. The wave pattern is more noticeable, and simpler, away from the surface. The **500 hPa pressure level** (at a height of around 5.5 km at the latitude of the British Isles) is commonly used to describe the free atmosphere westerlies, although wind speeds are greater at higher levels. On a day-to-day basis the flow can be very complex, but if the flow is averaged over a few days there tends to be five waves in the westerlies encircling the Northern Hemisphere. Waves with this sort of wavelength are called planetary waves. If the averaging period is increased to seasonal time-scales, then the further smoothing produces three waves in the Northern Hemisphere winter and four in the summer (Figure 2.7). The changeover from the winter to summer westerly pattern occurs relatively abruptly around June and back again, equally abruptly, around October (see Chapter 8). There is a general relationship between westerly vigour and the number of planetary waves, even over shorter periods. The stronger the westerly (or zonal) flow, the smaller the number of planetary waves.

In reality, the planetary waves drift slowly eastwards, but statistical smoothing highlights the major regions of occurrence. The main planetary wave **troughs** are situated over eastern North America and over the east coast of Asia. These troughs are especially pronounced in winter. A weaker planetary wave trough is located over Europe, between about



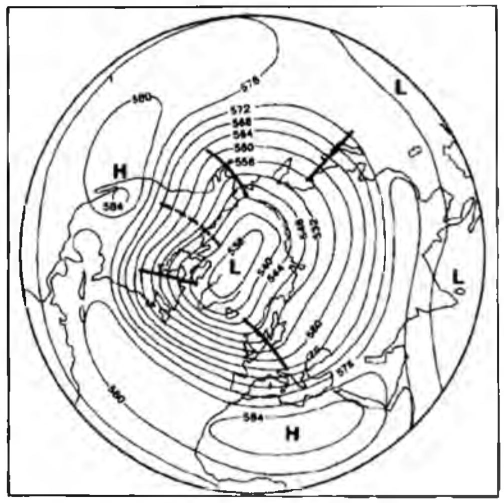


Figure 2.7 Average height (tens of metres) of the 500 hPa surface in January (left) and July (right). The mid-latitude westerlies blow parallel to the contours – the closer the contour spacing the stronger the wind. Dashed bold lines indicate the average position of ridges, continuous bold lines the position of troughs.

10–60°E. The position and strength of this European trough are much more variable than are those of the North American and East Asian troughs, which have more anchored positions. The European trough is very sensitive to changes in the westerly circulation. This is an important characteristic for the European and British climate since, as we shall see later, the overlying westerly waves exert a strong control on surface weather patterns. It is of some interest that the average position of the European trough axis, when averaged by decade, has varied by as much as 20° longitude over the last 200 years. A planetary wave crest (ridge) is located to the west of the European trough, with a position somewhere between 10°E and 30°W. This ridge is less discernible in summer.

The preferred positions of the planetary waves owe much to the continental distributions of land masses and to the presence of major mountain ranges. Experiments using dishpans to model the Earth–atmosphere system have replicated some of the major features of the general circulation without an underlying topography (see Box 2.1). The nature of the underlying surface, however, strongly conditions the location and character of the planetary waves.

BOX 2.1 DISHPAN EXPERIMENTS

Dishpans are physical models of the Earth–atmosphere system, so named because the Earth is modelled by a dishpan and the atmosphere is modelled by a fluid contained therein. A temperature gradient is applied across the dishpan (the outside representing the 'equator' and the centre the 'pole') and the dishpan is rotated. The resulting circulation of the fluid reproduces many features observed in the atmosphere, including the westerly planetary waves (as well as circulations which resemble cyclones and anticyclones). The dishpan models do not have the equivalent of mountain ranges or land masses, so the planetary waves and other circulation features form as a consequence of a pole–equator temperature gradient impressed on a rotating fluid system, irrespective of the underlying topography. Nevertheless, in the real Earth–atmosphere system, it is obvious that the location and precise character of the planetary waves are strongly conditioned by the nature of the underlying surface. Before we return to the real atmosphere, it is worth pointing out another aspect of dishpan circulation behaviour that is analogous to observed behaviour in the atmosphere. Some dishpan experiments exhibit semi-regular fluctuations in the westerly wavelength and amplitude. The real atmosphere also tends to demonstrate this type of behaviour; the so-called (Zonal) Index Cycle which, typically, spans several weeks. One extreme is strong westerly flow (high Zonal Index) with small amplitude waves; the other extreme is weaker westerly flow with large amplitude waves producing more north-to-south and south-to-north (meridional) flow.

An important property of atmospheric circulations is **vorticity**, which is a measure of spin within a fluid. If we assume that an airstream, flowing west-to-east, is approaching a large mountain barrier, such as the North American Rockies, then as it is forced to rise over the mountain range and since the tropopause acts as a sort of 'lid', it becomes vertically squeezed as it passes over the peaks. Meteorology textbooks show mathematically that there are important links between vertical squeezing (and stretching) and vorticity. In our case, it leads to a decrease in the vorticity of the airstream which in the Northern Hemisphere represents a clockwise turning – anticyclonic curvature. As it passes over the mountain barrier, therefore, the air starts to turn equatorwards. As the air flows beyond the mountain barrier, however, it is allowed to stretch vertically. This now leads to an increase in vorticity, which in the Northern Hemisphere leads to anticlockwise turning – cyclonic curvature. The result is a (cyclonic) trough in the westerlies in the lee of the mountains. This explains the anchoring of a pronounced wave trough in the westerlies over eastern North America (Figure 2.7).

