9 Weather and climate

'There is really no such thing as bad weather, only different types of good weather.'

John Ruskin, Quote from Lord Avebury

'When two Englishmen meet, their first talk is of the weather.'

Samuel Johnson, *The Idler*

The science of meteorology is the study of atmospheric phenomena; it includes the study of both weather and climate. The distinction between climate and weather is one of scale. **Weather** refers to the state of the atmosphere at a local level, usually on a short timescale of minutes to months. It emphasises aspects of the atmosphere that affect human activity, such as sunshine, cloud, wind, rainfall, humidity and temperature. **Climate** is concerned with the long-term behaviour of the atmosphere in a specific area. Climatic characteristics are represented by data on temperature, pressure, wind, precipitation, humidity, etc. which are used to calculate daily, monthly and yearly averages (Framework 8, page 246) and to build up global patterns (Chapter 12).

Structure and composition of the atmosphere

The atmosphere is an envelope of transparent, odourless gases held to the Earth by gravitational attraction. While the furthest limit of the atmosphere is said by international convention to be at 1000 km, most of the atmosphere, and therefore our climate and weather, is concentrated within 16 km of the Earth's surface at the Equator and 8 km at the poles. Fifty per cent of atmospheric mass is within 5.6 km of sea-level and 99 per cent is within 40 km. Atmospheric pressure decreases rapidly with height but, as recordings made by radiosondes, weather balloons and more recently weather satellites have shown, temperature changes are more complex. Changes in temperature mean that the atmosphere can be conveniently divided into four distinctive layers (Figure 9.1); moving outwards from the Earth's surface:

1 **Troposphere** Temperatures in the troposphere decrease by 6.4°C with every 1000 m increase in altitude (environmental lapse rate, page 216). This is because the Earth's surface is warmed by incoming solar radiation which in turn heats the air next to it by conduction, convection and radiation. Pressure falls as the effect of gravity decreases, although wind speeds usually increase with height. The layer is unstable and contains most of the atmosphere's water vapour, cloud, dust and pollution. The tropopause, which forms the upper limit to the Earth's climate and weather, is marked by an isothermal layer where temperatures remain constant despite any increase in height.

2 **Stratosphere** The stratosphere is characterised by a steady increase in temperature (temperature inversion, page 217) caused by a concentration of **ozone** (O_3) (Places 27, page 209). This gas absorbs incoming **ultra-violet (UV) radiation** from the sun. Winds are light in the lower parts, but increase with height; pressure continues to fall and the air is dry. The stratosphere, like the two layers above it, acts as a protective shield against meteorites which usually burn out as they enter the Earth's gravitational field. The **stratopause** is another isothermal layer where temperatures do not change with increasing height.

3 **Mesosphere** Temperatures fall rapidly as there is no water vapour, cloud, dust or ozone to absorb incoming radiation. This layer experiences the atmosphere's lowest temperatures (–90°C) and strongest winds (nearly 3000 km/hr). The **mesopause**, like the tropopause and stratopause, shows no change in temperature.

4 **Thermosphere** Temperatures rise rapidly with height, perhaps to reach 1500°C. This is due to an increasing proportion of atomic oxygen in the atmosphere which, like ozone, absorbs incoming ultra-violet radiation.

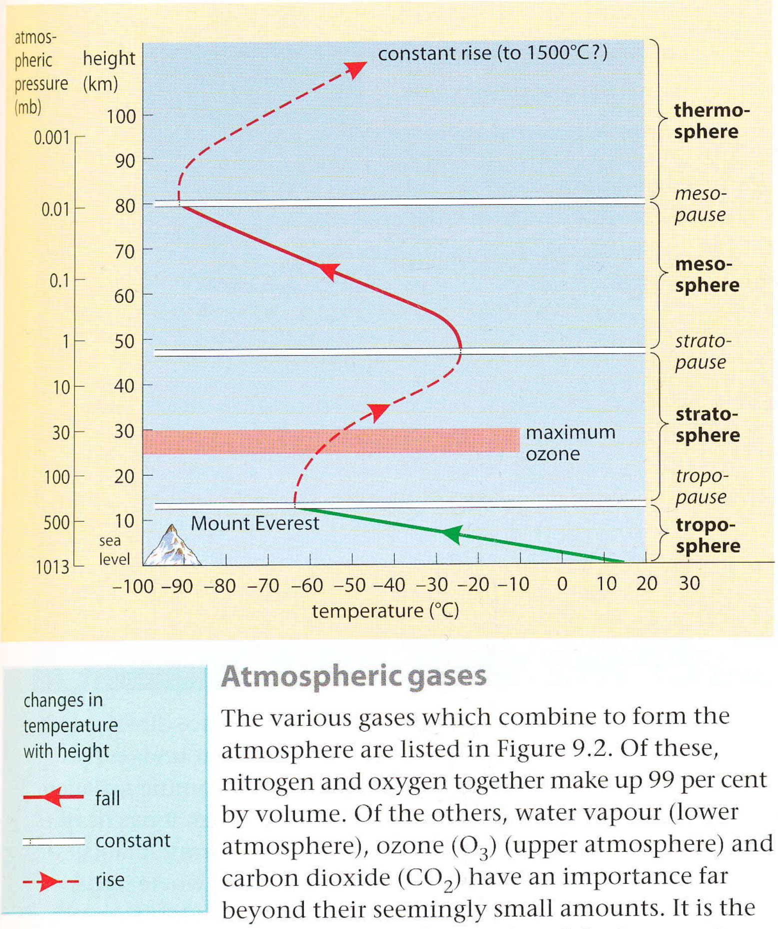


Figure 9.1

The vertical structure of the atmosphere

Atmospheric gases

The various gases which combine to form the atmosphere are listed in Figure 9.2. Of these, nitrogen and oxygen together make up 99 per cent by volume. Of the others, water vapour (lower atmosphere), ozone (O_3) (upper atmosphere) and carbon dioxide (CO_2) have an importance far beyond their seemingly small amounts. It is the depletion of O_3 (Places 27) and the increase in CO_2 (Case Study 9B) which are causing concern to scientists.

Energy in the atmosphere

The sun is the Earth's prime source of energy. The Earth receives energy as incoming **short-wave** solar radiation (also referred to as **insolation**). It is this energy that controls our planet's climate and weather and which, when converted by photosynthesis in green plants, supports all forms of life. The amount of incoming radiation received by the Earth is determined by four astronomical factors (Figure 9.3): the solar constant, the distance from the sun, the altitude of the sun in the sky, and the length of night and day. Figure 9.3 is theoretical in that it assumes there is no atmosphere around the Earth. In reality, much insolation is absorbed, reflected and scattered as it passes through the atmosphere (Figure 9.4).

Absorption of incoming radiation is mainly by ozone, water vapour, carbon dioxide and particles of ice and dust. It occurs in, and is limited to, the infra-red part of the spectrum. Clouds and, to a lesser extent, the Earth's surface **reflect** considerable amounts of radiation back into space. The ratio between incoming radiation and the amount reflected, expressed as a percentage, is known as the **albedo**. The albedo varies with cloud type from 30–40 per cent in thin clouds, to 50–70 per cent in thicker stratus and 90 per cent in cumulo-nimbus (when only 10 per cent reaches the atmosphere below cloud level). Albedos also vary over different land surfaces, from less than 10 per cent over

Figure 9.2

The composition of the atmosphere

Gas		Percentage by volume	Importance for weather and climate	Other functions/source
Permanent gases:	nitrogen	78.09	Mainly passive	Needed for plant growth.
	oxygen	20.95		Produced by photosynthesis; reduced by deforestation.
Variable gases:	water vapour	0.20–4.0	Source of cloud formation and precipitation, reflects/absorbs incoming long-wave radiation. Keeps global temperatures constant. Provides majority of natural 'greenhouse effect'.	Essential for life on Earth. Can be stored as ice/snow.
	carbon dioxide	0.03	Absorbs long-wave radiation from Earth and so contributes to 'greenhouse effect'. Its increase due to human activity is a major cause of global warming.	Used by plants for photosynthesis; increased by burning fossil fuels and by deforestation.
	ozone	0.00006	Absorbs incoming short-wave ultra-violet radiation.	Reduced/destroyed by chlorofluorocarbons (CFCs).
	pollutants	trace	Sulphur dioxide, nitrogen oxide, methane. Absorb long-wave radiation, cause acid rain and contribute to the greenhouse effect.	From industry, power stations and car exhausts.
Inert gases:	argon	0.93		
	helium, neon, krypton	trace		
Non-gaseous:	dust	trace	Absorbs/reflects incoming radiation. Forms condensation nuclei necessary for cloud formation.	Volcanic dust, meteoritic dust, soil erosion by wind.

Note: the figures refer to dry air and so the variable amount of water vapour is not usually taken into consideration.

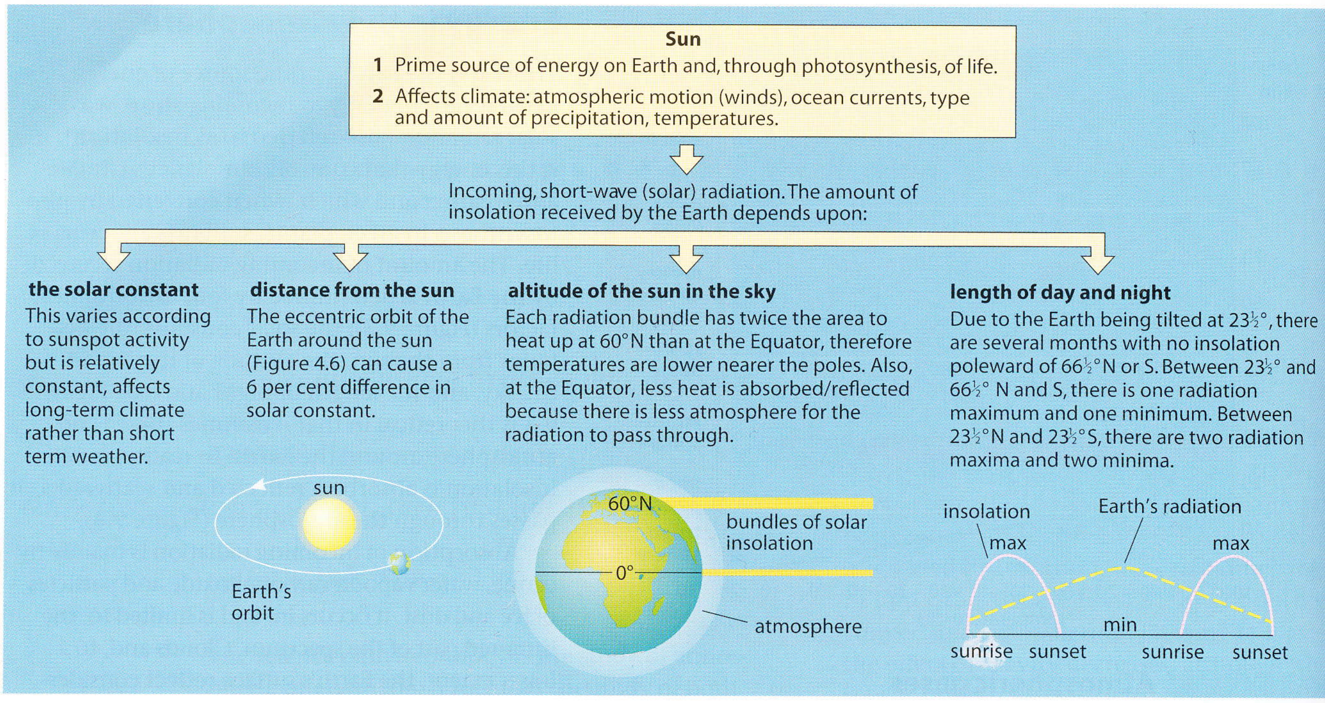


Figure 9.3

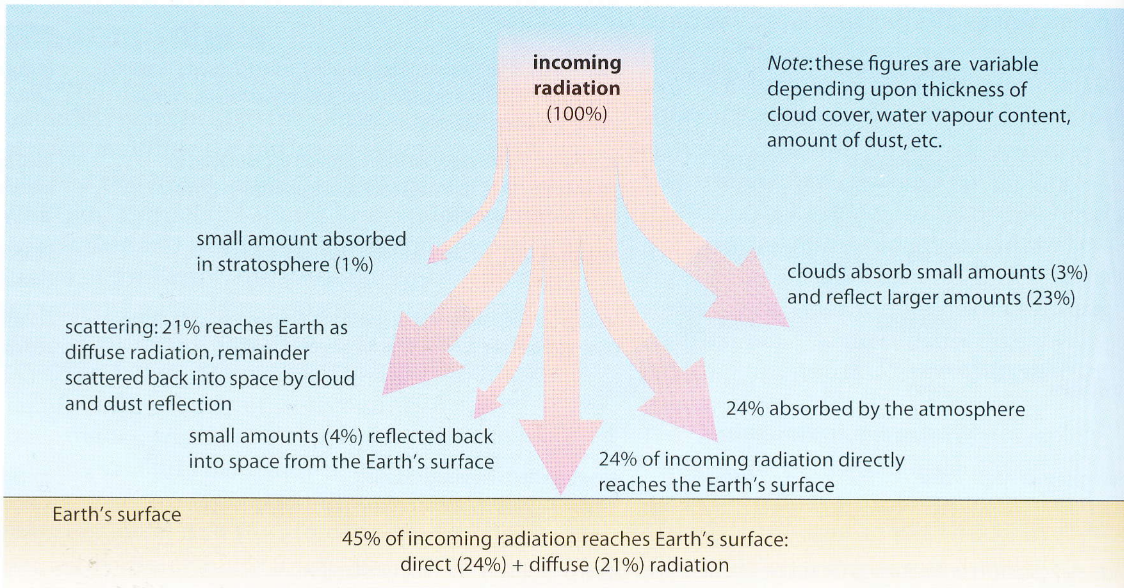
Incoming radiation received by the Earth (assuming that there is no atmosphere)

oceans and dark soil, to 15 per cent over coniferous forest and urban areas, 25 per cent over grasslands and deciduous forest, 40 per cent over light-coloured deserts and 85 per cent over reflecting fresh snow. Where deforestation and overgrazing occur, the albedo increases. This reduces the possibility of cloud formation and precipitation and increases the risk of **desertification** (Case Study 7). **Scattering** occurs when incoming radiation is diverted by particles of dust, as from volcanoes and deserts, or by molecules of gas. It takes place in all directions and some of the radiation will reach the Earth's surface as **diffuse** radiation.

As a result of absorption, reflection and scattering, only about 24 per cent of incoming radiation reaches the Earth's surface directly, wit a further 21 per cent arriving at ground-level as diffuse radiation (Figure 9.4). Incoming radiatior is converted into heat energy when it reaches the Earth's surface. As the ground warms, it radiates energy back into the atmosphere where 94 per cent is absorbed (only 6 per cent is lost to space), mainly by water vapour and carbon dioxide – the greenhouse effect (Case Study 9B). Without the natural greenhouse effect, which traps so much o the outgoing radiation, world temperatures woul be 33°C lower than they are at present and life or Earth would be impossible. (During the ice age, it was only 4°C cooler.) This outgoing (terrestrial) radiation is **long-wave** or **infra-red** radiation.

Figure 9.4

The solar energy cascade



Places 27 The atmosphere: ozone

The major concentration of ozone is in the stratosphere, 25–30 km above sea-level (Figure 9.1). Ozone acts as a shield protecting the Earth from the damaging effects of ultra-violet (UV) radiation from the sun. An increase in UV radiation means an increase in sunburn and skin cancer (fair skin is at greater risk than dark skin), snow-blindness, cataracts and eye damage, ageing and skin wrinkling in humans, as well as having a major impact on Antarctic organisms.

A depletion in ozone above the Antarctic was first observed, by chance, by the British Antarctic Survey in 1977, and the first 'hole' was described in a scientific paper published in 1985. The term 'hole' is misleading as it means a depletion in ozone of over 50 per cent (not a 100 per cent loss). Each Antarctic spring (September to November) the temperature falls so low that it causes ozone to be destroyed in a chemical reaction with chlorine. At the time there were two main sources of chlorine:

- the release of chlorofluorocarbons (CFCs) from aerosols such as hairsprays, deodorants, refrigerator coolants and manufacturing processes that produced foam packaging (a long-term effect)
- from major volcanic eruptions, e.g. Mount Pinatubo (Case Study 1 – a short-term effect).

The 1985 paper was followed by a spate of experiments aimed at trying to establish the causes and probable effects of ozone depletion. Within two years – a remarkably short time for international action – the Montreal Protocol was signed by which the more industrialised countries agreed to set much lower limits for CFC production, and subsequently to reduce this to zero. The agreement came so quickly, and CFC production dropped so rapidly, that the Montreal Protocol has been held up as a 'model' international environmental agreement.

Initially, ozone depletion continued. The first Arctic 'hole' was observed in 1989 following the coldest-ever recorded January in that region. The 'hole' over Antarctica continued to grow each year until 2003, by which time it had reached its maximum extent and was affecting populated parts of Chile and New Zealand. Since then, mainly due to most of the harmful CFCs having been replaced by gases less toxic to ozone (though still greenhouse gases), there have been encouraging signs of ozone replacement and hopes are high that ozone concentrations will return to normal by the middle or latter part of this century – a rare success story for international environment management.

In contrast, vehicle exhaust systems generate dangerous quantities of ozone close to the Earth's surface, especially during calm summer anticyclonic conditions (page 234). Under extreme conditions, nitrogen oxide from exhausts reacts with VOCs (volatile organic compounds) in sunlight to create a petrochemical smog. This can cause serious damage to the health of people (especially those with asthma) and animals.