Once such a large-scale wave has been initiated, there are good reasons why a wave form should be maintained downstream. In large-scale motion, at mid-troposphere levels (around 500–600 hPa), there is a need to conserve absolute vorticity. This is the sum of atmospheric vorticity and the local vorticity of the Earth's surface (the Earth is 'spinning' and therefore has vorticity). The local value of the vorticity of the Earth's surface is known as the **Coriolis parameter** (named after Gustave-Gaspard Coriolis, see Chapter 14) and is proportional to the sine of the latitude. It is therefore at a minimum at the Equator and a maximum at the poles. The large-scale westerly motion downstream of the North American trough has a poleward component; hence the Coriolis parameter is increasing. To maintain absolute vorticity, the atmospheric vorticity must decrease. The airstream starts to take on increasing anticyclonic curvature, eventually turning through a ridge, and so eventually taking on an equatorward component. Now the Coriolis parameter is decreasing and absolute vorticity must be conserved by the atmosphere adopting more cyclonic curvature. This downstream oscillation would continue indefinitely in the absence of any other factors and explains the point made above that the North American trough largely controls the behaviour of the European trough. In reality, other factors do come into play and the character of downstream waves is influenced by, for example, surface temperature patterns.

Cyclone waves in the westerlies

On a day-to-day basis, the smoothed planetary wave pattern is obscured by the superimposition of smaller wavy perturbations which appear, grow and decay

Figure 2.8 Schematic representation of the development of a cyclone wave. The heavier lines represent the flow at upper levels, with the heaviest line showing the most rapidly moving air. The frontal depression at the surface is denoted by the warm (semi-circles) and cold (triangles) fronts and by the lightest lines which represent the surface isobars (lines of constant atmospheric pressure). Adapted from McIlveen (1992) (see p. 32 Note 13).

over a few days. These waves are the result of instabilities in the smooth, planetary flow and they generally propagate rapidly eastwards, apparently steered by the flow in the planetary waves, at a rate of about 10–12° longitude per day. Typically, their amplitude increases by a factor of two to three over a couple of days. Their wavelength is of the order of 3–6,000 km; so the number of waves around the hemisphere is between 6 and 10. These unstable perturbations are known as **baroclinic waves**, or cyclone waves. One of the most important destabilising factors responsible for these waves is the marked increase of wind speed with height in the polar front jet stream zone. This, combined with the strong horizontal temperature gradients concentrated into this zone, produces the sort of instability which leads to cyclone wave development.

Figure 2.8 shows the development of a cyclone wave at the level of the polar front jet stream and its associated frontal depression at the Earth's surface. In order for these cyclone waves to grow, the kinetic energy of the wave must increase. The kinetic energy is made available by warm air rising and cold air sinking. The upsliding of warm air, exchanging with the downsliding of cold air across the polar front zone, takes place along a slope of only around 1°. The horizontal and vertical temperature gradients in this zone are such, however, that they make the conversion of potential energy (dependent on the relative positions of warm and cold air) to kinetic energy more effective than the very shallow slope of the exchange might suggest. As far as energy conversion is concerned, what is happening is convection (although we generally understand convection to be a process which involves greater vertical exchange); hence the term 'slantwise convection'. The cyclone waves pass kinetic energy into the westerly wind belt, playing the crucial role in the maintenance of the general circulation which has been mentioned before. Early notions of the mid-latitude cyclones were that they were akin to turbulent eddies, which are maintained by the energy of the mean flow. The opposite is in fact the case.

Frontal depressions often form in families. Figure 2.9 is an idealised picture of four planetary waves in the 500 hPa flow and associated frontal depressions (or cyclone waves). Four families of frontal depressions are shown with new depressions forming on the trailing cold front of the 'parent', or **occluding**, depression. The depression families are seen to lie on the poleward-moving limbs of the planetary waves, downstream of the troughs. The reason for this is as follows.

The planetary wave troughs have cyclonic curvature and the planetary wave ridges anticyclonic curvature. For reasons we will not explain here, this

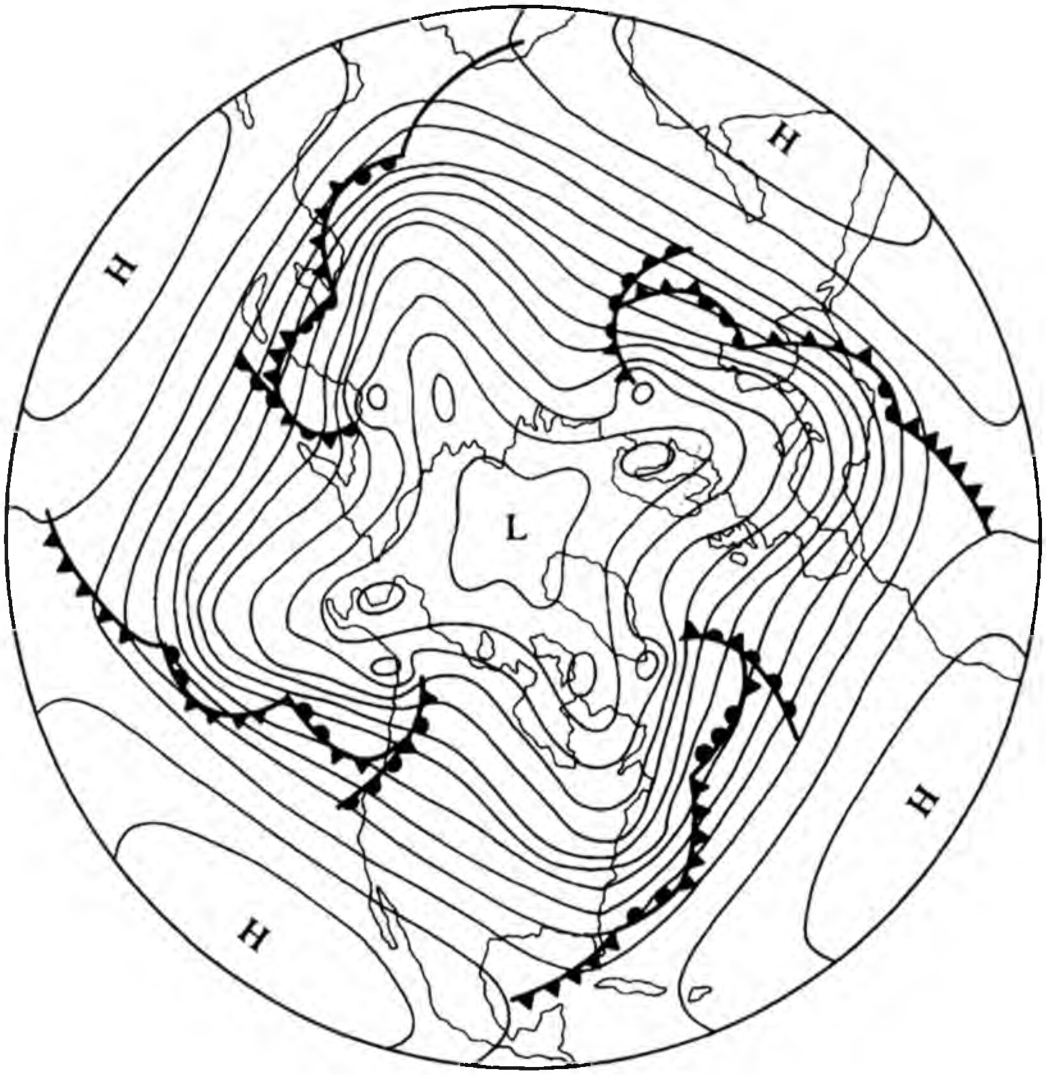


Figure 2.9 Four idealised planetary waves, showing how the formation of families of frontal depressions is favoured under the poleward-moving limb of the waves (see text and Figure 2.10). In the real atmosphere on a day-to-day basis, the westerly wave pattern is more complex than this. The isolines represent heights of the 500 hPa pressure surface.

has the effect of horizontally stretching volumes of air passing through the poleward-moving limb of the planetary waves at upper levels, whereas the effect is one of horizontal squeezing on the equatorward-moving limb. This upper stretching (**divergence**) and upper squeezing (**convergence**) produces compensatory patterns of convergence and divergence near the surface (Figure 2.10). The surface convergence leads to an increase in cyclonic vorticity, and the surface divergence leads to an increase in anti-

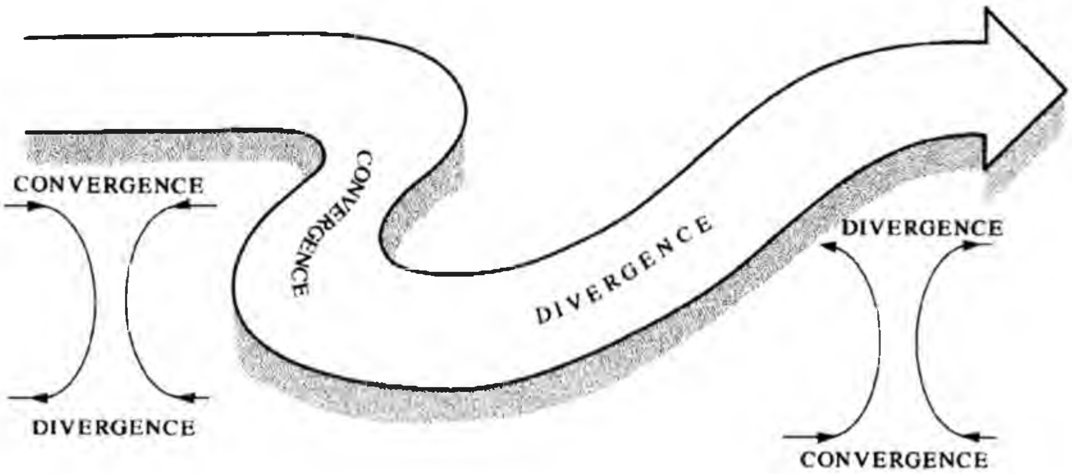


Figure 2.10 Convergence at upper levels in the equatorward-moving limb of a planetary wave is compensated by divergence closer to the Earth's surface. Under the poleward-moving limb, the upper divergence is compensated by surface convergence.

cyclonic vorticity. Consequently, cyclone development is encouraged downstream of a planetary wave trough, and is inhibited upstream of the trough, where anticyclone development is more favoured. The upper flow pattern is, therefore, seen to be an important control on the development of cyclones and travelling anticyclones.

From our discussion it is probably quite easy to get the impression that the polar front is a continuous band snaking round the hemisphere, coinciding with the path of the polar front jet stream (although Figures 2.5 and 2.9 have indicated that this is a misleading simplicity). In reality, the geography of the Earth's surface, including the influence of ocean currents, favours certain zones of 'frontogenesis'. One such zone stretches from the south-eastern United States, across the North Atlantic, towards the British Isles. It is weaker, and less tilted, in summer compared to winter (in particular, its eastern end shifts northwards). The zone off North America is the strongest part of the North Atlantic Polar Front. In winter there is a large temperature contrast between a land mass with an extensive snowcover and warm offshore currents. The positioning of the poleward-moving limb of the planetary wave anchored over North America (and, hence, upper divergence) over one of the strongest frontal zones in the Northern Hemisphere favours the development of cyclonic waves. These are then steered by the upper flow, growing as they move towards Iceland and northern Scandinavia. This depression track is much weaker in summer and tends to take a more northerly course over the eastern North Atlantic. Something like 170 depressions a year follow this track, although there are significant variations from year to year and from decade to decade. There are also variations in the typical depression tracks, depending on, amongst other factors, variations in the westerly wave pattern. Some of these variations will be discussed below.