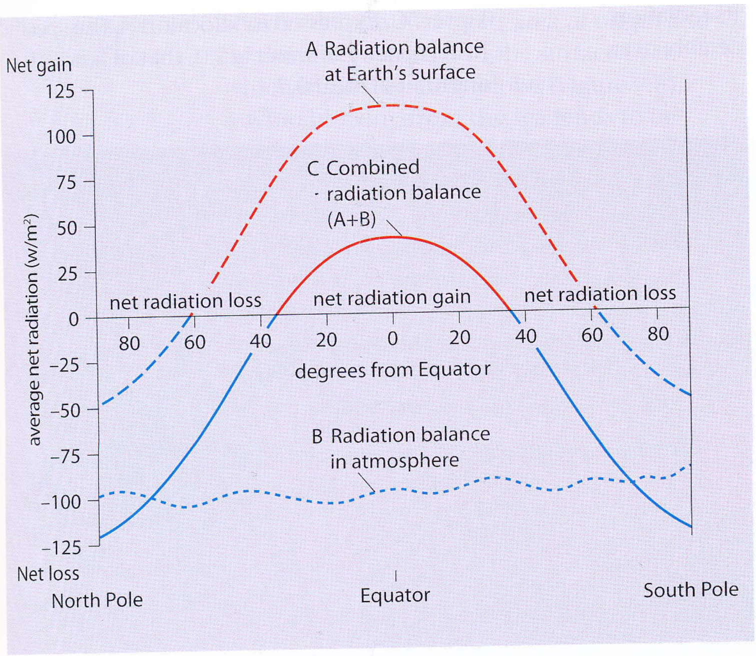


Figure 9.5

The heat budget

The heat budget

Since the Earth is neither warming up nor cooling down, there must be a balance between incoming insolation and outgoing terrestrial radiation. Figure 9.5 shows that:

- there is a net gain in radiation everywhere on the Earth's surface (curve A) except in polar latitudes which have high albedo surfaces
- there is a net loss in radiation throughout the atmosphere (curve B)
- after balancing the incoming and outgoing radiation, there is a net surplus between 35°S and 40°N (the difference in latitude is due to the larger land masses of the northern hemisphere) and a net deficit to the poleward sides of those latitudes (curve C).

This means that there is a **positive heat balance** within the tropics and a **negative heat balance** both at high latitudes (polar regions) and high altitudes. Two major **transfers** of heat, therefore, take place to prevent tropical areas from over-heating (Figure 9.6).

1 **Horizontal heat transfers** Heat is transferred away from the tropics, thus preventing the Equator from becoming increasingly hotter and the poles increasingly colder. Winds (air movements including jet streams, page 227; hurricanes, page 235; and depressions, page 230) are responsible for 80 per cent of this heat transfer, and ocean currents for 20 per cent (page 211).

2 **Vertical heat transfers** Heat is also transferred vertically, thus preventing the Earth's surface from getting hotter and the atmosphere colder. This is achieved through **radiation**, **conduction**, **convection** and the transfer of **latent heat**. Latent heat is the amount of heat energy needed to change the state of a substance without affecting its temperature. When ice changes into water or water into vapour, heat is taken up to help with the processes of melting and evaporation. This absorption of heat results in the cooling of the atmosphere. When the process is reversed – i.e. vapour condenses into water or water freezes into ice – heat energy is released and the atmosphere is warmed.

Variations in the radiation balance occur at a number of spatial and temporal scales. Regional differences may be due to the uneven distribution of land and sea, altitude, and the direction of prevailing winds. Local variations may result from **aspect** and amounts of cloud cover. Seasonal and diurnal variations are related to the altitude of the sun and the length of night and day.

Global factors affecting insolation

Factors that influence the amount of insolation received at any point, and therefore its radiation balance and heat budget, vary considerably over time and space.

Long-term factors

These are relatively constant at a given point.

- **Height above sea-level** The atmosphere is not warmed directly by the sun, but by heat radiated from the Earth's surface and distributed by conduction and convection. As the height of mountains increases, they present a decreasing area of land surface from which to heat the surrounding air. In addition, as the density or pressure of the air decreases, so too does its ability to hold heat (Figure 9.1). This is because the molecules in the air which receive and retain heat become fewer and more widely spaced as height increases.
- **Altitude of the sun** As the angle of the sun in the sky decreases, the land area heated by a given ray and the depth of atmosphere through which that ray has to pass both increase. Consequently, the amount of insolation lost through absorption, scattering and reflection also increases. Places in lower latitudes therefore have higher temperatures than those in higher latitudes.
- **Land and sea** Land and sea differ in their ability to absorb, transfer and radiate heat energy. The sea is more transparent than the land, and is capable of absorbing heat down to a depth of 10 metres. It can then transfer this heat to greater depths through the movements of waves and currents. The sea also has a greater **specific heat capacity** than that of land. Specific heat capacity is the amount of energy required to raise the temperature of 1 kg of a substance by 1°C, expressed in kilojoules per kg per °C. Expressed in kilocalories, the specific heat capacity of water is 1.0, that of land is 0.5 and that of sand 0.2.

Figure 9.6

Heat transfers in the atmosphere

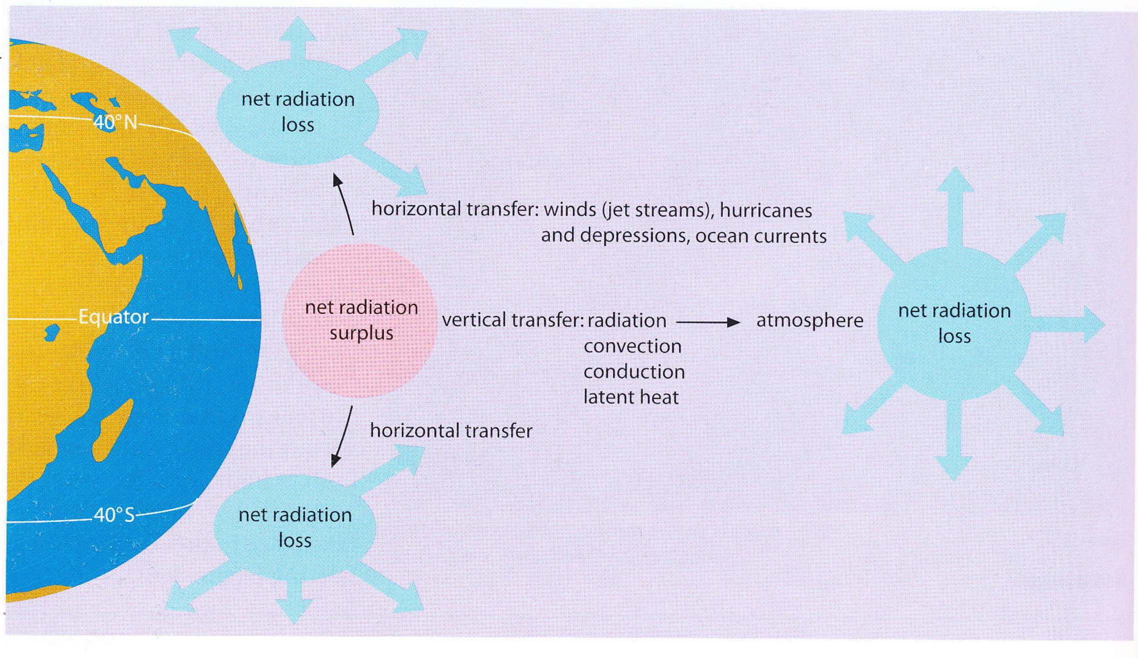
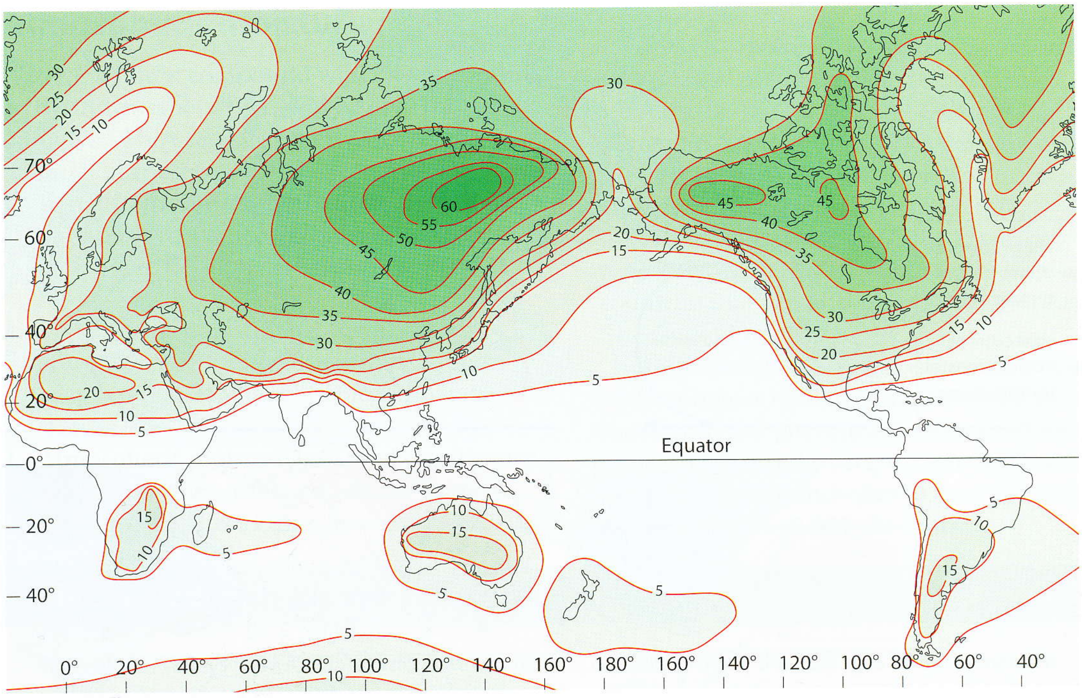


Figure 9.7

Mean annual ranges in global temperature (°C)



This means that water requires twice as much energy as soil and five times more than sand to raise an equivalent mass to the same temperature. During summer, therefore, the sea heats up more slowly than the land. In winter, the reverse is the case and land surfaces lose heat energy more rapidly than water. The oceans act as efficient 'thermal reservoirs'. This explains why coastal environments have a smaller annual range of temperature than locations at the centres of continents (Figure 9.7).

- **Prevailing winds** The temperature of the wind is determined by its area of origin and by the characteristics of the surface over which it subsequently blows (Figure 9.8). A wind blowing from the sea tends to be warmer in winter and cooler in summer than a corresponding wind coming from the land.
- **Ocean currents** These are a major component in the process of horizontal transfer of heat energy. Warm currents carry water polewards and raise the air temperature of the maritime environments where they flow. Cold currents carry water towards the Equator and so lower the temperatures of coastal areas (Figure 9.9). The main ocean currents follow circular routes – clockwise in the northern hemisphere, anti-clockwise in the southern hemisphere.

Figure 9.10 shows the difference between the mean January temperature of a place and the mean January temperatures of other places with the same latitude; this difference is known as a **temperature anomaly**. (The term 'temperature anomaly' is used specifically to describe temperature differences from a mean. It should not be confused with the more general definition of 'anomaly' which refers to something that does not fit into a general pattern.) For example, Stornoway (Figure 9.10) has a mean January temperature of 4°C, which is 20°C higher than the average for other locations lying at 58°N. Such anomalies result primarily from the uneven heating and cooling rates of land and sea and are intensified by the horizontal transfer of energy by ocean currents and prevailing winds. Remember that the sun appears overhead in the southern hemisphere at this time of year (January) and isotherms have been reduced to sea-level – i.e. temperatures are adjusted to eliminate some of the effects of relief, thus emphasising the influence of prevailing winds, ocean currents and continentality.

Figure 9.8

Simplified diagram showing the effect of prevailing winds on land and sea temperatures

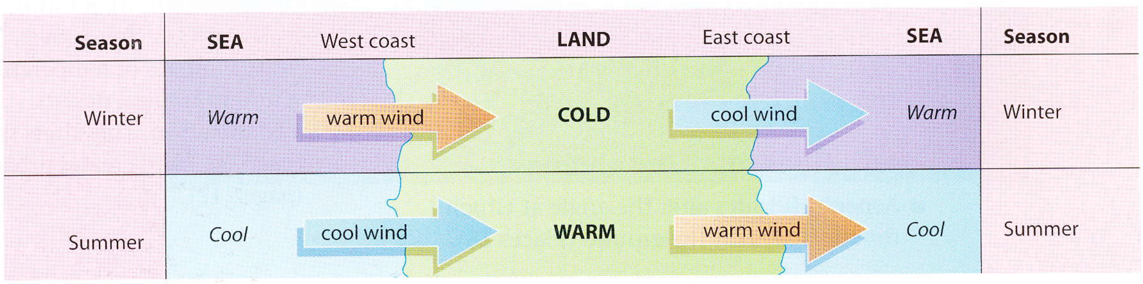


Figure 9.9

Major ocean currents

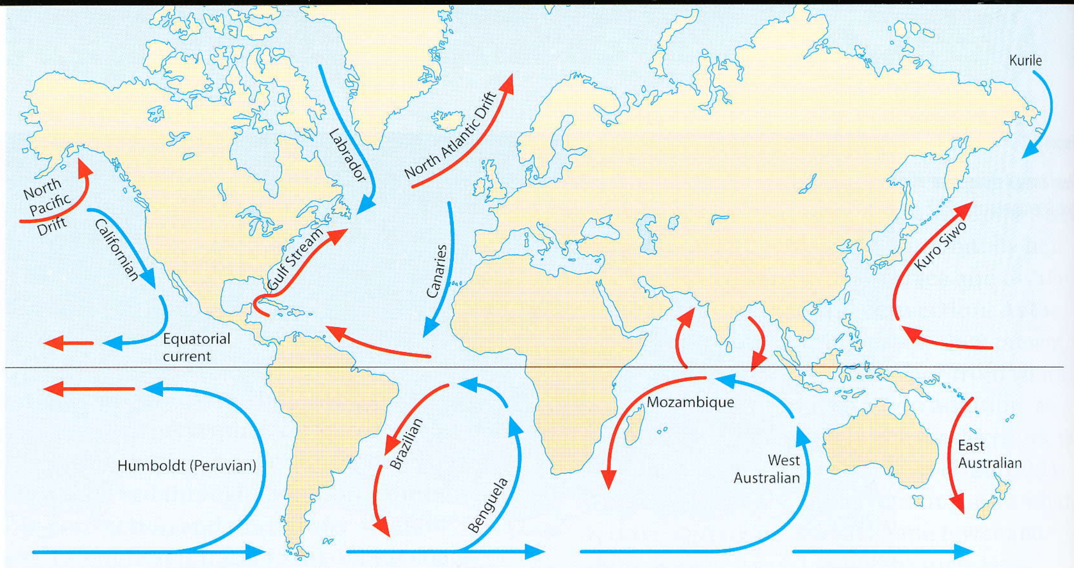
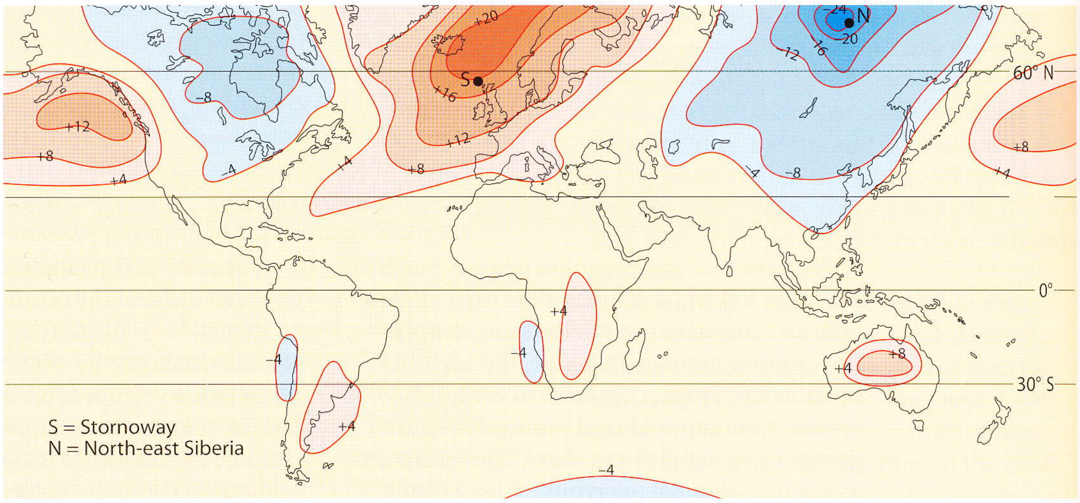


Figure 9.10

Temperature anomalies for January (*after* D.C. Money)



Short-term factors

- **Seasonal changes** At the spring and autumn equinoxes (21 March and 22 September) when the sun is directly over the Equator, insolation is distributed equally between both hemispheres. At the summer and winter solstices (21 June and 22 December) when, due to the Earth's tilt, the sun is overhead at the tropics, the hemisphere experiencing 'summer' will receive maximum insolation.
- **Length of day and night** Insolation is only received during daylight hours and reaches its peak at noon. There are no seasonal variations at the Equator, where day and night are of equal length throughout the year. In extreme contrast, polar areas receive no insolation during part of the winter when there is continuous darkness, but may receive up to 24 hours of insolation during part of the summer when the sun never sinks below the horizon ('the lands of the midnight sun').

Local influences on insolation

- **Aspect** Hillsides alter the angle at which the sun's rays hit the ground (Places 28). In the northern hemisphere, north-facing slopes, being in shadow for most or all of the year, are cooler than those facing south. The steeper the south-facing slope, the higher the angle of the sun's rays to it and therefore the higher will be the temperature. North- and south-facing slopes are referred to, respectively, as the **adret** and **ubac**.
- **Cloud cover** The presence of cloud reduces both incoming and outgoing radiation. The thicker the cloud, the greater the amount of absorption, reflection and scattering of insolation, and of terrestrial radiation. Clouds may reduce daytime temperatures, but they also act as an insulating blanket to retain heat at night. This means that tropical deserts, where skies are clear, are warmer during the day and cooler at night than humid equatorial regions with a greater cloud cover. The world's greatest diurnal ranges of temperature are therefore found in tropical deserts.
- **Urbanisation** This alters the albedo (page 207) and creates urban 'heat islands' (page 242).

Atmospheric moisture

Water is a liquid compound which is converted by heat into vapour (gas) and by cold into a solid (ice). The presence of water serves three essential purposes:

1 It maintains life on Earth: flora, in the form of natural vegetation (biomes) and crops; and fauna, i.e. all living creatures, including humans.
2 Water in the atmosphere, mainly as a gas, absorbs, reflects and scatters insolation to keep our planet at a habitable temperature (Figure 9.4).
3 Atmospheric moisture is of vital significance as a means of transferring surplus energy from tropical areas either horizontally to polar latitudes or vertically into the atmosphere to balance the heat budget (Figure 9.5).