We have spent some time discussing the westerlies and cyclonic waves and their role in the general circulation of the atmosphere. We have concentrated on the area of the North Atlantic, although there are other regions, in both hemispheres, of important frontogenesis and cyclone wave development. We will return to some features of the global-scale circulation later, but the reason for the present emphasis is that the travelling cyclones, and their intervening high pressure ridges, create the essential character of the British weather. Seasonal fluctuations in their behaviour control much of the regional-scale climate, and long-term variations in behaviour – responding to some other feature of the circulation of the atmosphere and oceans – can help explain much of the observed longer-term fluctuations in the climate of the British Isles.

We do not intend to describe in detail the particular pattern of weather associated with frontal depressions. Most readers will be familiar with the precipitation bands and temperature changes associated with the passage of the **warm** and **cold fronts** in a frontal depression, which can swing over the British Isles even when the depression centre is passing far to the north. The mobility and precise path of a particular system are important characteristics for daily weather. The severe windstorms which afflicted Western Europe in the early 1990s are a graphic example (see Chapter 11). The passage of the intervening high pressure ridge, with its different weather (weaker winds, clearer skies, cool winter nights, etc.; although high pressure can also produce persistently cloudy skies in winter) generally provides a clear contrast to their fellow travelling depressions. Together, they can produce a characteristic 2–3 day sequence of weather – although this sequence is regularly disturbed.

Blocking

When we were discussing the westerlies, we noted that the wave pattern exhibits fluctuations in wavelength and amplitude. This characteristic is known as the Index Cycle.[3] The term is derived from the parameter, the Zonal Index, which is a measure of the strength of the middle-latitude westerly winds.

At low values of the Zonal Index, the westerly flow is weak and the wave pattern becomes so exaggerated that there are large areas of higher-than-average and lower-than-average atmospheric pressure (positive and negative pressure anomalies, respectively) encircling mid-latitudes. Once the wave pattern starts to become strongly perturbed in one longitudinal band, there is a tendency for the large-scale amplitude pattern to spread throughout the mid-latitudes within days. A chain of four positive and four negative pressure anomalies encircling the globe is a common pattern. When the positive anomalies become well-established and remain quasi-stationary, they are known as 'blocks'. The longitudes of the planetary wave ridges at 150°W and 15°W (the latter position being off the west European coastline, see Figure 2.7) are particularly favoured for the development of blocks.

When the mid-latitude flow is zonal (high Zonal Index), the vigorous westerly flow over extensive regions means that fast-developing cyclonic waves move quickly eastwards. When the flow is meridional (low Zonal Index), then the development and passage of the cyclonic waves is 'blocked' over extensive regions. **Blocking** is most common in spring/early summer, although it can occur at any time of year. A typical position for a block to be centred is about 15°W. In line with the seasonal shifts in the planetary wave pattern, however, there is a tendency for the preferred position of the block axis to move from east of the British Isles in winter to the west of Ireland in May, and continue out into the Atlantic to its most westerly position in summer, whence it starts its slow progress eastwards again. The high frequency of blocking in spring/early summer when averaged over 100 years is quite pronounced. This common spring/early summer block allows more airflow from the north, south and east, at the expense of westerly, progressive, conditions. By the end of June, with the block declining or starting to shift eastwards, more westerly flow has been resumed with generally more precipitation. Some climatologists have described this late June period as heralding the '**European Monsoon**'. Chapter 8 further discusses this characteristic of the annual cycle of weather over the British Isles.

Fluctuations in the Zonal Index, and the associated hemispheric-scale adjustments to the mid-latitude flow, operate on a time-scale such that a particular region's weather over significant parts of whole seasons may be strongly influenced by them. That being so, at least part of the year-to-year variations

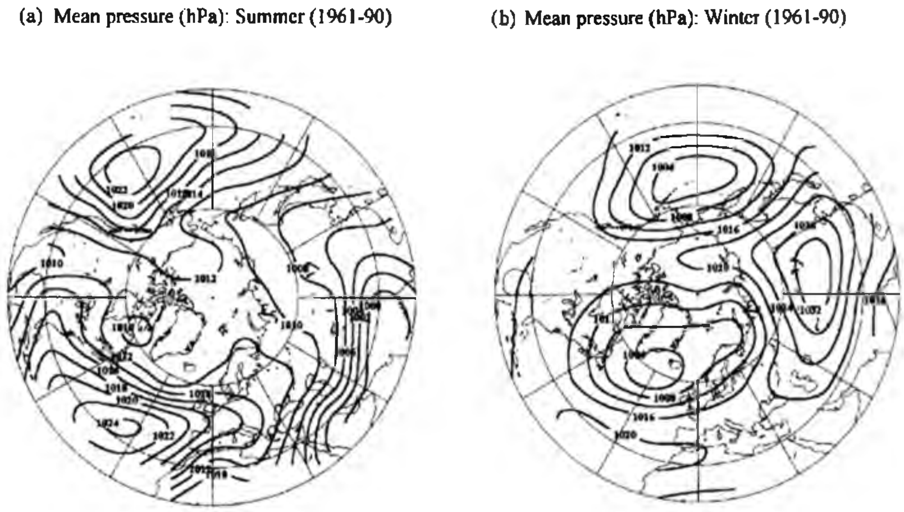


Figure 2.11 Average mean sea-level pressure (hPa) for the Northern Hemisphere (1961–90), summer (a) and winter (b). Note the change in contour interval between the seasons. Pressure gradients are stronger in winter and, as a result, the atmospheric circulation is more vigorous.

which are such a strong feature of short-term climatic variations can be explained by the 'Index Cycle'.