Despite this need for water, its existence in a form readily available to plants, animals and humans is limited. It has been estimated that 97.2 per cent of the world's water is in the oceans and seas; in this form, it is only useful to plants tolerant of saline conditions (**halophytes**, page 291) and to the populations of a few wealthy countries that can afford desalinisation plants (the Gulf oil states).

Approximately 2.1 per cent of water in the hydrosphere is held in storage as polar ice and snow. Only 0.7 per cent is fresh water found either in lakes and rivers (0.1 per cent), as soil moisture and groundwater (0.6 per cent), or in the atmosphere (0.001 per cent).

Places 28 An alpine valley: aspect

Many alpine valleys in Switzerland and Austria have an east–west orientation which means that their valley sides face either north or south. South-facing adret slopes are much warmer and drier than those facing north (Figure 9.11). The south-facing slopes have more plant species, a higher tree-line, and a greater land use with alpine pastures at higher altitudes and fruit and hay lower down; also, they usually provide the best sites for settlement. In contrast, north-facing ubac slopes are snow-covered for a much longer period, they are less suited to farming, the tree-line is lower, and they tend to be left forested. However, on the valley floors, as severe frosts are likely to occur during times of temperature inversion (page 217), sensitive plants and crops do not flourish.

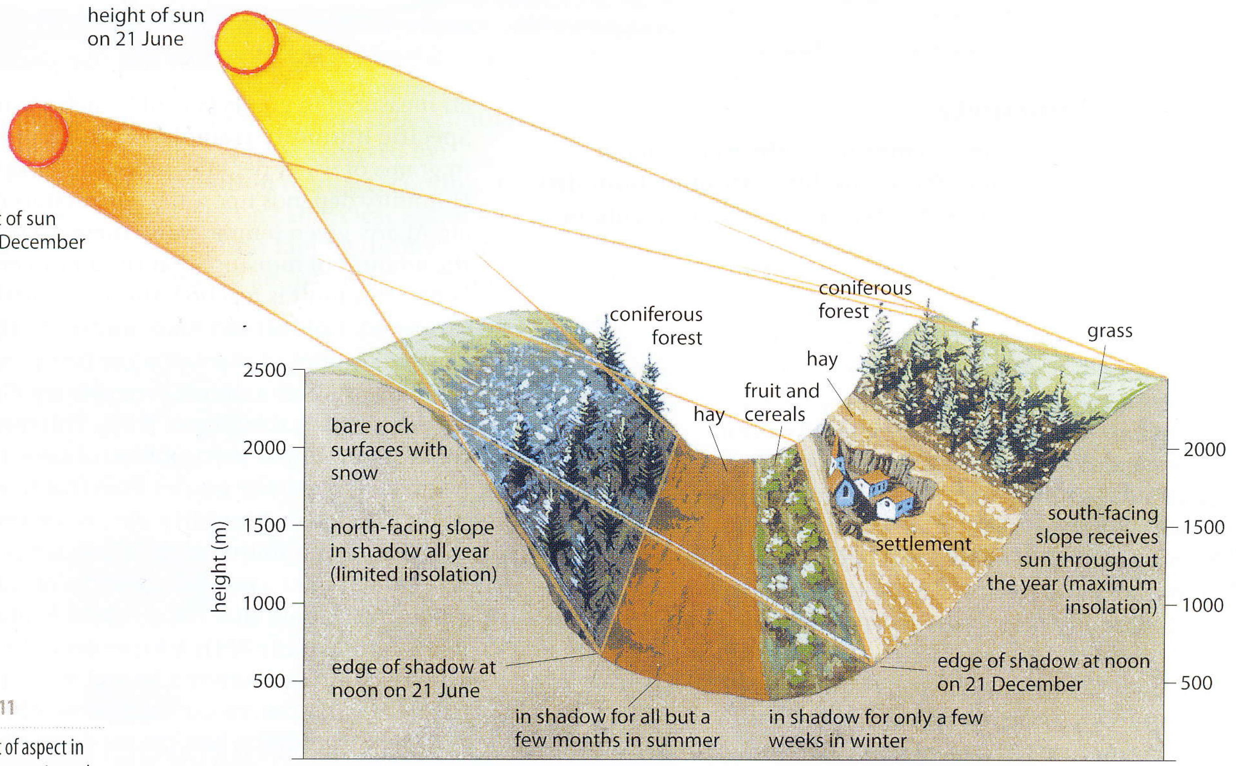


Figure 9.11

The effect of aspect in an east–west oriented alpine valley in the northern hemisphere

Atmospheric moisture

Water is a liquid compound which is converted by heat into vapour (gas) and by cold into a solid (ice). The presence of water serves three essential purposes:

1. It maintains life on Earth: flora, in the form of natural vegetation (biomes) and crops; and fauna, i.e. all living creatures, including humans.
2. Water in the atmosphere, mainly as a gas, absorbs, reflects and scatters insolation to keep our planet at a habitable temperature (Figure 9.4).
3. Atmospheric moisture is of vital significance as a means of transferring surplus energy from tropical areas either horizontally to polar latitudes or vertically into the atmosphere to balance the heat budget (Figure 9.5).

Despite this need for water, its existence in a form readily available to plants, animals and humans is limited. It has been estimated that 97.2 per cent of the world's water is in the oceans and seas; in this form, it is only useful to plants tolerant of saline conditions (**halophytes**, page 291) and to the populations of a few wealthy countries that can afford desalinisation plants (the Gulf oil states).

Approximately 2.1 per cent of water in the hydrosphere is held in storage as polar ice and snow. Only 0.7 per cent is fresh water found either in lakes and rivers (0.1 per cent), as soil moisture and groundwater (0.6 per cent), or in the atmosphere (0.001 per cent).

Places 28 An alpine valley: aspect

Many alpine valleys in Switzerland and Austria have an east–west orientation which means that their valley sides face either north or south. South-facing adret slopes are much warmer and drier than those facing north (Figure 9.11). The south-facing slopes have more plant species, a higher tree-line, and a greater land use with alpine pastures at higher altitudes and fruit and hay lower down; also, they usually provide the best sites for settlement. In contrast, north-facing ubac slopes are snow-covered for a much longer period, they are less suited to farming, the tree-line is lower, and they tend to be left forested. However, on the valley floors, as severe frosts are likely to occur during times of temperature inversion (page 217), sensitive plants and crops do not flourish.

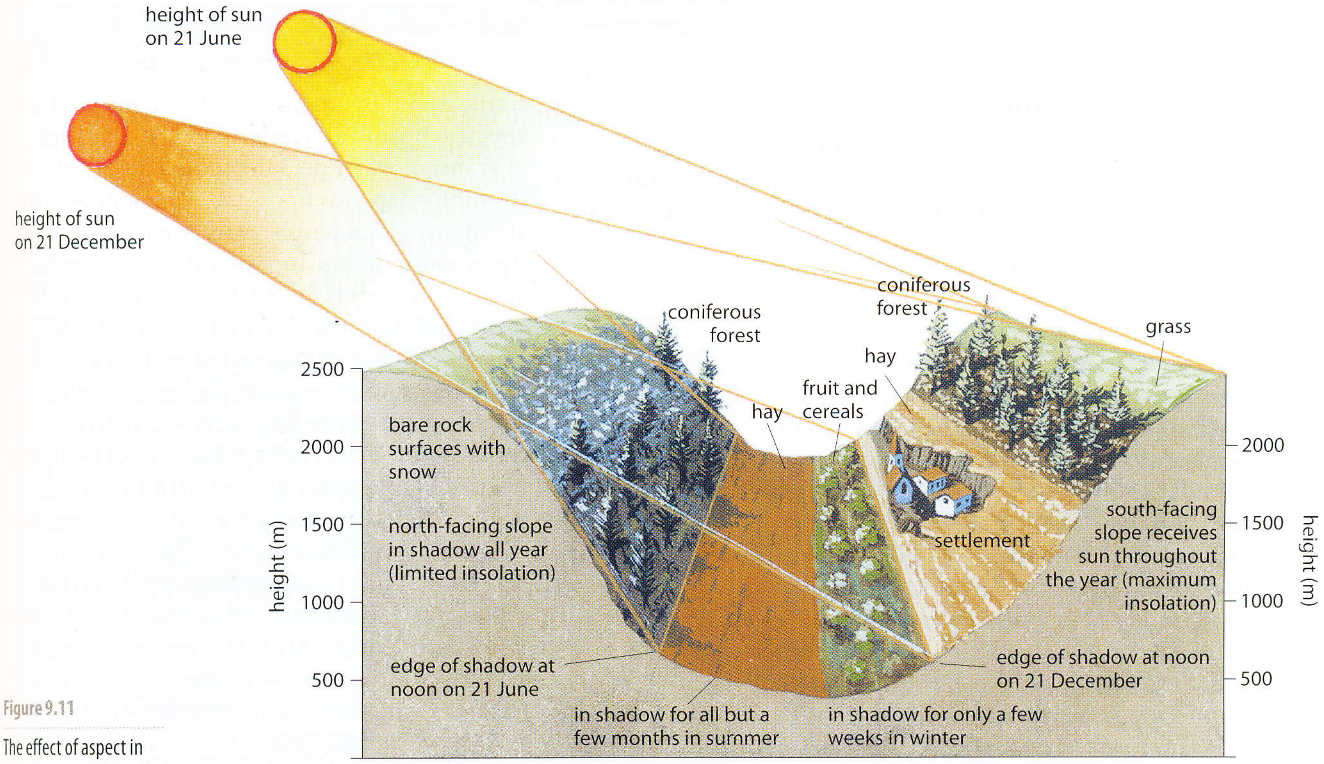
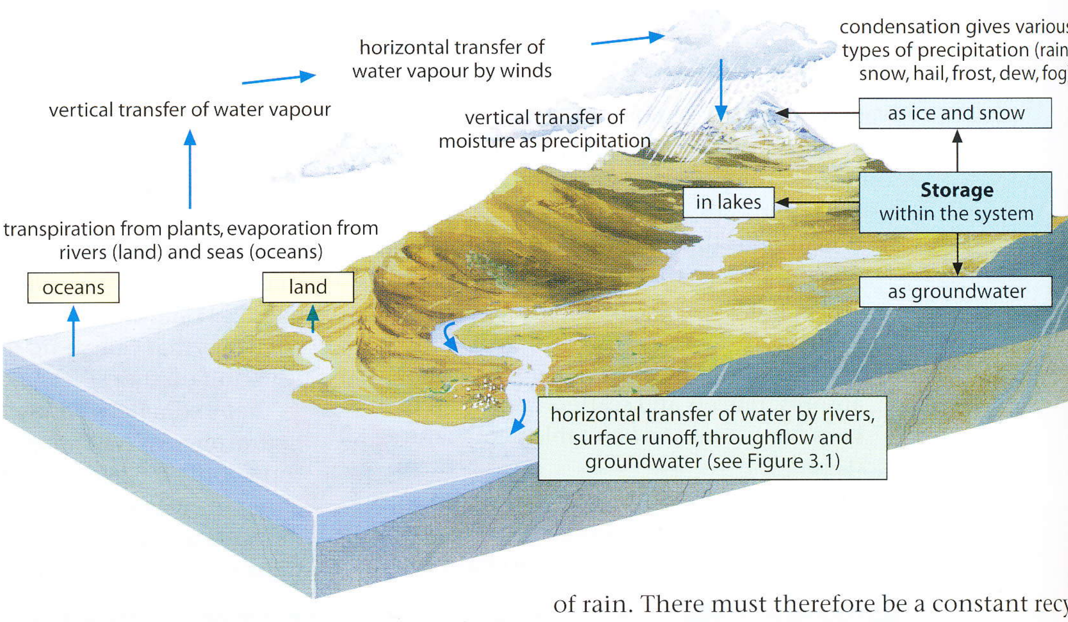


Figure 9.11

The effect of aspect in an east–west oriented alpine valley in the northern hemisphere

Figure 9.12

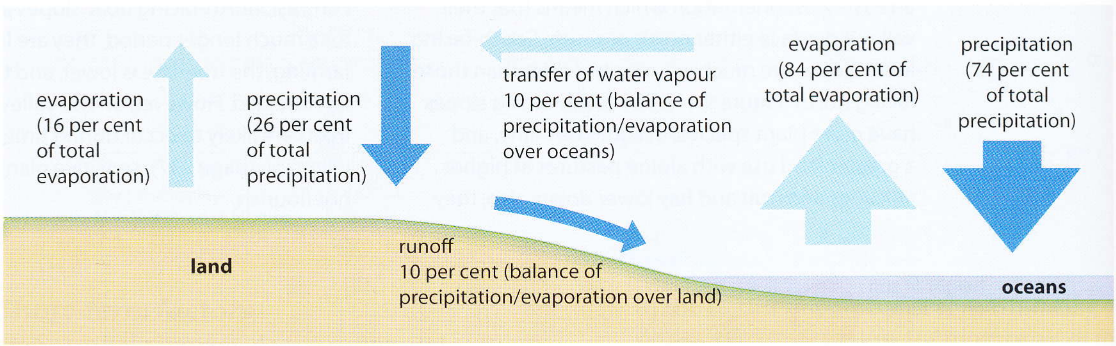
The hydrological cycle (compare with Figure 3.1)



At any given time, the atmosphere only holds, on average, sufficient moisture to give every place on the Earth 2.5 cm (about 10 days' supply) of rain. There must therefore be a constant recycling of water between the oceans, atmosphere and land (Figure 9.13). This recycling is achieved through the **hydrological cycle** (Figure 9.12).

Figure 9.13

The world's water balance

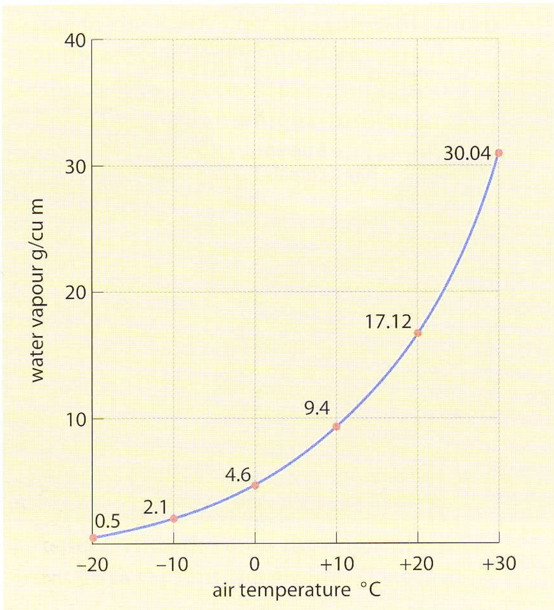


Humidity

Humidity is a measure of the water vapour content in the atmosphere. **Absolute humidity** is the mass of water vapour in a given volume of air measured in grams per cubic metre (g/m^3). **Specific humidity** is similar but is expressed in grams of water per kilogram of air (g/kg). Humidity depends upon the temperature of the air. At any given temperature, there is a limit to the amount of moisture that the air can hold. When this limit is reached, the air is said to be **saturated**. Cold air can hold only relatively small quantities of vapour before becoming saturated but this amount increases rapidly as temperatures rise (Figure 9.14). This means that the amount of precipitation obtained from warm air is generally greater than that from cold air. **Relative humidity** (RH) is the amount of water vapour in the air at a given temperature expressed as a percentage of the maximum amount of vapour that the air could hold at that temperature. If the RH is 100 per cent, the air is saturated. If it lies between 80 and 99 per cent, the air is said to be 'moist' and the weather is humid or clammy. When the RH drops to 50 per cent, the air is 'dry'– figures as low as 10 per cent have been recorded over hot deserts.

Figure 9.14

Air temperatures and absolute humidity for saturated air



If unsaturated air is cooled and atmospheric pressure remains constant, a critical temperature will be reached when the air becomes saturated (i.e. RH = 100 per cent). This is known as the **dew point**. Any further cooling will result in the condensation of excess vapour, either into water droplets where condensation nuclei are present, or into ice crystals if the air temperature is below 0°C. This is shown in the following worked example.

1 The early morning air temperature was 10°C. Although the air could have held 100 units of water at that temperature, at the time of the reading it held only 90. This meant that the RH was 90 per cent.
2 During the day, the air temperature rose to 12°C. As the air warmed it became capable of holding more water vapour, up to 120 units. Owing to evaporation, the reading reached a maximum of 108 units which meant that the RH remained at 90 per cent – i.e. (108 ÷ 120) × 100.
3 In the early evening, the temperature fell to 10°C at which point, as stated above, it could hold only 100 units. However, the air at that time contained 108 units so, as the temperature fell, dew point was reached and the 8 excess units of water were lost through condensation.

Condensation

This is the process by which water vapour in the atmosphere is changed into a liquid or, if the temperature is below 0°C, a solid. It usually results from air being cooled until it is saturated. Cooling may be achieved by:

1 **Radiation** (contact) **cooling** This typically occurs on calm, clear evenings. The ground loses heat rapidly through terrestrial radiation and the air in contact with it is then cooled by conduction. If the air is moist, some vapour will condense to form radiation fog, dew, or – if the temperature is below freezing point – hoar frost (page 221).
2 **Advection cooling** This results from warm, moist air moving over a cooler land or sea surface. Advection fogs in California and the Atacama Desert (Places 24, page 180 and page 122) are formed when warm air from the land drifts over cold offshore ocean currents (Figure 9.9).

As both radiation and advection involve horizontal rather than vertical movements of air, the amount of condensation created is limited.

3 **Orographic** and **frontal uplift** Warm, moist air is forced to rise either as it crosses a mountain barrier (orographic ascent, page 220) or when it meets a colder, denser mass of air at a front (page 229).
4 **Convective** or **adiabatic cooling** This is when air is warmed during the daytime and rises in pockets as **thermals** (Figure 9.15). As the air expands, it uses energy and so loses heat and the temperature drops. Because air is cooled by the reduction of pressure with height rather than by a loss of heat to the surrounding air, it is said to be adiabatically cooled (see lapse rates, page 216).

As both orographic and adiabatic cooling involve vertical movements of air, they are more effective mechanisms of condensation.

Condensation does not occur readily in clean air. Indeed, if air is absolutely pure, it can be cooled below its dew point to become **supersaturated** with an RH in excess of 100 per cent. Laboratory tests have shown that clean, saturated air can be cooled to –40°C before condensation or, in this case, **sublimation**. Sublimation is when vapour condenses directly into ice crystals without passing through the liquid state. However, air is rarely pure and usually contains large numbers of condensation nuclei. These microscopic particles, referred to as **hygroscopic nuclei** because they attract water, include volcanic dust (heavy rain always accompanies volcanic eruptions); dust from windblown soil; smoke and sulphuric acid originating from urban and industrial areas; and salt from sea spray. Hygroscopic nuclei are most numerous over cities, where there may be up to 1 million per cm^3, and least common over oceans (only 10 per cm^3). Where large concentrations are found, condensation can occur with an RH as low as 75 per cent – as in the smogs of Los Angeles (Figure 9.25 and Case Study 15A).

Figure 9.15

Convective cooling

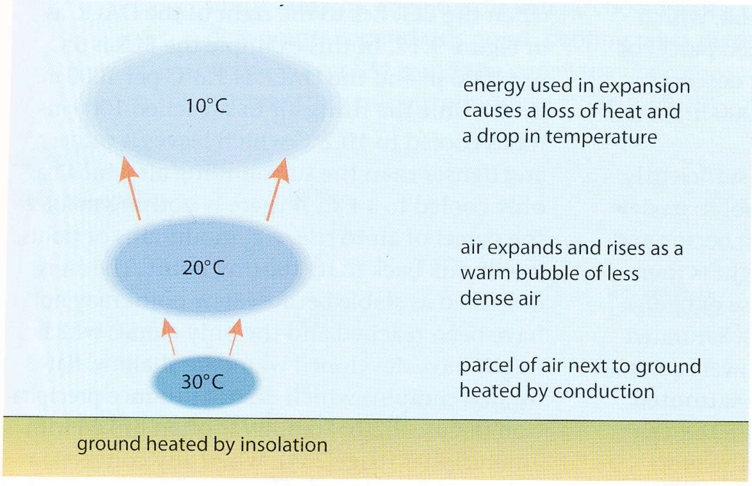
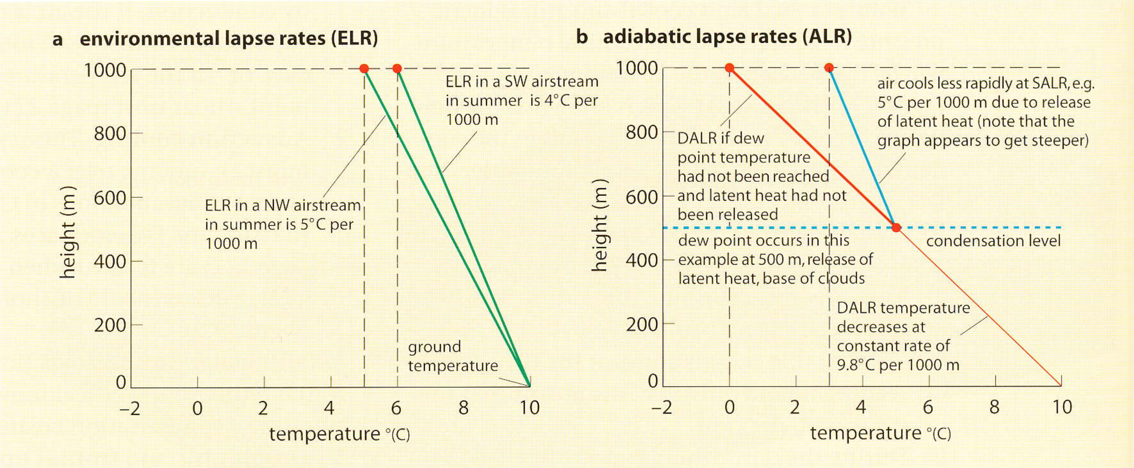


Figure 9.16

Examples of lapse rates shown in temperature–height diagrams (tephigrams)



Lapse rates

The **environmental lapse rate** (ELR) is the decrease in temperature usually expected with an increase in height through the troposphere (Figure 9.1). The ELR is approximately 6.5°C per 1000 m, but varies according to local air conditions. It may vary due to several factors: **height** – ELR is lower nearer ground-level; **time** – it is lower in winter or during a rainy season; over different **surfaces** – it is lower over continental areas; and between different **air masses** (Figure 9.16a).

The **adiabatic lapse rate** (ALR) describes what happens when a parcel of air rises and the decrease in pressure is accompanied by an associated increase in volume and a decrease in temperature (Figure 9.15). Conversely, descending air will be subject to an increase in pressure causing a rise in temperature. In either case, there is negligible mixing with the surrounding air. There are two adiabatic lapse rates:

1 If the upward movement of air does not lead to condensation, the energy used by expansion will cause the temperature of the parcel of air to fall at the **dry adiabatic lapse rate** (DALR on Figure 9.16b). The DALR, which is the rate at which an unsaturated parcel of air cools as it rises or warms as it descends, remains constant at 9.8°C per 1000 m (i.e. approximately 1°C per 100 m).

2 When the upward movement is sufficiently prolonged to enable the air to cool to its dew point temperature, condensation occurs and the loss in temperature with height is then partly compensated by the release of latent heat (Figure 9.16b and page 210). Saturated air, which therefore cools at a slower rate than unsaturated air, loses heat at the **saturated adiabatic lapse rate** (SALR). The SALR can vary because the warmer the air the more moisture it can hold, and so the greater the amount of latent heat released following condensation. The SALR may be as low as 4°C per 1000 m and as high as 9°C per 1000 m. It averages about 5.4°C per 1000 m (i.e. approximately 0.5°C per 100 m). Should temperatures fall below 0°C, then the air will cool at the **freezing adiabatic lapse rate** (FALR). This is the same as the DALR as very little moisture is present at low temperatures.

Air stability and instability

Parcels of warm air which rise through the lower atmosphere cool adiabatically. The rate and maintenance of any vertical uplift depend on the temperature–density balance between the rising parcel and the surrounding air. In a simplified form, this balance is the relationship between the environmental lapse rate and the dry and saturated adiabatic lapse rates.

Stability

The state of **stability** is when a rising parcel of unsaturated air cools more rapidly than the air surrounding it. This is shown diagrammatically when the ELR lies to the right of the DALR, as in Figure 9.17. In this example the ELR is 6°C per 1000 m and the DALR is 9.8°C per 1000 m. By the time the rising air has reached 1000 m, it has cooled to 10.2°C which leaves it colder and denser than the surrounding air which has only cooled to 14°C. If there is nothing to force the parcel of air to rise, e.g. mountains or fronts, it will sink back to its starting point. The air is described as stable because dew point may not have been reached and the only clouds which might have developed would be shallow, flat-topped cumulus which do not produce precipitation (Figure 9.20). Stability is often linked with anticyclones (page 234), when any convection currents are suppressed by sinking air to give dry, sunny conditions.

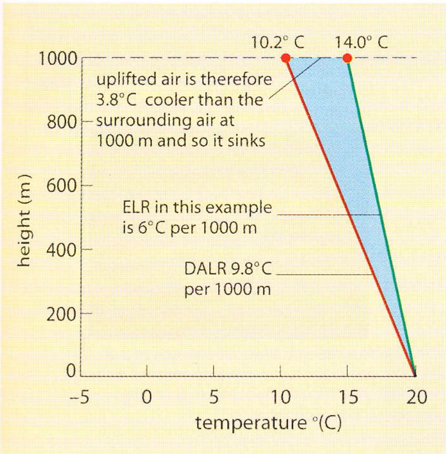


Figure 9.17

Stability: changes in lapse rates and air temperature with height

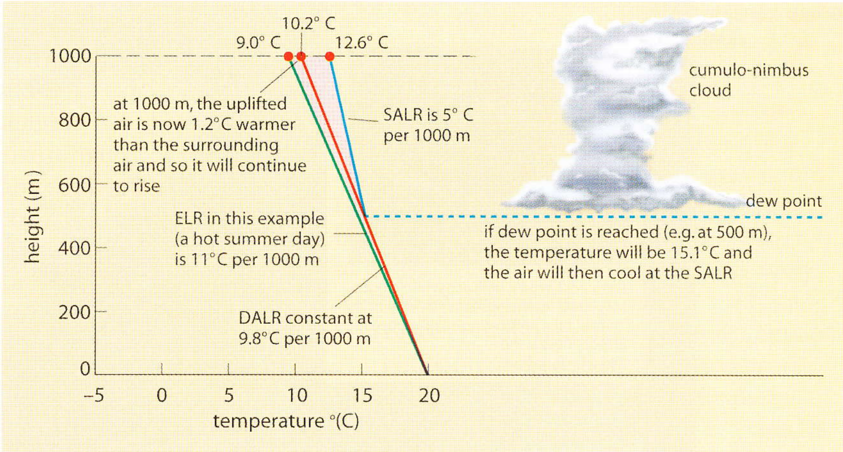


Figure 9.18

Instability and cloud development

Instability

Conditions of **instability** arise in Britain on hot days. Localised heating of the ground warms the adjacent air by conduction, creating a higher lapse rate. The resultant parcel of rising unsaturated air cools less rapidly than the surrounding air. In this case, as shown in Figure 9.18, the ELR lies to the left of the DALR. The rising air remains warmer and lighter than the surrounding air. Should it be sufficiently moist and if dew point is reached, then the upward movement may be accelerated to produce towering cumulus or cumulo-nimbus type cloud (Figure 9.20). Thunderstorms are likely (Figure 9.21) and the saturated air, following the release of latent heat, will cool at the SALR.

Conditional instability

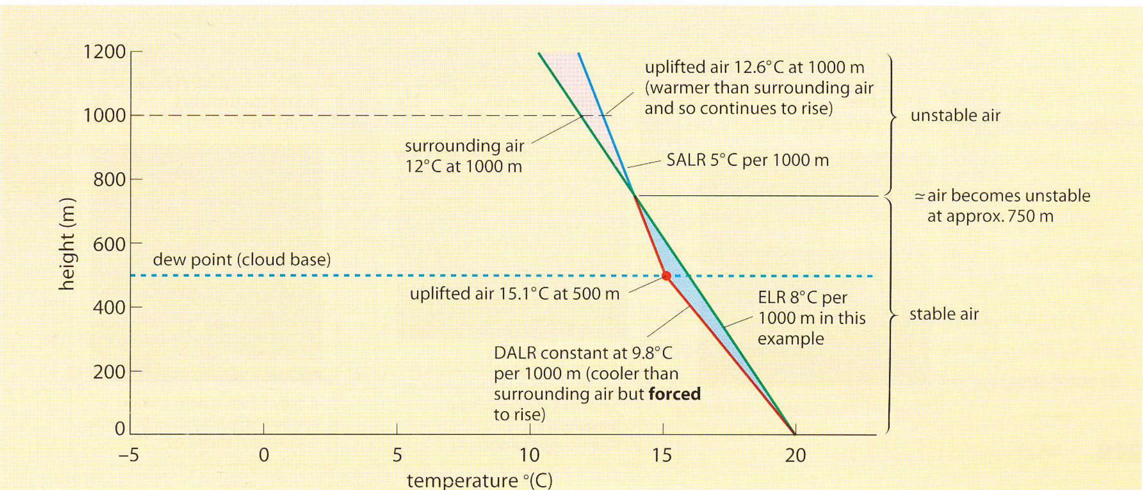
This type of instability occurs when the ELR is lower than the DALR but higher than the SALR. In Britain, it is the most common of the three conditions. The rising air is stable in its lower layers and, being cooler than the surrounding air, would normally sink back again. However, if the mechanism which initially triggered the uplift remains, then the air will be cooled to its dew point. Beyond this point, cooling takes place at the slower SALR and the parcel may become warmer than the surrounding air (Figure 9.19). It will now continue to rise freely, even if the uplifting mechanism is removed, as it is now in an unstable state. Instability is conditional upon the air being forced to rise in the first place, and later becoming saturated so that condensation occurs. The associated weather is usually fine in areas at altitudes below condensation level, but cloudy and showery in those above.

Temperature inversions

As the lapse rate exercises have shown, the temperature of the air usually decreases with altitude, but there are certain conditions when the reverse occurs. **Temperature inversions**, where warmer air overlies colder air, may occur at three levels in the atmosphere. Figure 9.1 showed that temperatures increase with altitude in both the stratosphere and the thermosphere. Inversions can also occur near ground-level and high in the troposphere. High-level inversions are found in depressions where warm air overrides cold air at the warm front or is undercut by colder air at the cold front (page 229). Low-level, or ground, inversions usually occur under anticyclonic conditions (page 234 and Figure 9.24) when there is a rapid loss of heat from the ground due to radiation at night, or when warm air is advected over a cold surface. Under these conditions, fog and frost (page 221 and Figure 9.23) may form in valleys and hollows.

Figure 9.19

Conditional instability



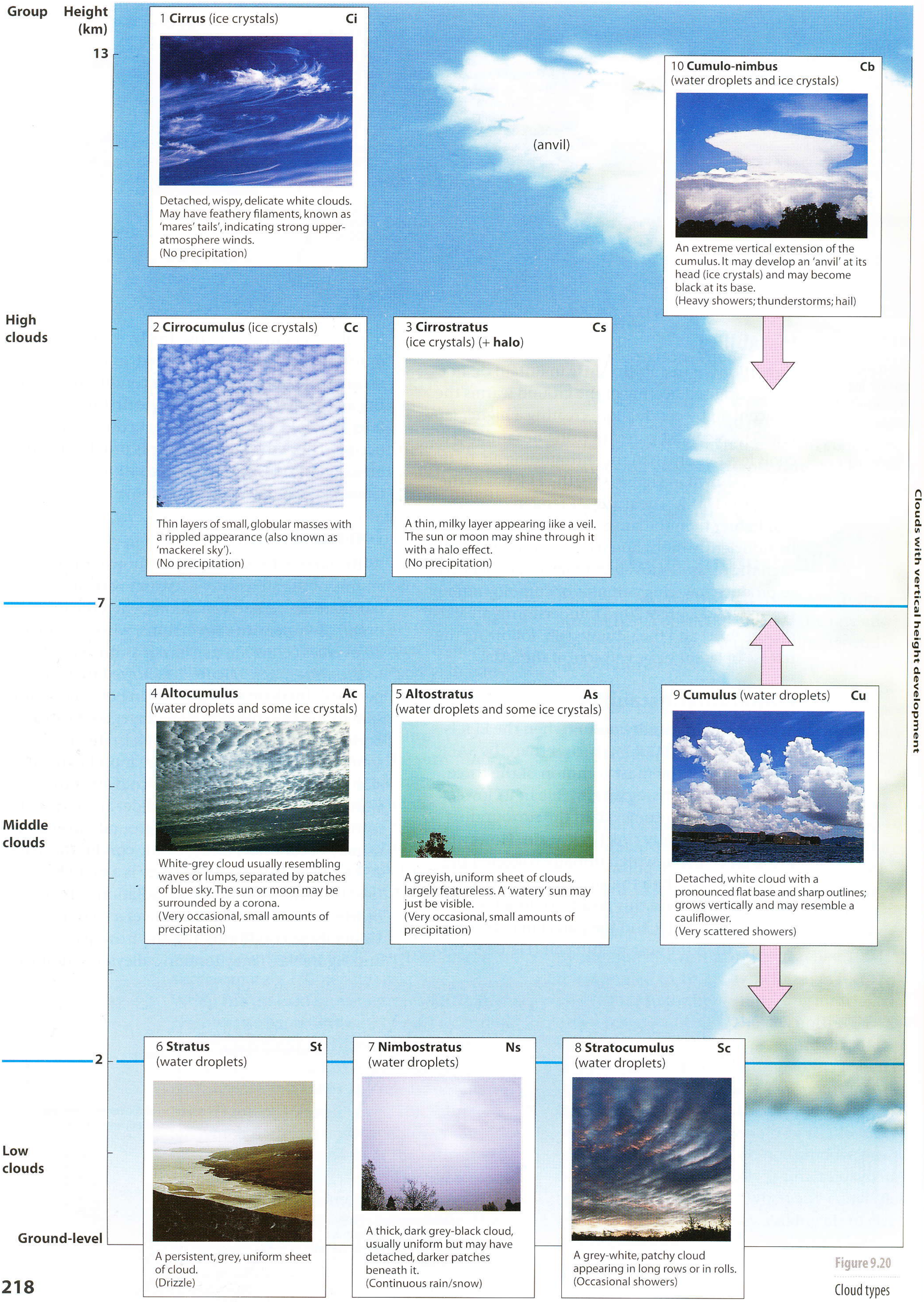


Figure 9.20
Cloud types

Clouds

Clouds form when air cools to dew point and vapour condenses into water droplets and/or ice crystals. There are many different types of cloud, but they are often difficult to distinguish as their form constantly changes. The general classification of clouds was proposed by Luke Howard in 1803. His was a descriptive classification, based on cloud shape and height (Figure 9.20). He used four Latin words: **cirrus** (a lock of curly hair); **cumulus** (a heap or pile); **stratus** (a layer); and **nimbus** (rain-bearing). He also compiled composite names using these four terms, such as cumulo-nimbus, cirrostratus; and added the prefix 'alto-' for middle-level clouds.

Precipitation

Condensation produces minute water droplets, less than 0.05 mm in diameter, or, if the dew point temperature is below freezing, ice crystals. The droplets are so tiny and weigh so little that they are kept buoyant by the rising air currents which created them. So although condensation forms clouds, clouds do not necessarily produce precipitation. As rising air currents are often strong, there has to be a process within the clouds which enables the small water droplets and/or ice crystals to become sufficiently large to overcome the uplifting mechanism and fall to the ground.