SURFACE PRESSURE PATTERNS

Average surface airflow is parallel to the isobars and its vigour is proportional to the pressure gradient – the tighter the isobars, the stronger the flow. The most noticeable difference between the maps of summer and winter surface pressure over the Northern Hemisphere (Figure 2.11) is the replacement of winter high pressure by summer low pressure over Eurasia. This produces the very pronounced seasonal reversal of flow which is the south-east Asian Monsoon. As far as the most immediate features of relevance for the British Isles are concerned, the dominant centres of action are the area of low pressure near Iceland (the Icelandic Low), most marked in winter, and high pressure to the west of Spain, which extends over western Europe in summer as the Icelandic Low migrates westwards.

The high pressure area is known as the Azores High and is an extension of the permanent subtropical high pressure over the Atlantic. This is matched by another over the Pacific and which are broad zones of subsiding air reflecting descent in the Hadley Cell (Figure 2.5). The Icelandic Low is really a statistical manifestation of the passage of travelling cyclones over this part of the Atlantic. The eastern part of Europe is influenced by the western extremity of the Siberian High in winter; this is a shallow high pressure caused by radiational cooling of the Eurasian land mass. During periods of low Zonal Index, there can be outflows of very cold air from the Siberian High which lead to cold weather over the British Isles. In summer, the high pressure is replaced by low pressure, caused by heating of the Eurasian land mass. The surface pressure patterns confirm what we already know; winter flow over the Atlantic and into the British Isles is much stronger than summer flow. The summer reduction in the Icelandic Low reflects the decline in cyclone vigour and the northwards expansion of the Azores High reflects the tendency of all components of the global climate to 'follow the Sun'.

Since the two centres of action – the Icelandic Low and the Azores High – dominate the pattern of surface pressure over the eastern Atlantic, a useful index to describe conditions upwind of the British Isles is the difference in pressure between the Azores region and over Iceland. This is, in effect, a 'local' Zonal Index. Besides describing part of the annual variation in surface pressure over the Atlantic, this Azores/Iceland pressure index also characterises changes in the strengths and positions of the Icelandic Low and the Azores High. The behaviour which the index characterises is known as the **North Atlantic Oscillation** (NAO). The oscillation is the link between the two centres of action – when the Azores High is more intense (higher pressure), the Icelandic Low also tends to be more intense (lower pressure). This NAO signal, when averaged over several years, is present for all seasons, although it changes its precise character with the seasons. The NAO is an important component of the interannual variability of the whole Northern Hemisphere circulation.

It will come as no surprise that there are links between the behaviour of the NAO index and the weather experienced over the British Isles during a particular year (see Chapter 9 for its relationship with temperature). Changes in the circulation patterns over the Atlantic are associated with shifts in storm tracks – high values of the NAO index push storms further into Northern Europe, accompanied by higher temperatures than usual. Stronger Atlantic westerlies increase the atmospheric transport of moisture into northern Europe. This leads to heavier precipitation over the northern half of the British Isles, although indications are that precipitation may be reduced over the southern half. So, even though the initial control is the North Atlantic large-scale circulation, we still have to consider the sub-regional scale response. This reflects the similarity of scale between the British Isles and the synoptic systems which produce the Islands' day-to-day weather.

On occasions the NAO index is negative; that is, the normal south-to-north pressure gradient is reversed. This is an extreme circulation mode, reflecting a strong pattern of blocking and leads to flow with an easterly component over the North Atlantic European sector. As we indicated earlier, periods of weeks or, occasionally whole seasons, can be dominated by such conditions (see Box 2.2).

LINKS WITH THE OCEAN

Variations in the NAO have been linked to sea-surface temperature (SST) changes in the North Atlantic. From year to year, the SST patterns are probably caused (or forced) by the atmospheric circulation, with the surface wind influencing the ocean circulation and hence the distribution of SST anomalies. The picture is not clear, however, and there is evidence that SST patterns in the western part of the North Atlantic influence the British weather on time-scales of months.[4] Warm SST anomalies in this part of the ocean tend to precede a greater incidence of cyclonic circulations over the British Isles in the following months, whereas a cold SST anomaly is frequently followed by months which are more anti-cyclonic in character.