There are currently two main theories that attempt to explain the rapid growth of water droplets:

1. The **ice crystal mechanism** is often referred to as the Bergeron–Findeisen mechanism. It appears that when the temperature of air is between –5°C and –25°C, supercooled water droplets and ice crystals exist together. Supercooling takes place when water remains in the atmosphere after temperatures have fallen below 0°C – usually due to a lack of condensation nuclei. Ice crystals are in a minority because the **freezing nuclei** necessary for their formation are less abundant than condensation nuclei. The relative humidity of air is ten times greater above an ice surface than over water. This means that the water droplets evaporate and the resultant vapour condenses (sublimates) back onto the ice crystals which then grow into hexagonal-shaped snowflakes. The flakes grow in size – either as a result of further condensation or by fusion as their numerous edges interlock on collision with other flakes. They also increase in number as ice splinters break off and form new nuclei. If the air temperature rises above freezing point as the snow falls to the ground, flakes melt into raindrops. Experiments to produce rainfall artificially by cloud-seeding are based upon this process. The Bergeron–Findeisen theory is supported by evidence from temperate latitudes where rainclouds usually extend vertically above the freezing level. Radar and high-flying aircraft have reported snow at high altitudes when it is raining at sea-level. However, as clouds rarely reach freezing point in the tropics, the formation of ice crystals is unlikely in those latitudes.
2. The **collision and coalescence** process was suggested by Longmuir. 'Warm' clouds (i.e. those containing no ice crystals), as found in the tropics, contain numerous water droplets of differing sizes. Different-sized droplets are swept upwards at different velocities and, in doing so, collide with other droplets. It is thought that the larger the droplet, the greater the chance of collision and subsequent coalescence with smaller droplets. When coalescing droplets reach a radius of 3 mm, their motion causes them to disintegrate to form a fresh supply of droplets. The thicker the cloud (cumulo-nimbus), the greater the time the droplets have in which to grow and the faster they will fall, usually as thundery showers.

Latest opinions suggest that these two theories may complement each other, but that a major process of raindrop enlargement has yet to be understood.

Types of precipitation

Although the definition of precipitation includes sleet, hail, dew, hoar frost, fog and rime, only rain and snow provide significant totals in the hydrological cycle.

Rainfall

There are three main types of rainfall, distinguished by the mechanisms which cause the initial uplift of the air. Each mechanism rarely operates in isolation.

1. **Convergent** and **cyclonic** (**frontal**) rainfall results from the meeting of two air streams in areas of low pressure. Within the tropics, the trade winds, blowing towards the Equator, meet at the inter-tropical convergence zone or ITCZ (page 226). The air is forced to rise and, in conjunction with convection currents, produces the heavy afternoon thunderstorms associated with the equatorial climate (page 316). In temperate latitudes, depressions form at the boundary of two air masses. At the associated fronts, warm, moist, less dense air is forced to rise over colder, denser air, giving periods of prolonged and sometimes intense rainfall. This is often augmented by orographic precipitation.

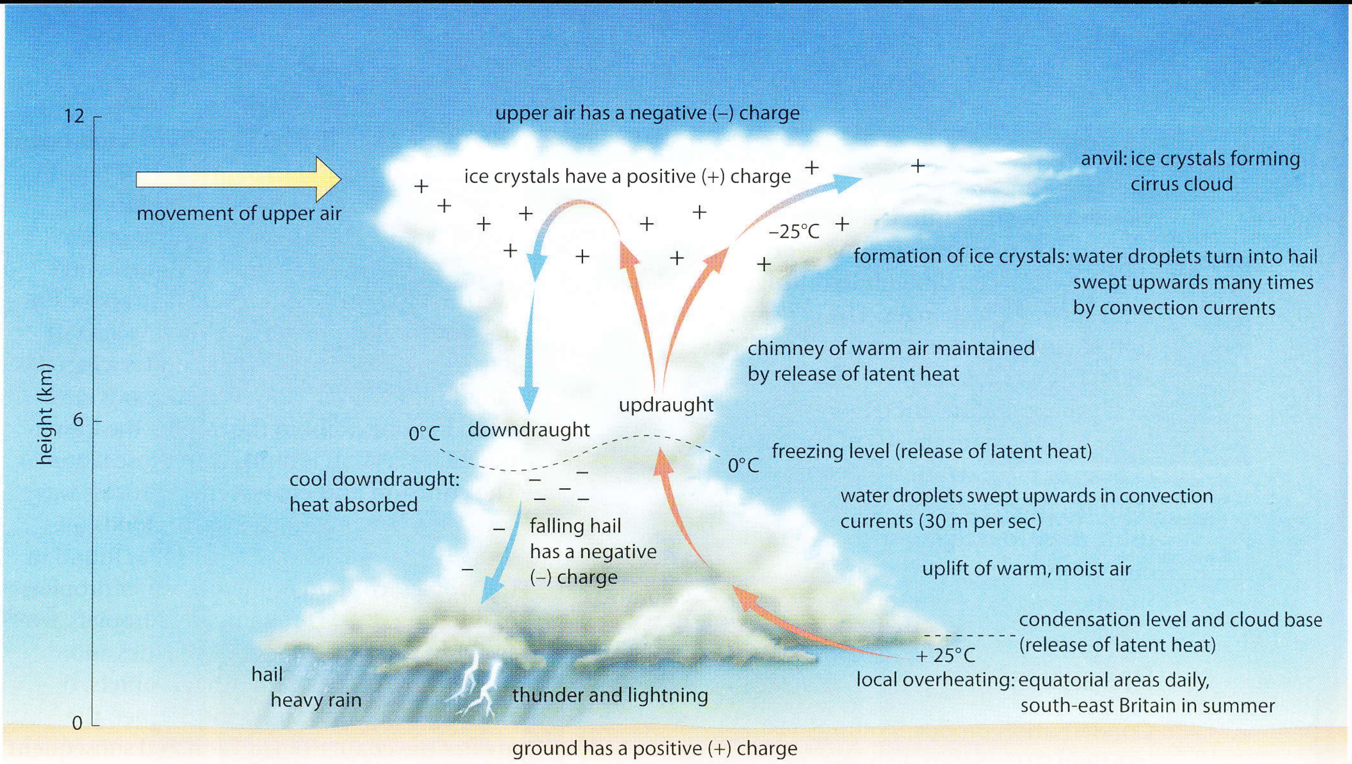


Figure 9.21

Convectional rainfall: the development of a thunderstorm

2 **Orographic** or **relief** rainfall results when near-saturated, warm maritime air is forced to rise where confronted by a coastal mountain barrier. Mountains reduce the water-holding capacity of rising air by enforced cooling and can increase the amounts of cyclonic rainfall by retarding the speed of depression movement. Mountains also tend to cause air streams to converge and funnel through valleys. Rainfall totals increase where mountains are parallel to the coast, as is the Canadian Coast Range, and where winds have crossed warm offshore ocean currents, as they do before reaching the British Isles. As air descends on the leeward side of a mountain range, it becomes compressed and warmed and condensation ceases, creating a **rainshadow** effect where little rain falls.

3 **Convectional** rainfall occurs when the ground surface is locally overheated and the adjacent air, heated by conduction, expands and rises. During its ascent, the air mass remains warmer than the surrounding environmental air and it is likely to become unstable (page 217) with towering cumulonimbus clouds forming. These unstable conditions, possibly augmented by frontal or orographic uplift, force the air to rise in a 'chimney' (Figure 9.21). The updraught is maintained by energy released as latent heat at both condensation and freezing levels. The cloud summit is characterised by ice crystals in an anvil shape, the top of the cloud being flattened by upper-air movements. When the ice crystals and frozen water droplets, i.e. hail, become large enough, they fall in a downdraught. The air through which they fall remains cool as heat is absorbed by evaporation. The downdraught reduces the warm air supply to the 'chimney' and therefore limits the lifespan of the storm. Such storms are usually accompanied by thunder and lightning. How storms develop immense amounts of electric charge is still not fully understood. One theory suggests that as raindrops are carried upwards into colder regions, they freeze on the outside. This ice-shell compresses the water inside it until the shell bursts and the water freezes into positively-charged ice crystals while the heavier shell fragments, which are negatively charged, fall towards the cloud base inducing a positive charge on the Earth's surface (Figure 9.21). **Lightning** is the visible discharge of electricity between clouds or between clouds and the ground. **Thunder** is the sound of the pressure wave created by the heating of air along a lightning flash. Convection is one process by which surplus heat and energy from the Earth's surface are transferred vertically to the atmosphere in order to maintain the heat balance (Figure 9.6).

Thunderstorms associated with the so-called **Spanish plume** can affect southern England several times during a hot, sultry summer. They occur when very hot air over the Sierra Nevada mountains (southern Spain) moves northwards over the Bay of Biscay where it draws in cooler, moist air. Should the resultant storm reach Britain, it can cause flash flooding, landslips and electricity blackouts.

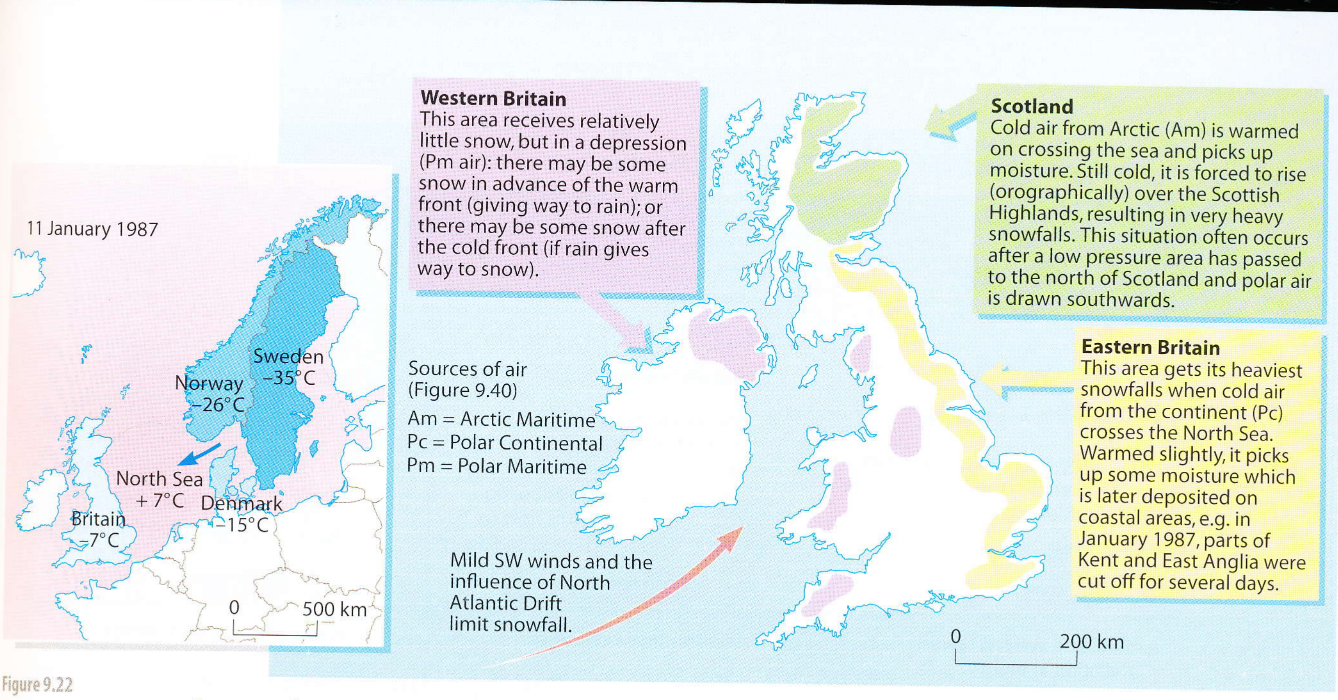


Figure 9.22

Causes of uneven snowfall patterns across Britain

Snow, sleet, glazed frost and hail

Snow forms under similar conditions to rain (Bergeron–Findeisen process) except that as dew point temperatures are under 0°C, then the vapour condenses directly into a solid (sublimation, page 215). Ice crystals will form if hygroscopic or freezing nuclei are present and these may aggregate to give snowflakes. As warm air holds more moisture than cold, snowfalls are heaviest when the air temperature is just below freezing. As temperatures drop, it becomes 'too cold for snow'. Figure 9.22 shows the typical conditions under which snow might fall in Britain.

Sleet is a mixture of ice and snow formed when the upper air temperature is below freezing, allowing snowflakes to form, and the lower air temperature is around 2 to 4°C, which allows their partial melting.

Glazed frost is the reverse of sleet and occurs when water droplets form in the upper air but turn to ice on contact with a freezing surface. When glazed frost forms on roads, it is known as 'black ice'.

Hail is made up of frozen raindrops which exceed 5 mm in diameter. It usually forms in cumulo-nimbus clouds, resulting from the uplift of air by convection currents, or at a cold front. It is more common in areas with warm summers where there is sufficient heat to trigger the uplift of air, and less common in colder climates. Hail frequently proves a serious climatic hazard in cereal-growing areas such as the American Prairies.

Figure 9.23

Temperature inversion: radiation fog in a valley, Iceland

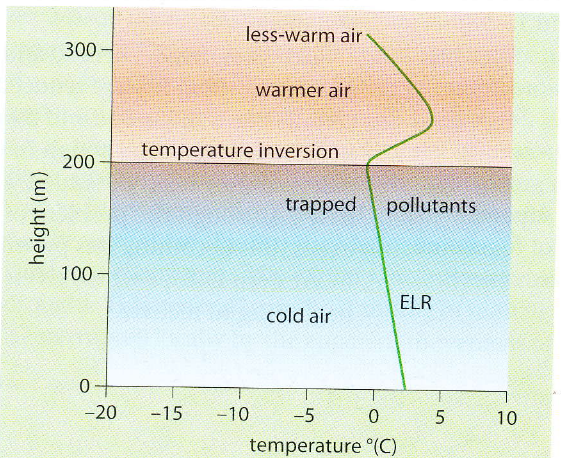


Figure 9.24

A low-level temperature inversion

Dew, hoar frost, fog and rime

Dew, hoar frost and radiation fog all form under calm, clear, anticyclonic conditions when there is rapid terrestrial radiation at night. Dew point is reached as the air cools by conduction and moisture in the air, or transpired from plants, condenses. If dew point is above freezing, **dew** will form; if it is below freezing, **hoar frost** develops. Frost may also be frozen dew. Dew and hoar frost usually occur within 1 m of ground-level.

If the lower air is relatively warm, moist and contains hygroscopic nuclei, and if the ground cools rapidly, **radiation fog** may form. Where visibility is more than 1 km it is mist, if less than 1 km, fog. In order for radiation fog to develop, a gentle wind is needed to stir the cold air adjacent to the ground so that cooling affects a greater

Figure 9.25

Formation of radiation fog and smog

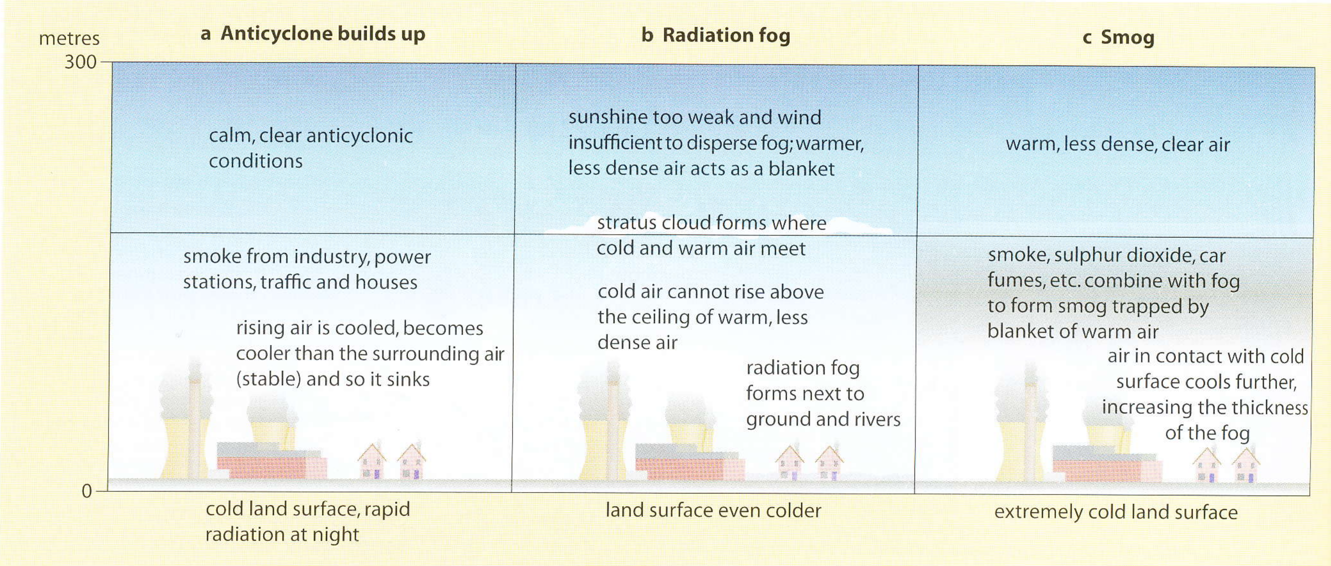


Figure 9.26

Rime frost, North Carolina, USA

thickness of air. Radiation fogs usually occur in valleys, are densest around sunrise, and consist of droplets which are sufficiently small to remain buoyant in the air. Fog is likely to thicken if temperature inversion takes place (Figures 9.23 and 9.24), i.e. when cold surface air is trapped by overlying warmer, less dense air. It is under such conditions, in urban and industrial areas, that smoke and other pollutants released into the air are retained as smog (Figures 9.25 and 15.55).

Advection fog forms when warm air passes over or meets with cold air to give rapid cooling. In the coastal Atacama Desert (Places 24, page 180), sufficient droplets fall to the ground as 'fog-drip' to enable some vegetation growth.

Rime (Figure 9.26) occurs when supercooled droplets of water, often in the form of fog, come into contact with, and freeze on, solid objects such as telegraph poles and trees.

Acid rain

This is an umbrella term for the presence in rainfall of a series of pollutants which are produced mainly by the burning of fossil fuels. Coal-fired power stations, heavy industry and vehicle exhausts emit sulphur dioxide and nitrogen oxides. These are carried by prevailing winds across seas and national frontiers to be deposited either directly onto the Earth's surface as dry deposition or to be converted into acids (sulphuric and nitric acid) which fall to the ground in rain as wet deposition. Clean rainwater has a pH value of 5.6, which is slightly acidic due to the natural presence of carbonic acid (dissolved carbon dioxide). Today, rainfall over most of north-west Europe has a pH of about 5, the lowest ever recorded being 2.2 (the same as lemon juice).

The effects of acid rain include the increase in water acidity which caused the deaths of fish and plant life, mainly in Scandinavian rivers and lakes, and the pollution of fresh water supplies. Forests can be destroyed as important soil nutrients (calcium and potassium) are washed away and replaced by manganese and aluminium, both of which are harmful to root growth. In time trees shed their needles (coniferous) and leaves (deciduous) and become less resistant to drought, frost and disease.

However, between 1980 and 2000 emissions of sulphur dioxide were reduced by nearly 60 per cent in Western Europe and by about 30 per cent in North America (although in China and South-east Asia they nearly doubled, albeit from a low base). Although the problem of acid rain still exists, it is becoming less prominent, especially in Western Europe where rivers and lakes are beginning to recover.

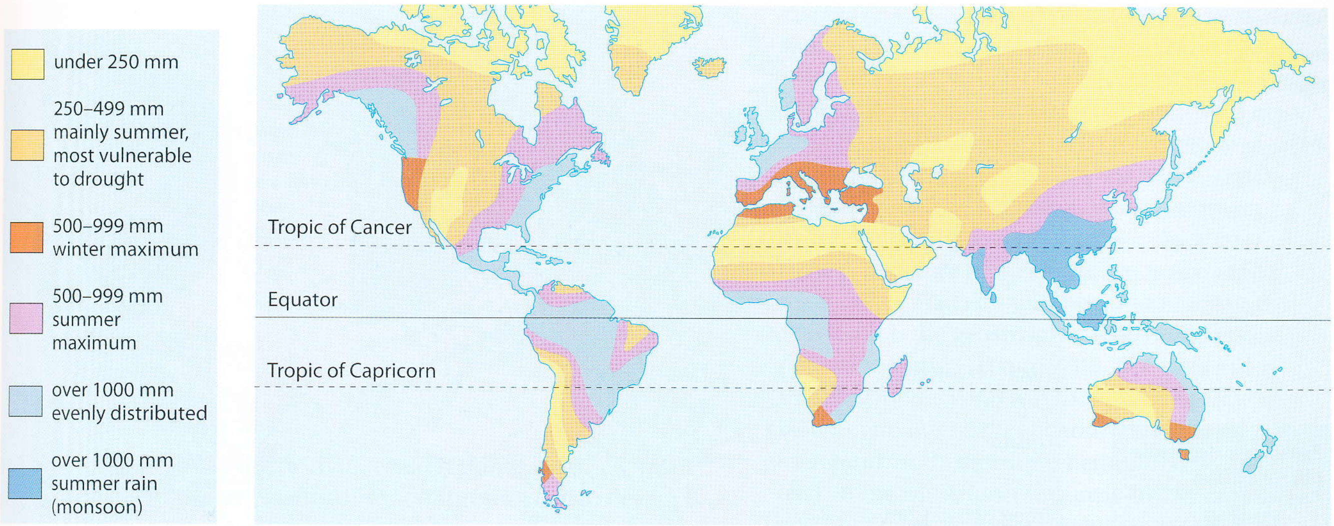


Figure 9.27

World precipitation: mean annual totals and seasonal distribution

World precipitation: distribution and reliability

Geographers are interested in describing distributions and in identifying and accounting for any resultant patterns. Where precipitation is concerned, geographers have, in the past, concentrated on long-term distributions which show either mean annual amounts or seasonal variations. Long-term fluctuations vary considerably across the globe but, nevertheless, a map showing world precipitation does show identifiable patterns (Figure 9.27).

Equatorial areas have high annual rainfall totals due to the continuous uplift of air resulting from the convergence of the trade winds and strong convectional currents (page 226). The presence of the ITCZ ensures that rain falls throughout the year. Further away from the Equator, rainfall totals decrease and the length of the dry season increases. These tropical areas, especially those inland, experience convectional rainfall in summer, when the sun is overhead, followed by a dry winter. Latitudes adjacent to the tropics receive minimal amounts as they correspond to areas of high pressure caused by subsiding, and therefore warming, air (Figure 7.2).

To the poleward side of this arid zone, rainfall quantities increase again and the length of the dry season decreases. These temperate latitudes receive large amounts of rainfall, spread evenly throughout the year, due to cyclonic conditions and local orographic effects. Towards the polar areas, where cold air descends to give stable conditions, precipitation totals decrease and rain gives way to snow. Between 30° and 40° north and south (in the west of continents) the Mediterranean climate is characterised by winter rain and summer drought. This general latitudinal zoning of rainfall is interrupted locally by the apparent movement of the overhead sun, the presence of mountain ranges or ocean currents, the monsoon, and continentality (distance from the sea).

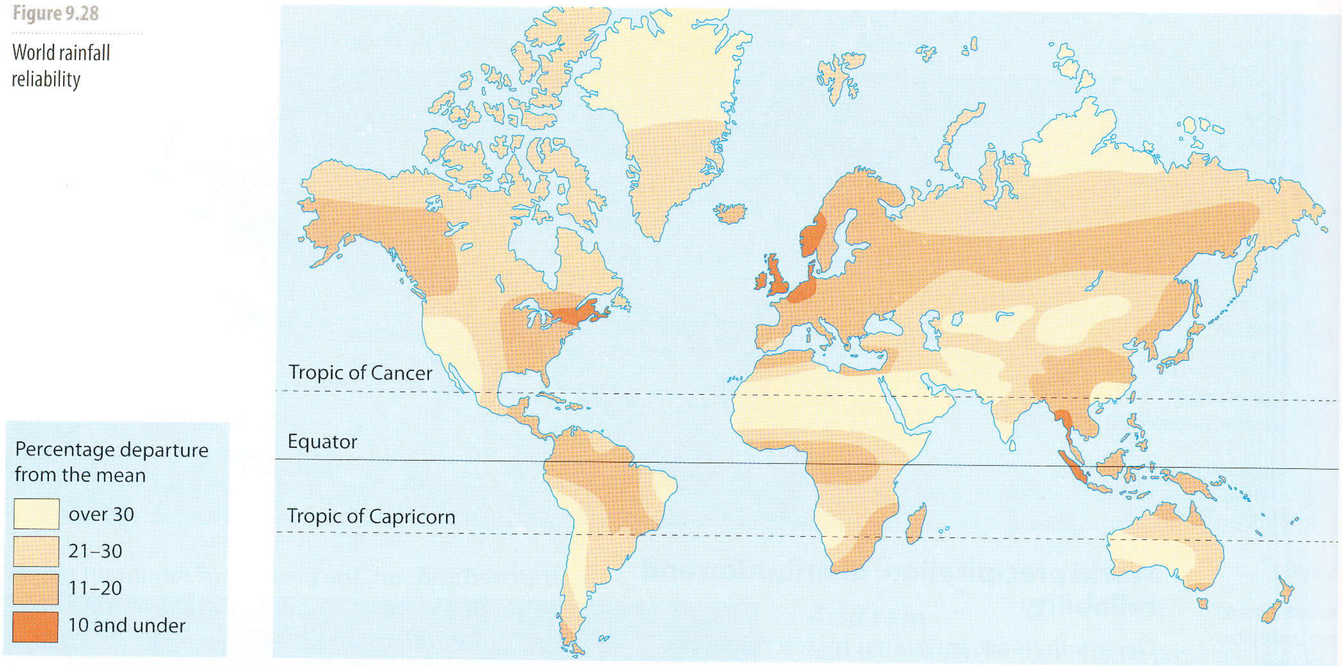
More recently, geographers have become increasingly concerned with shorter-term variations. In many parts of the world, economic development and lifestyles are more closely linked to the duration, intensity and reliability of rainfall than to annual amounts. Precipitation is more valuable when it falls during the growing season (Canadian Prairies) and less effective if it occurs when evapotranspiration rates are at their highest (Sahel countries). In the same way, lengthy episodes of steady rainfall as experienced in Britain provide a more beneficial water supply than storms of a short and intensive duration which occur in tropical semi-arid climates. This is because moisture penetrates the soil more gradually and the risks of soil erosion, flooding and water shortages are reduced.

Of utmost importance is the reliability of rainfall. There appears to be a strong positive correlation (Framework 19, page 612) between rainfall totals and rainfall reliability – i.e. as rainfall totals increase, so too does rainfall reliability. In Britain and the Amazon Basin, rainfall is reliable with relatively little variation in annual totals from year to year (Figure 9.28).

Elsewhere, especially in monsoon or tropical continental climates, there is a pronounced wet and dry season. Consequently, if the rains fail one year, the result can be disastrous for crops, and possibly also for animals and people. The most vulnerable areas, such as north-east Brazil and the Sahel countries, lie near to desert margins (Figure 9.28). Here, where even a small variation of 10 per cent below the mean can be critical, many places often experience a variation in excess of 30 per cent.

Figure 9.28

World rainfall reliability

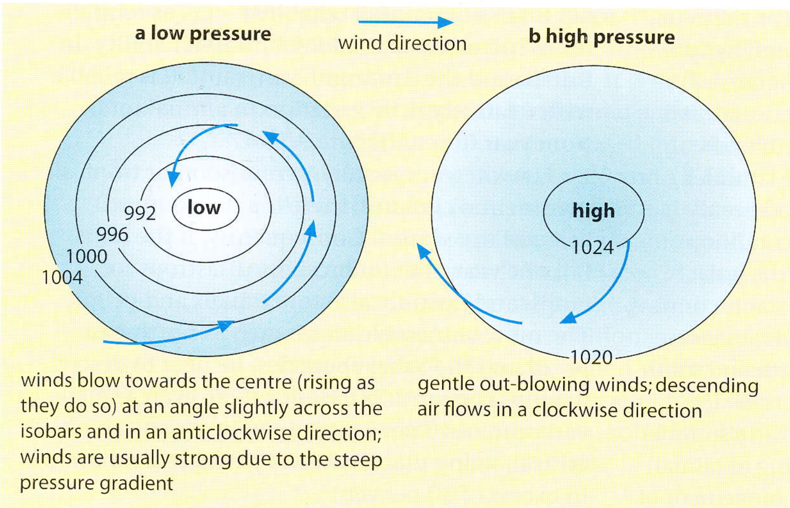


Atmospheric motion

The movement of air in the atmospheric system may be vertical (i.e. rising or subsiding) or horizontal; in the latter case it is commonly known as **wind**. Winds result from differences in air pressure which in turn may be caused by differences in temperature and the force exerted by gravity, as pressure decreases rapidly with height (Figure 9.1). An increase in temperature causes air to heat, expand, become less dense and rise, creating an area of low pressure below. Conversely, a drop in temperature produces an area of high pressure. Differences in pressure are shown on maps by **isobars**, which are lines joining places of equal pressure. To draw isobars, pressure readings are normally reduced to represent pressure at sea-level. Pressure is measured in **millibars** (mb) and it is usual for isobars to be drawn at 4 mb intervals.

Figure 9.29

The two basic pressure systems affecting Britain



Average pressure at sea-level is 1013 mb. However, the isobar pattern is usually more important in terms of explaining the weather than the actual figures. The closer together the isobars, the greater the difference in pressure – **the pressure gradient** – and the stronger the wind. Wind is nature's way of balancing out differences in pressure as well as temperature and humidity.

Figure 9.29 shows the two basic pressure systems which affect the British Isles. In addition to the differences in pressure, wind speed and wind direction, the diagrams also show that winds blow neither directly at right-angles to the isobars along the pressure gradient, nor parallel to them. This is due to the effects of the Coriolis force and of friction, which are additional to the forces exerted by the pressure gradient and gravity.

The Coriolis force

If the Earth did not rotate and was composed entirely of either land or water, there would be one large convection cell in each hemisphere (Figure 9.30). Surface winds would be parallel to pressure gradients and would blow directly from high to low pressure areas. In reality, the Earth does rotate and the distribution of land and sea is uneven. Consequently, more than one cell is created (Figures 9.34 and 9.35) as rising air, warmed at the Equator, loses heat to space – there is less cloud cover to retain it – and as it travels further from its source of heat. A further consequence is that moving air appears to be deflected to the right in the northern hemisphere and to the left in the southern hemisphere. This is a result of the Coriolis force.

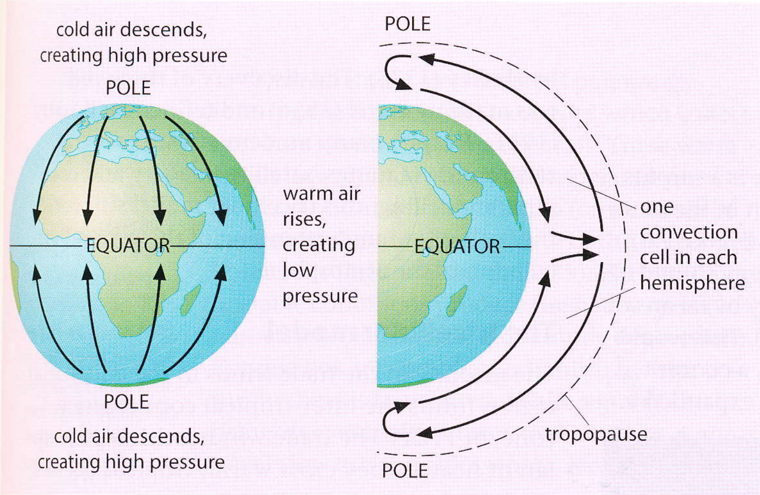


Figure 9.30

Air movement on a rotation-free Earth

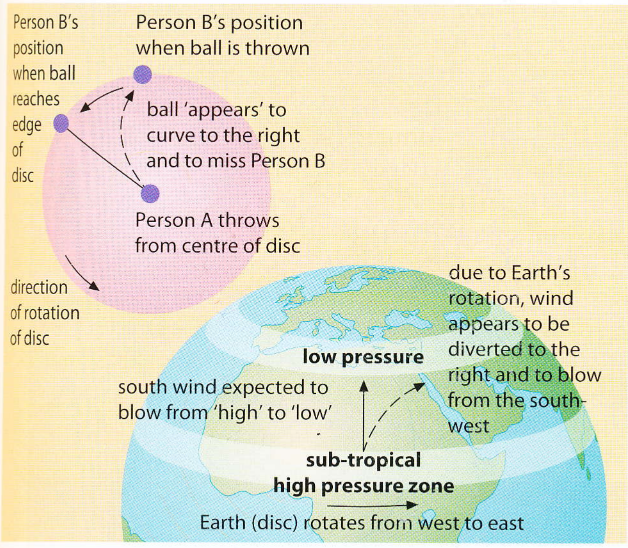
Figure 9.31

The Coriolis force in the northern hemisphere

Imagine that Person A stands in the centre of a large rotating disc and throws a ball to Person B, standing on the edge of that moving disc. As Person A watches, the ball appears to take a curved path away from Person B – due to the fact that, while the ball is in transit, Person B has been moved to a new position by the rotation of the disc (Figure 9.31). Similarly, the Earth's rotation through 360° every 24 hours means that a wind blowing in a northerly direction in the northern hemisphere appears to have been diverted to the right on a curved trajectory by 15° of longitude for every hour (though to an astronaut in a space shuttle, the path would look straight). This helps to explain why the prevailing winds blowing from the tropical high pressure zone approach Britain from the south-west rather than from the south. In theory, if the Coriolis force acted alone, the resultant wind would blow in a circle.

Winds in the upper troposphere, unaffected by friction with the Earth's surface, show that there is a balance between the forces exerted by the pressure gradient and the Coriolis deflection. The result is the **geostrophic wind** which blows parallel to isobars (Figure 9.32). The existence of the geostrophic wind was recognised in 1857 by a Dutchman, Buys Ballot, whose law states that 'if you stand, in the northern hemisphere, with your back to the wind, low pressure is always to your left and high pressure to your right'.

Friction, caused by the Earth's surface, upsets the balance between the pressure gradient and the Coriolis force by reducing the effect of the latter. As the pressure gradient becomes relatively more important when friction is reduced with altitude, the wind blows across isobars towards the low pressure (Figure 9.29). Deviation from the geostrophic wind is less pronounced over water because its surface is smoother than that of land.



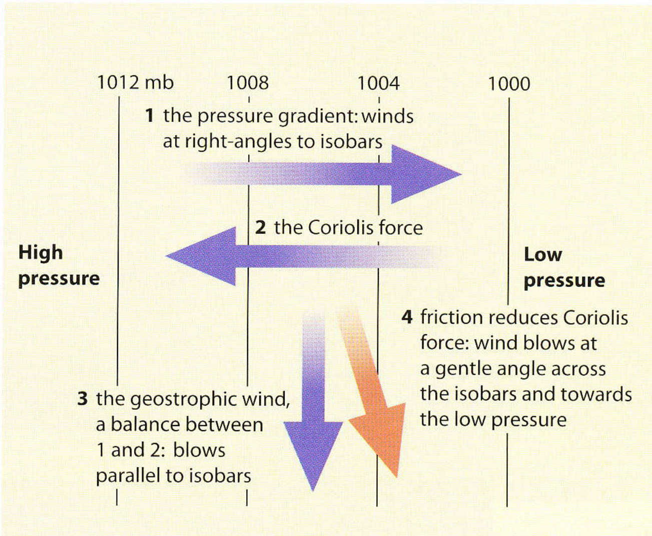


Figure 9.32

The geostrophic wind and the effect of friction (in the northern hemisphere)

A hierarchy of atmospheric motion

An appreciation of the movement of air is fundamental to an understanding of the workings of the atmosphere and its effects on our weather and climate. The extent to which atmospheric motion influences local weather and climate depends on winds at a variety of scales and their interaction in a hierarchy of patterns. One such hierarchy, which is useful in studying the influence of atmospheric motion, was suggested by B.W. Atkinson in 1988. Although defining four levels, he stressed that there were important interrelationships between each (Figure 9.33).