The precise linking mechanism between the ocean and the atmosphere appears to be related to the shift in the position of the maximum surface temperature gradient, affecting the formation and path of cyclone waves. Over time-scales of several years to decades, although two-way interactions between the atmosphere and ocean still operate, there are indications that the SST anomalies (this time over a larger area of the North Atlantic Ocean) are playing an important role in forcing the circulation of the overlying atmosphere, and in influencing the climate of Europe. Figure 2.12 shows that SSTs over the North Atlantic were relatively low up to the 1920s, higher up to the 1960s, then lower thereafter. There are indications that the high SSTs in the 1940s and 1950s were associated with the production of more cyclones over the mid-North Atlantic Ocean at around 45°N. There are also some hints of links between the SST anomalies and the frequency of different types of circulation over the British Isles. Robert Ratcliffe and Roy Murray amongst others[5] have emphasised, however, that it is likely to be the precise pattern of SSTs which is important for

BOX 2.2 BLOCKING AND EXTREME SEASONAL WEATHER

One of the lowest values of the NAO index occurred in 1963. This winter (January to March) was one of the coldest in the last 250 years in the British Isles, when the temperature in parts of England did not rise above 0°C for three months, because of the persistent easterly flow from the cold European mainland. The first three months of 1996 were also rather cold, as a result of blocking highs in the Scandinavia to east North Atlantic region, leading to persistent easterly or northerly flow over the British Isles. This recent cold winter was a timely reminder to a public, which had been fed an oversimplified diet of global warming by many parts of the media, that interannual variability is still a strong characteristic of the British Isles climate. Understanding the regional response of climate to the enhanced greenhouse effect demands rather more sophisticated consideration (see Chapter 15).

The 1963 block was centred over Iceland. Blocks do not have to occur over the Atlantic to have a dominating control on the British weather over a season. Another extreme winter, in 1947, was caused by a blocking high over Scandinavia. This was less noticeable in the NAO index, but had the effect of steering depressions further south than usual, over the southern half of the British Isles, producing much snowfall (snow fell somewhere over the British Isles every day from 22 January to 17 March in 1947). In this case, the Atlantic depressions provided the moisture source for a deep-white winter, whereas the extremely cold 1963 winter was relatively deficient in snowfall because of the easterly flow from the dry European mainland. So, although the prime control on the British climate comes from the atmospheric circulation over the Atlantic, and part of this control can be represented by simple indices such as the NAO index, we need to remember that the precise configuration of anomalous circulations is important.

Blocks can also produce anomalously hot or dry summers. The prolonged drought of 1975–6 (one of the driest 18 month periods on record over England and Wales, see Chapter 10) resulted from blocking summer highs over, or close to, the British Isles. The clear settled conditions resulted in high temperatures and the rain-bearing depressions were steered to the north and to the south, around the blocks. Persistent high pressure also dominated in the very hot summer of 1995; a particularly pronounced ridge in the westerlies occupied a position which stretched from the British Isles to as far east as 25°E (cf. Figure 2.7).

the British climate, in particular the position/orientation of the zone of maximum SST gradient across the North Atlantic.

SST patterns in parts of the North Atlantic are intimately linked with sea-ice distributions and there are clearly established relationships between sea-ice extent around Iceland and the European climate.[6] Periods of extensive northerly winds over the north-east Atlantic and western Europe, for example, bring cooling to the continental landmass and ice to the shores of Iceland. The warmth of the 1920s and 1930s round the North Atlantic sector was associated with strong westerly winds and a period when ice was a relatively infrequent visitor to Iceland. These relationships arise because, on this geographical scale, the atmospheric circulation is the primary cause of the fluctuations in ice and climate conditions. Locally, however, the advance and retreat of the ice edge is associated with a marked change in surface heating and albedo and can exert a strong influence on the overlying atmosphere.

The interactions between the ocean surface and the atmosphere over the North Atlantic, and the consequent 'downstream' impact on the weather or

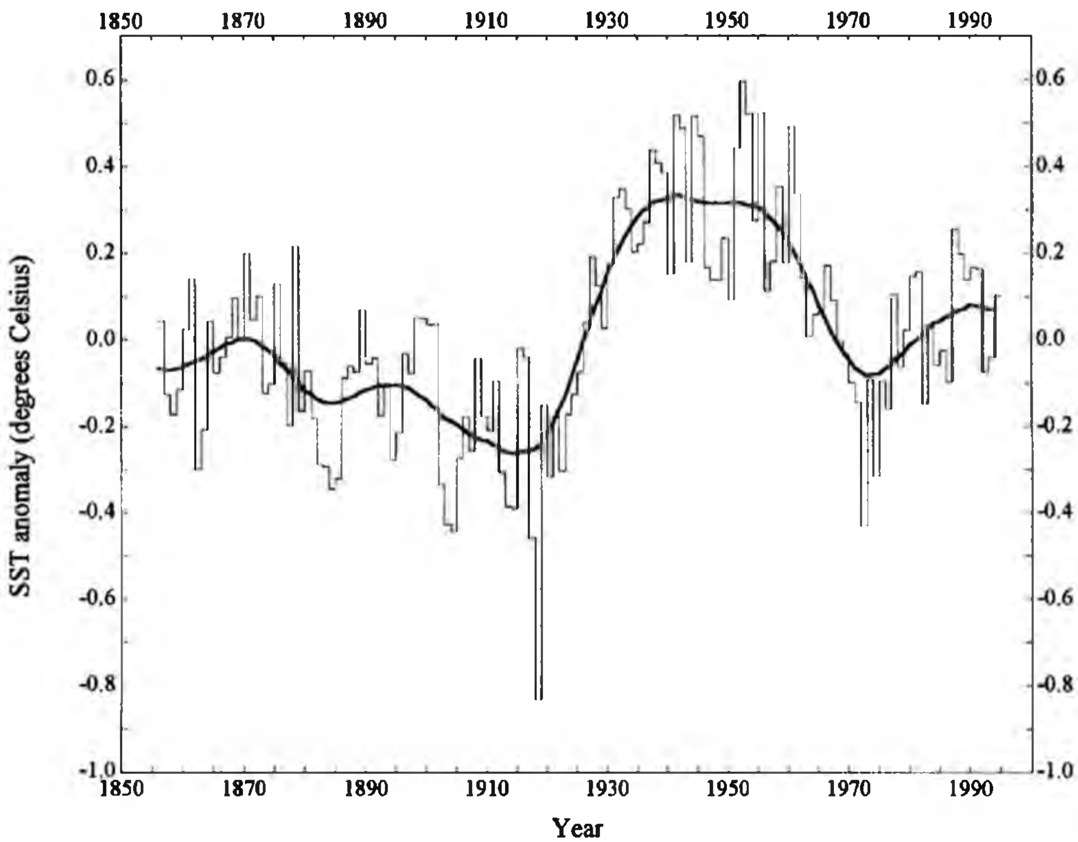


Figure 2.12 Annual sea-surface temperature anomalies, with respect to the 1961–90 mean, in the North Atlantic Ocean from 1856 to 1995. The smooth line is the result of applying a filter which emphasises variations on time-scales greater than 30 years. The region is defined as 20° to 70°N and from 0° to 80°W.

climate of the British Isles, are complex and two-way. The nature of the interaction depends on time-scale and geographical scale. We must also consider the atmosphere–ocean interactions on a wider, indeed global, scale. Similarly, we have to extend our interest to deep ocean circulations, not only in the northern part of the Atlantic Basin, but further south and, as for the atmosphere, to other ocean basins.

First, we examine the deep overturning of water which is particularly vigorous in the Atlantic Ocean due to the formation of North Atlantic Deep Water (NADW). The sinking of this water mass in the northernmost parts of the ocean causes the Gulf Stream and its extension – the **North Atlantic Drift** – to turn more northward to replace the sinking water. This does not happen in the Pacific Ocean which has no deep water formation. The sinking of the NADW results from its high density, which is a consequence of its high salinity as much as its low temperature. It is self-sustaining, to some degree, since its high salinity is due to the northward transport of saline water from more tropical latitudes. Its warmth is also important, since it induces evaporation, further increasing salinity.

Since the sinking is sensitive to changes in the input of freshwater to the North Atlantic, the suggestion has been made that the overturning in the Atlantic could vary in strength, stop, or even reverse. Such switches could occur over very short time-scales. There is evidence that the warming from the last **Ice Age** was interrupted around 10,000 years ago by a dramatic return towards Ice Age temperatures, sometimes called the **Younger Dryas** period, only to be followed by a rapid recovery, all within 1,000 years (see Chapter 5). This rapid climate deterioration was probably caused by the Atlantic overturning being disturbed by a large input of freshwater of low density into the North Atlantic from the melting of ice. The mechanisms which are likely to be involved in these fluctuations are complex.[7]

There are some tantalising indications from computer models that the sinking in the North Atlantic may fluctuate over shorter time-scales – possible oscillations of around 40–60 years have been reported.[8] The reason for such oscillatory behaviour is uncertain, but some aspects may be triggered by a short-term change in the input of freshwater in the sinking region north of 60°N (more or less ice melt, or even heavy precipitation). Other aspects may be more self-sustaining. The implication for the British climate is that these multi-decadal oscillations may be reflected in changes in the North Atlantic SST patterns which, we know, have an important downstream influence. The reason why these computer experiments are tantalising is that climate reconstructions from tree-rings from parts of Europe also exhibit oscillatory-type behaviour on the same time-scale.[9]

We have mentioned the **teleconnection** character of the North Atlantic Oscillation. The most pronounced teleconnections globally are those associated with the El Niño Southern Oscillation (ENSO). This is a surface pressure oscillation across the tropical Pacific, related to Pacific Ocean currents.[10] The ENSO dominates interannual climate variability in tropical latitudes and there is a pronounced teleconnection between it and surface pressure over the North Pacific Ocean and North America. Although strong ENSO signals have not been detected in the atmospheric circulation over the North Atlantic/European sector, there are some indications of possible linkages with European-scale weather patterns.[11] There appear to be weak links with some aspects of the British climate, particularly the frequency of anticyclonic and cyclonic weather types in winter, and winter precipitation over England and Wales. The strongest links appear to be in January and February.[12] It is possible that more pronounced ENSO signals in British climate will emerge as research progresses.

AIR MASSES

We have already discussed how air flowing over the North Atlantic to the British Isles has a different character to that flowing out of Siberia in winter – relatively warm and moist versus cold and dry. The concept of air masses is a useful one and underpins the usefulness of the weather type classifications (see Chapter 8), since direction of airflow is one of the bases of identification of many types. When air resides in a source region for weeks it starts to develop a homogeneous character, a few kilometres deep, which it acquires from its source region. The source regions are geographically distinct, cover hundreds of thousands of square kilometres and differ between summer and winter.

Two of the winter land source regions are Canada and (approximately) the former Soviet Union, the origin of 'continental polar' air masses. The character of air masses from these source regions will clearly differ considerably between summer and winter. There are also seasonally differing source regions for 'continental arctic' and 'continental tropical' air masses, as well as 'maritime arctic', 'maritime polar' and 'maritime tropical' air masses. The path the air mass takes from its source region to the British Isles is important. For example, a relatively cool air mass flowing over a warmer surface will often bring convective clouds, good visibility and gusty winds; warmer air flowing over a similarly warm surface may produce stratus cloud, fog and poor visibility. Many weather features result from such modification of an air mass along its path; the air mass type may change from such modification – 'continental polar' air flowing out of Canada over the warm North Atlantic Drift may have developed into a cool moist 'maritime polar' air mass by mid-Atlantic, producing bright periods and some showers.