Figure 9.33

A hierarchy of atmosphere motion systems (*after* Atkinson, 1988)

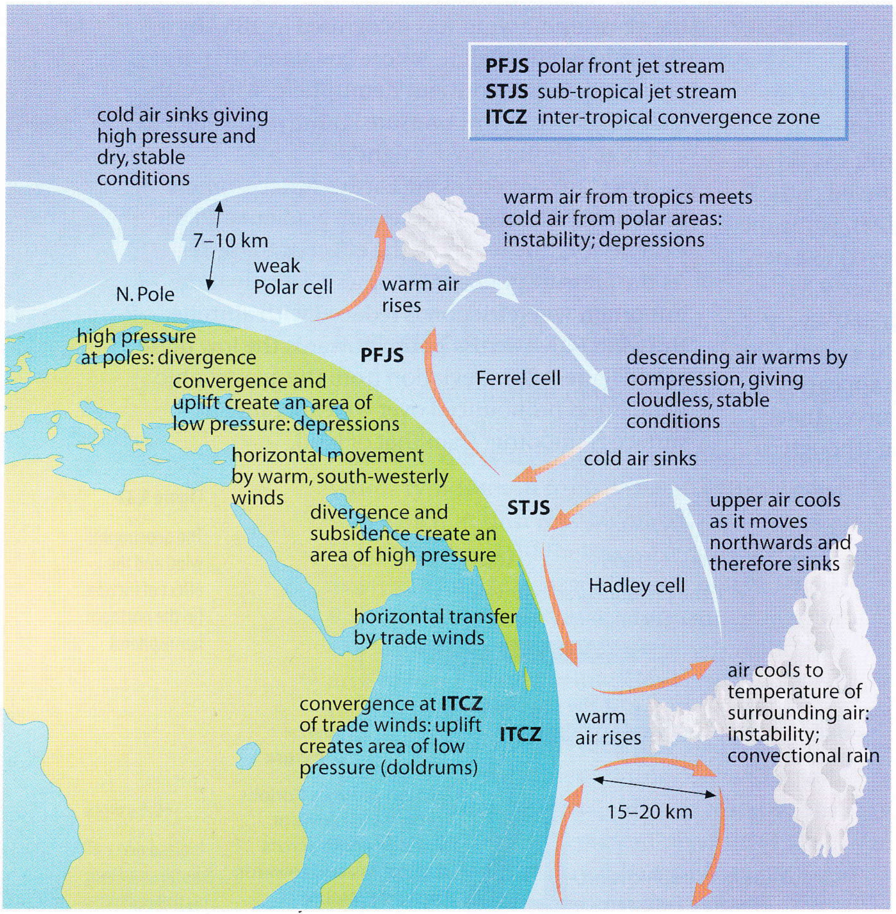
Scale	Characteristic horizontal size (km)	Systems
1 Planetary	5000–10 000	Rossby waves, ITCZ
2 Synoptic (macro)	1000–5000	Monsoons, hurricanes, depressions, anticyclones
3 Meso-scale	10–1000	Land and sea breezes, mountain and valley winds, föhn, thunderstorms
4 Small (micro)	0.1–10	Smoke plumes, urban turbulence

Planetary scale: atmospheric circulation

It has already been shown that there is a surplus of energy at the Equator and a deficit in the outer atmosphere and nearer to the poles (Figure 9.6). Therefore, theoretically, surplus energy should be transferred to areas with a deficiency by means of a single convective cell (Figure 9.30). This would be the case for a non-rotating Earth, a concept first advanced by Halley (1686) and expanded by Hadley (1735). The discovery of three cells was made by Ferrel (1856) and refined by Rossby (1941). Despite many modern advances using **radiosonde** readings, satellite imagery and computer modelling, this tricellular model still forms the basis of our understanding of the general circulation of the atmosphere.

Figure 9.34

Tricellular model showing atmospheric circulation in the northern hemisphere



The tricellular model

The meeting of the trade winds in the equatorial region forms the inter-tropical convergence zone, or **ITCZ**. The trade winds, which pick up latent heat as they cross warm, tropical oceans, are forced to rise by violent convection currents. The unstable, warm, moist air is rapidly cooled adiabatically to produce the towering cumulo-nimbus clouds, frequent afternoon thunderstorms and low pressure characteristic of the equatorial climate (page 316). It is these strong upward currents that form the 'power-house of the general global circulation' and which turn latent heat first into sensible heat and later into potential energy. At ground-level, the ITCZ experiences only very gentle, variable winds known as the **doldrums**.

As rising air cools to the temperature of the surrounding environmental air, uplift ceases and it begins to move away from the Equator. Further cooling, increasing density, and diversion by the Coriolis force cause the air to slow down and to subside, forming the descending limb of the **Hadley cell** (Figures 9.34 and 9.35). In looking at the northern hemisphere (the southern is its mirror image), it can be seen that the air subsides at about 30°N of the Equator to create the sub-tropical high pressure belt with its clear skies and dry, stable conditions (Figure 9.36). On reaching the Earth's surface, the cell is completed as some of the air is returned to the Equator as the north-east trade winds.

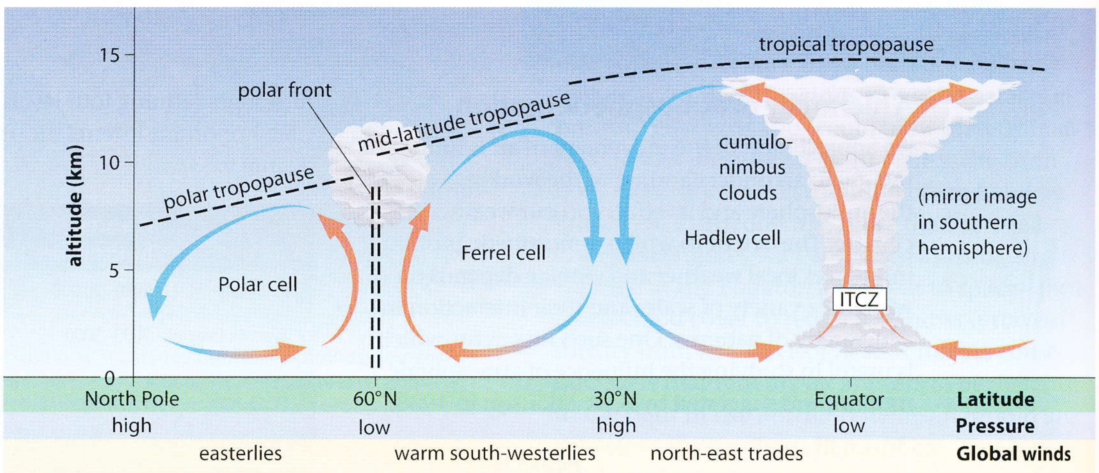


Figure 9.35

Tricellular model to show atmospheric circulation in the northern hemisphere and within the tropopause

The remaining air is diverted polewards, forming the warm south-westerlies which collect moisture when they cross sea areas. These warm winds meet cold Arctic air at the polar front (about 60°N) and are uplifted to form an area of low pressure and the rising limb of the **Ferrel** and **Polar cells** (Figures 9.34 and 9.35). The resultant unstable conditions produce the heavy cyclonic rainfall associated with mid-latitude **depressions**. Depressions are another mechanism by which surplus heat is transferred. While some of this rising air eventually returns to the tropics, some travels towards the poles where, having lost its heat, it descends to form another stable area of high pressure. Air returning to the polar front does so as the cold easterlies.

This overall pattern is affected by the apparent movement of the overhead sun to the north and south of the Equator. This movement causes the seasonal shift of the heat Equator, the ITCZ, the equatorial low pressure zone and global wind and rainfall belts. Any variation in the characteristics of the ITCZ – i.e. its location or width – can have drastic consequences for the surrounding climates, as seen in the Sahel droughts of the early 1970s and most of the 1980s (Case Study 7).

Figure 9.36

Image taken by the Meteosat geosynchronous satellite. Notice the clouds resulting from uplift at the ITCZ (not a continuous belt), the clear skies over the Sahara, the polar front over the north Atlantic, and a depression over Britain

Rossby waves and jet streams

Evidence of strong winds in the upper troposphere first came when First World War Zeppelins were blown off-course, and several inter-war balloons were observed travelling at speeds in excess of 200 km/hr. Pilots in the Second World War, flying at heights above 8 km, found eastward flights much faster and their return westward journeys much slower than expected, while north–south flights tended to be blown off-course. The explanation was found to be the **Rossby waves**, which often follow a meandering path (Figure 9.37a), distorting the upper-air westerlies. The number of meanders, or waves, varies seasonally, with usually four to six in summer and three in winter. These waves form a complete pattern around the globe (Figure 9.37b and c).

Further investigation has shown that the velocity of these upper westerlies is not internally uniform. Within them are narrow bands of extremely fast-moving air known as **jet streams**. Jet streams, which help in the rapid transfer of energy, can exceed speeds of 230 km/hr, which is sufficient to carry a balloon, or ash from a volcano, around the Earth within a week or two (Figure 9.39 and Case Study 1). Of five recognisable jet streams, two are particularly significant, with a third having seasonal importance.

The **polar front jet stream** (PFJS, Figure 9.34) varies between latitudes 40° and 60° in both hemispheres and forms the division between the Ferrel and Polar cells, i.e. the boundary between warm tropical and cold polar air. The PFJS varies in extent, location and intensity and is mainly responsible for giving fine or wet weather on the Earth's surface. Where, in the northern hemisphere, the jet stream moves south (Figure 9.38a), it brings with it cold air which descends in a clockwise direction to give dry, stable conditions associated with areas of high pressure (**anticyclones**, page 234). When the now-warmed jet stream backs northwards, it takes with it warm air which rises in an anticlockwise direction to give the strong winds and heavy rainfall associated with areas of low pressure (**depressions**, page 230). As the usual path of the PFJS over Britain is oblique – i.e. towards the north-east – this accounts for our frequent wet and windy weather. Occasionally, this path may be temporarily altered by a stationary or **blocking anticyclone** (Figures 9.38b and 9.48) which may produce extremes of climate such as the hot, dry summers of 1976 and 1989 or the cold January of 1987.

The **subtropical jet stream**, or STJS, occurs about 25° to 30° from the Equator and forms the boundary between the Hadley and Ferrel cells (Figures 9.34 and 9.35). This meanders less than the PFJS, has lower wind velocities, but follows a similar west–east path.

The **easterly equatorial jet stream** is more seasonal, being associated with the summer monsoon of the Indian subcontinent (page 239).

Figure 9.37

Rossby waves and jet streams (northern hemisphere)

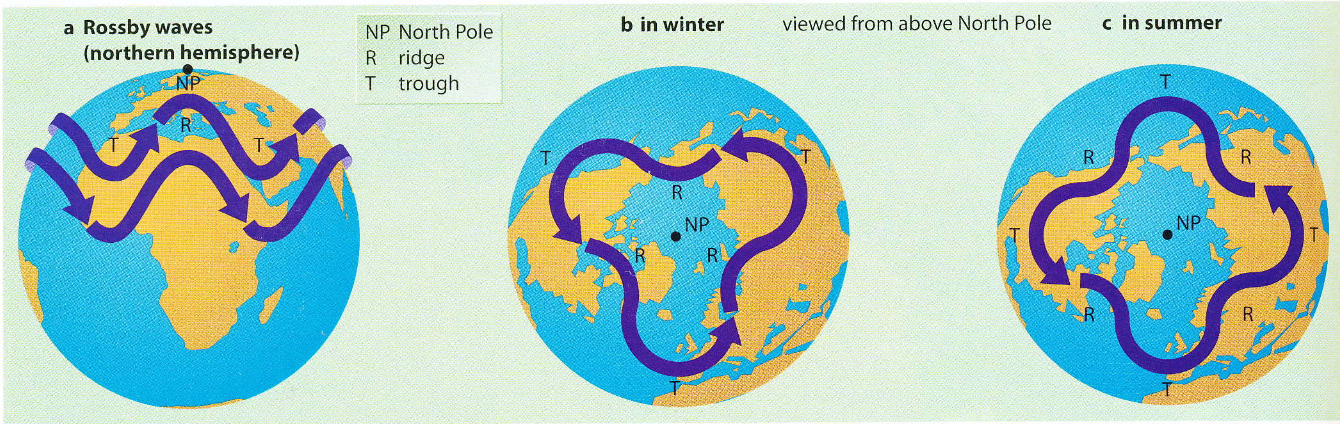


Figure 9.38

The polar front jet stream (PFJS) (northern hemisphere)

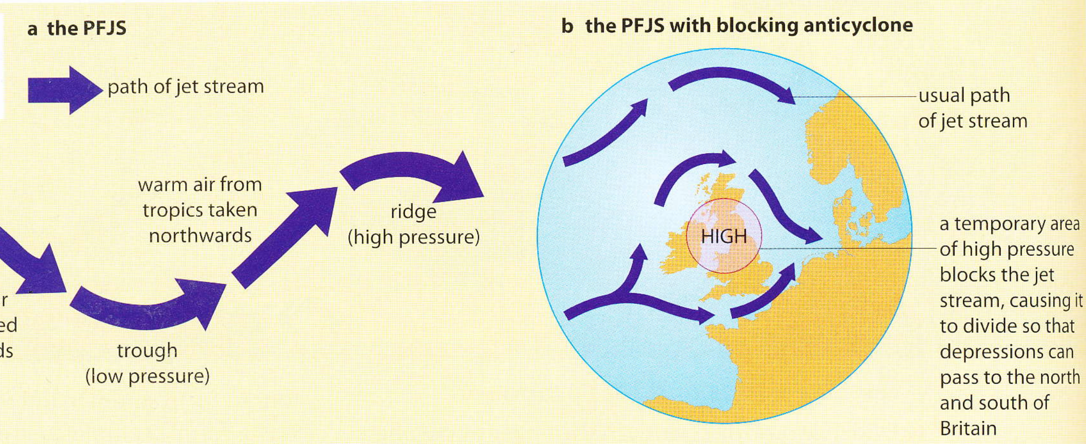
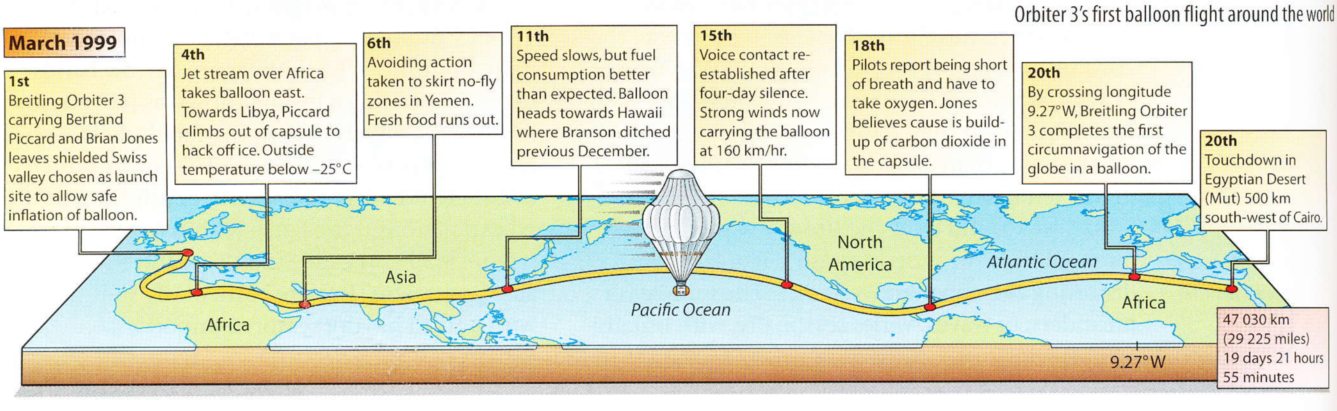


Figure 9.39

Orbiter 3's first balloon flight around the world



The **polar front jet stream** (PFJS, Figure 9.34) varies between latitudes 40° and 60° in both hemispheres and forms the division between the Ferrel and Polar cells, i.e. the boundary between warm tropical and cold polar air. The PFJS varies in extent, location and intensity and is mainly responsible for giving fine or wet weather on the Earth's surface. Where, in the northern hemisphere, the jet stream moves south (Figure 9.38a), it brings with it cold air which descends in a clockwise direction to give dry, stable conditions associated with areas of high pressure (**anticyclones**, page 234). When the now-warmed jet stream backs northwards, it takes with it warm air which rises in an anticlockwise direction to give the strong winds and heavy rainfall associated with areas of low pressure (**depressions**, page 230). As the usual path of the PFJS over Britain is oblique – i.e. towards the north-east – this accounts for our frequent wet and windy weather. Occasionally, this path may be temporarily altered by a stationary or **blocking anticyclone** (Figures 9.38b and 9.48) which may produce extremes of climate such as the hot, dry summers of 1976 and 1989 or the cold January of 1987.

The **subtropical jet stream**, or STJS, occurs about 25° to 30° from the Equator and forms the boundary between the Hadley and Ferrel cells (Figures 9.34 and 9.35). This meanders less than the PFJS, has lower wind velocities, but follows a similar west–east path.

The **easterly equatorial jet stream** is more seasonal, being associated with the summer monsoon of the Indian subcontinent (page 239).