SMALLER WEATHER SYSTEMS AND LOCAL INFLUENCES

A number of circulations on a smaller scale than the features we have been describing make contributions to the climatic character of the British Isles. This short section is not all-inclusive. We will not, for example, look at thunderstorms which can provide a significant proportion of summer rainfall (see Chapter 3), nor at mountain and valley wind systems which can influence local wind climatology, or cold nocturnal drainage flow which may fill 'frost hollows'.

Midway in scale between these circulations and the cyclones and travelling depressions are **polar lows**. We shall refer briefly to these circulations since, although they are relatively small (often several hundreds of kilometres) and shallow (5 km), they can produce severe weather over parts of the British Isles, with strong gusts and contributing to much of the heavier snowfalls. They represent a special case of cyclone formation in that they are generally non-frontal. They usually develop over the ocean in the northerly 'maritime polar' or 'maritime arctic' airflow to the rear of a cold front, often between Iceland and the British Isles (see Figure 2.13). The cold airflow across zones of relatively high SST gradients provides the mechanism for their formation.

Another type of low, but of entirely different origin and type, is the so-called heat low. Localised heating of land in summer can produce such features, usually in the afternoon. There are occasions, however, where they may survive night-time cooling and persist for some days. They can be relatively small-scale (for example, over East Anglia), or they may cover an area such as most of England. Thunderstorms may develop in the heat lows.[13]

Figure 2.13 An infra-red image of a polar low near the Faroes, 25 November 1978, observed from the NOAA 5 satellite. This day is classified as NW in the Lamb Catalogue (see Appendix B).

Within tens of kilometres of the coastline, the day-to-day weather may be modified by sea-breezes.[14] They are also caused by the daytime heating of the land, which produces a pressure gradient between the sea (high pressure) and the land (low pressure). During late morning a sea-breeze starts to blow inshore and penetrates inland causing moister, cooler conditions, frequently accompanied by cloud. Typically, during summer at some British locations, sea-breezes will blow on 20–30 per cent of days, but there are periods when they are much more frequent. Consequently, they may affect the character of a whole summer at some near-coastal locations.

Another important factor for local climate is the local orography. A glance at a long-term precipitation map of the British Isles provides clear confirmation of this (see Plate 4 or Appendix A). When moist air is forced to rise over high land, the air can be cooled to a point where condensation occurs and the precipitation process starts. There is a distinct west-to-east gradient in precipitation over the British Isles which largely reflects this orographic effect on moist air blowing in from the Atlantic. Another reason for enhanced precipitation over the highest land is that the passage of fronts can be slowed down. Detailed precipitation maps show a dependency of precipitation on elevation, even where the orography is not pronounced and even in eastern locations. Modest orographic enhancement of precipitation is apparent, for example, even in the far-eastern and relatively flat Norfolk. It goes without saying that cloud also is more common over higher land.

We have noted the importance of the land surface during our discussion of global-scale radiation and heat balances. It is not only albedo (which can change seasonally with vegetation changes) which is important, but also heat conductivity. The thermal conductivity of the soil is an important factor in the response of the surface to changes in net radiation. Soil conductivity is strongly influenced by water content. So, if the soil is coarse, sandy and dry (thus containing a lot of air; a good insulator), night-time radiational cooling will not be offset by the conduction of heat from lower levels in the soil. An example of this, again from East Anglia, is the sandy soil area of the Breckland area of Norfolk, where night-time minimum temperatures can be 3–4°C lower than in the surrounding areas where the soil is less freely drained (e.g., Santon Downham – see Chapter 3).

Other important surface differences occur in urban areas. The city fabric acts as a 'storage heater' maintaining night-time temperatures above those of the surrounding regions. The urban heat island tends to be most pronounced on calm, clear nights after sunny days but, for many towns and cities, is apparent even in yearly averages. Other climate variables, such as humidity and wind speed, are also modified to such an extent by some urban areas that differences are apparent in the long-term statistics for urban and adjacent rural locations.

THE SCENE IS SET

The next chapter describes the surface climatology of the British Isles and it is necessary for the reader to bear in mind that local influences, such as those we have introduced in the previous section, will be very important in modifying the climates of specific locations. Just as it is necessary to adapt a weather forecast for a region to take account of local conditions, so the broad climatology presented here must be modified to suit the reader's neighbourhood. It is possible to focus on very small scales – whole books have been written on the climates of a single city[15] – but the aim of this chapter, and of much of this book, has been to look outwards from the climate of the British Isles to the large-scale processes which deliver our average weather. When one considers the myriad factors which shape the climate of a particular area, it is clear why modelling weather and the climate system presents such a challenge. It is necessary to take account of all the processes discussed in this chapter, and many more, on spatial scales ranging from the global to the local. Computer power is limited, so compromises have to be made, often in sacrificing local detail. It is testament to the skills of those developing the models and forecasts that, despite the difficulties, they manage such a degree of accuracy.

NOTES

1 The reader wishing to learn more of the global climate system and the general circulation of the atmosphere is directed to R.G. Barry and R.J. Chorley, *Atmosphere, Weather and Climate* (6th edn) London, Routledge, 1995. For more technical accounts, see R. McIlveen, *Fundamentals of Weather and Climate*, London, Chapman and Hall, 1992, 497 pp., or J.T. Houghton, *The Physics of Atmospheres* (2nd edn), Cambridge, Cambridge University Press, 1986, 271 pp.

2 This radiation is longer in wavelength than that emitted by the Sun since the temperature of the Earth s atmosphere is much lower than that of the Sun – hotter bodies emit shorter wavelength radiation

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cycle has a characteristic time-scale of several weeks or so, but the period is very variable.

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13 See R. McIlveen, op. cit., for interesting descriptions of polar lows and heat lows.

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