Figure 9.37

Rossby waves and jet streams (northern hemisphere)

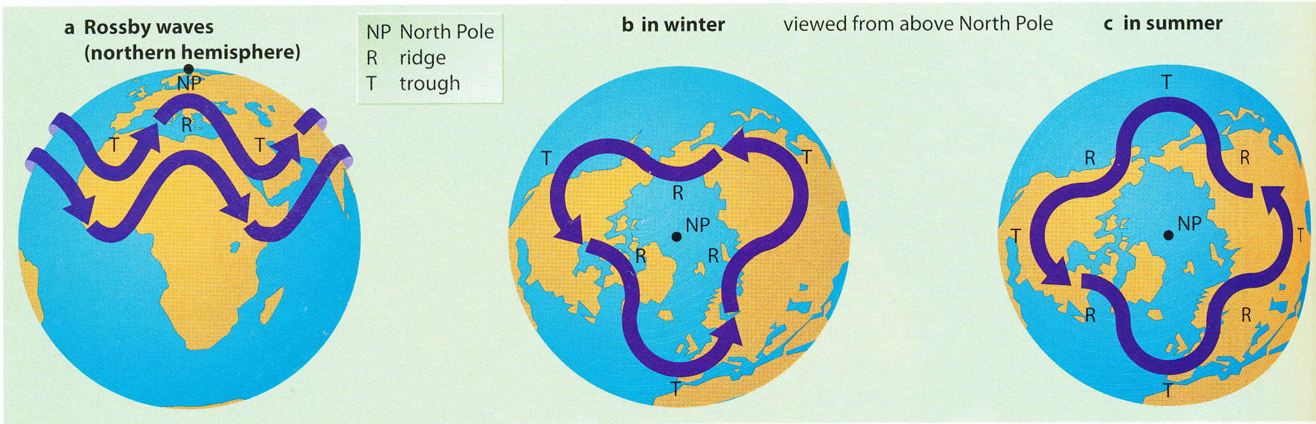


Figure 9.38

The polar front jet stream (PFJS) (northern hemisphere)

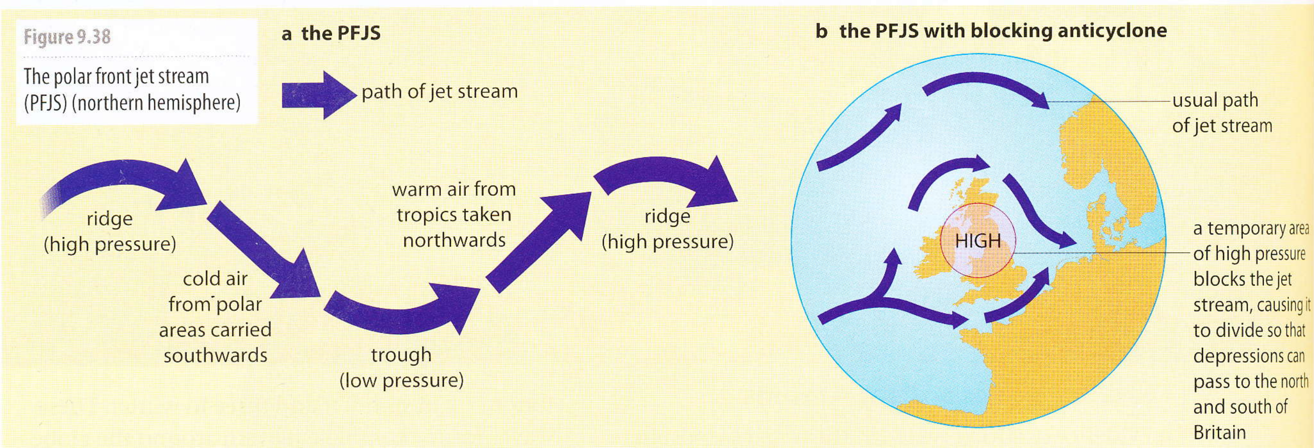
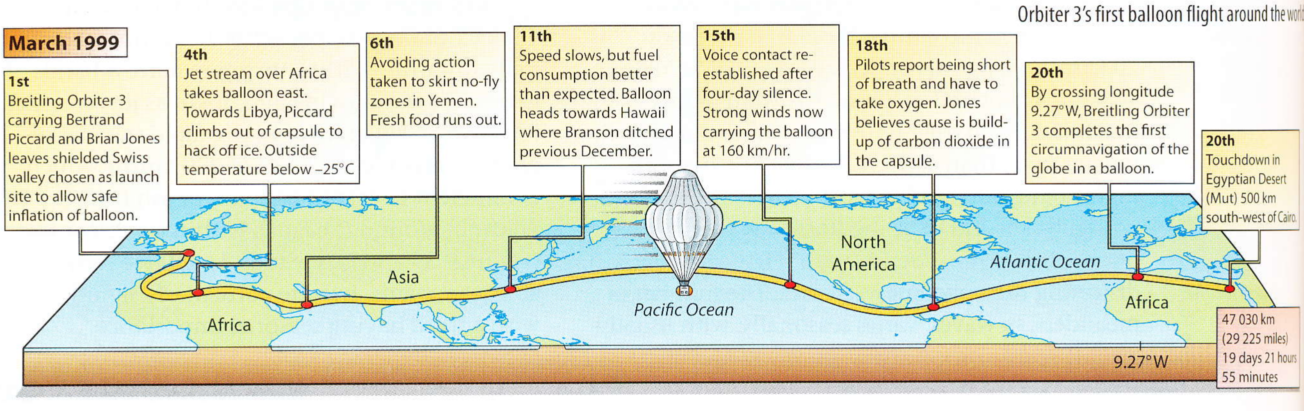


Figure 9.39

Orbiter 3's first balloon flight around the world



Macro-scale: synoptic systems

The concept of air masses is important because air masses help to categorise world climate types (Chapter 12). In regions where one air mass is dominant all year, there is little seasonal variation in weather, for example at the tropics and at the poles. Areas such as the British Isles, where air masses constantly interchange, experience much greater seasonal and diurnal variation in their weather.

Air masses and fronts: how they affect the British Isles

If air remains stationary in an area for several days, it tends to assume the temperature and humidity properties of that area. Stationary air is mainly found in the high pressure belts of the subtropics (the Azores and the Sahara) and in high latitudes (Siberia and northern Canada). The areas in which homogeneous air masses develop are called **source regions**. Air masses can be classified according to:

- the **latitude** in which they originate, which determines their temperature – Arctic (*A*), Polar (*P*), or Tropical (*T*)
- the nature of the **surface** over which they develop, which affects their moisture content – maritime (*m*), or continental (*c*).

The five major air masses which affect the British Isles at various times of the year (*Am*, *Pm*, *Pc*, *Tm* and *Tc*) are derived by combining these characteristics of latitude and humidity (Figure 9.40). When air masses move from their source region they are modified by the surface over which they pass and this alters their temperature, humidity and stability. For example, tropical air moving northwards is cooled and becomes more stable, while polar air moving south becomes warmer and increasingly unstable. Each air mass therefore brings its own characteristic weather conditions to the British Isles. The general conditions expected with each air mass are given in Figure 9.41. However, it should be remembered that each air mass is unique and dependent on: the climatic conditions in the source region at the time of its development; the path which it subsequently follows; the season in which it occurs; and, since it has a three-dimensional form, the vertical characteristics of the atmosphere at the time.

When two air masses meet, they do not mix readily, due to differences in temperature and density. The point at which they meet is called a **front**. A **warm front** is found where warm air is advancing and being forced to override cold air. A **cold front** occurs when advancing cold air undercuts a body of warm air. In both cases, the rising air cools and usually produces clouds, easily seen on satellite weather photographs (Figures 9.67 and 9.68); these clouds often generate precipitation. Fronts may be several hundred kilometres wide and they extend at relatively gentle gradients up into the atmosphere. The most notable type of front, the **polar front**, occurs when warm, moist, *Tm* air meets colder, drier, *Pm* air. It is at the polar front that **depressions** form. Depressions are areas of low pressure. They form most readily over the oceans in mid-latitudes, and track eastwards bringing cloud and rain to western margins of continents.

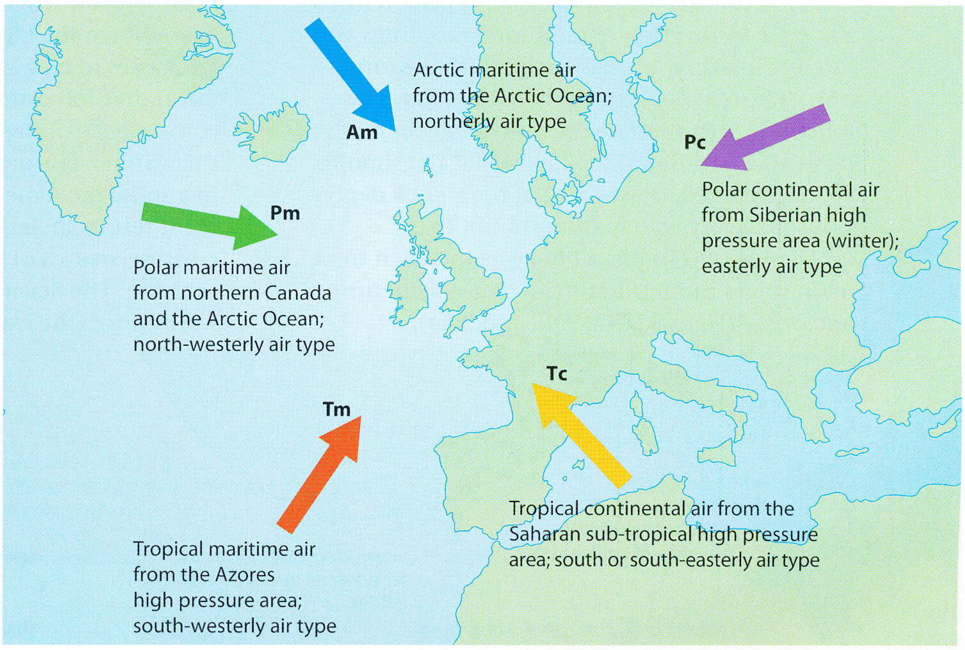


Figure 9.40

Air masses that affect the British Isles

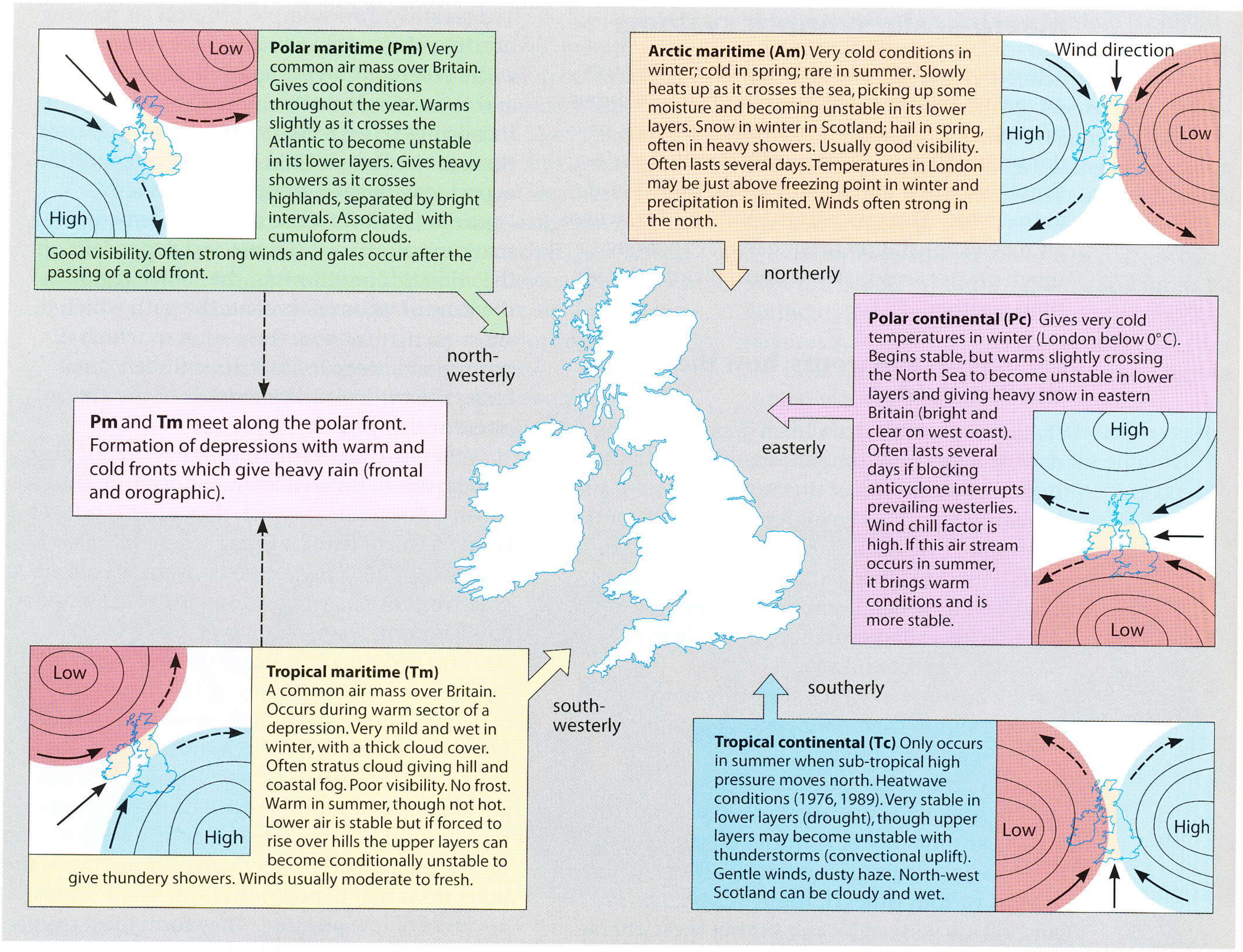


Figure 9.41

Air masses and the British weather

Depressions

The polar front theory was put forward by a group of Norwegian meteorologists in the early 1920s. Although some aspects have been refined since the innovation of radiosonde readings and satellite imagery, the basic model for the formation of frontal depressions remains valid. The following account describes a 'typical' or 'model' depression (Framework 12, page 352). It should be remembered, however, that individual depressions may vary widely from this model.

Depressions follow a life-cycle in which three main stages can be identified: embryo, maturity and decay (Figures 9.42, 9.43 and 9.44).

1 The **embryo depression** begins as a small wave on the polar front. It is here that warm, moist, tropical (*Tm*) air meets colder, drier, polar (*Pm*) air (Figure 9.42). Recent studies have shown the boundary between the two air masses to be a zone rather than the simple linear division claimed in early models. The convergence of the two air masses results in the warmer, less dense air being forced to rise in a spiral movement. This upward movement results in 'less' air at the Earth's surface, creating an area of below-average or low pressure. The developing depression, with its warm front (the leading edge of the tropical

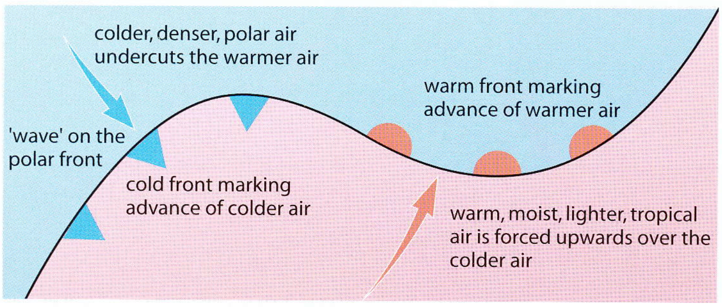


Figure 9.42

Life-cycle of a depression: Stage 1 – embryo depression air) and cold front (the leading edge of the polar air), usually moves in a north-easterly direction under the influence of the upper westerlies, i.e. the polar front jet stream.

2 A **mature depression** is recognised by the increasing amplitude of the initial wave (Figure 9.43). Pressure continues to fall as more warm air, in the warm sector, is forced to rise. As pressure falls and the pressure gradient steepens, the inward-blowing winds increase in strength. Due to the Coriolis force (page 224), these anticlockwise-blowing winds come from the south-west. As the relatively warm air of the warm sector continues to rise along the warm front, it eventually cools to dew point. Some of its vapour will condense to release large amounts of latent heat, and clouds will develop. Continued uplift and cooling will cause precipitation as the clouds become both thicker and lower.

Satellite photographs have shown that there is likely to be a band of 'meso-scale precipitation' extending several hundred kilometres in length and up to 150 km in width along, and just in front of, a warm front. As temperatures rise and the uplift of air decreases within the warm sector, there is less chance of precipitation and the low cloud may break to give some sunshine. The cold front moves faster and has a steeper gradient than the warm front (Figure 9.45).

Progressive undercutting by cold air at the rear of the warm sector gives a second episode of precipitation – although with a greater intensity and a shorter duration than at the warm front. This band of meso-scale precipitation may be only 10–50 km in width. Although the air behind the cold front is colder than that in advance of the warm front (having originated in and travelled through more northerly latitudes), it becomes unstable, forming cumulo-nimbus clouds and heavy showers. Winds often reach their maximum strength at the cold front and change to a more north-westerly direction after its passage (Figure 9.45).

3 The depression begins to **decay** when the cold front catches up the warm front to form an **occlusion** or **occluded front** (Figure 9.44). By this stage, the *Tm* air will have been squeezed upwards leaving no warm sector at ground-level. As the uplift of air is reduced, so too are (or will be) the amount of condensation, the release of latent heat and the amount and pattern of precipitation – there may be only one episode of rain. Cloud cover begins to decrease, pressure rises and wind speeds decrease as the colder air replaces the uplifted air and 'infills' the depression.

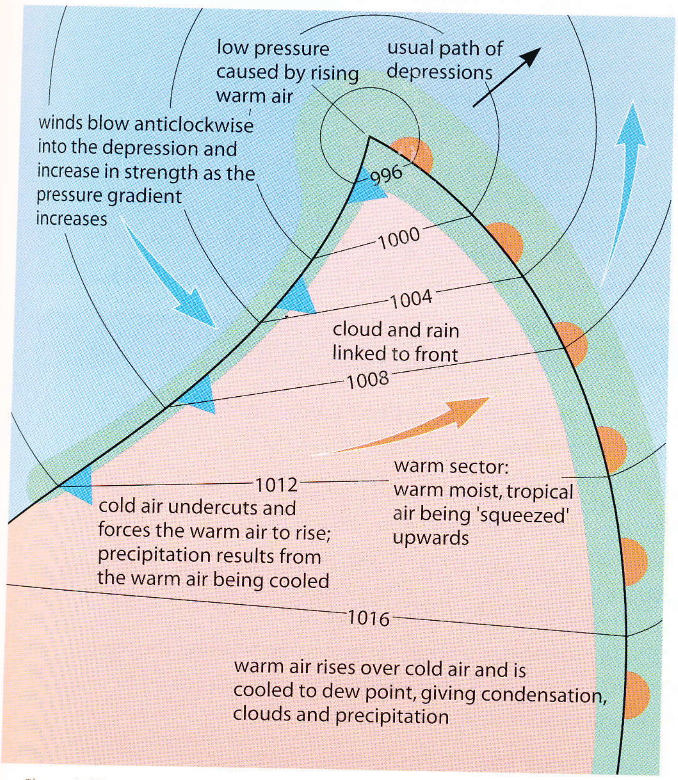


Figure 9.43

Life-cycle of a depression: Stage 2 – maturity

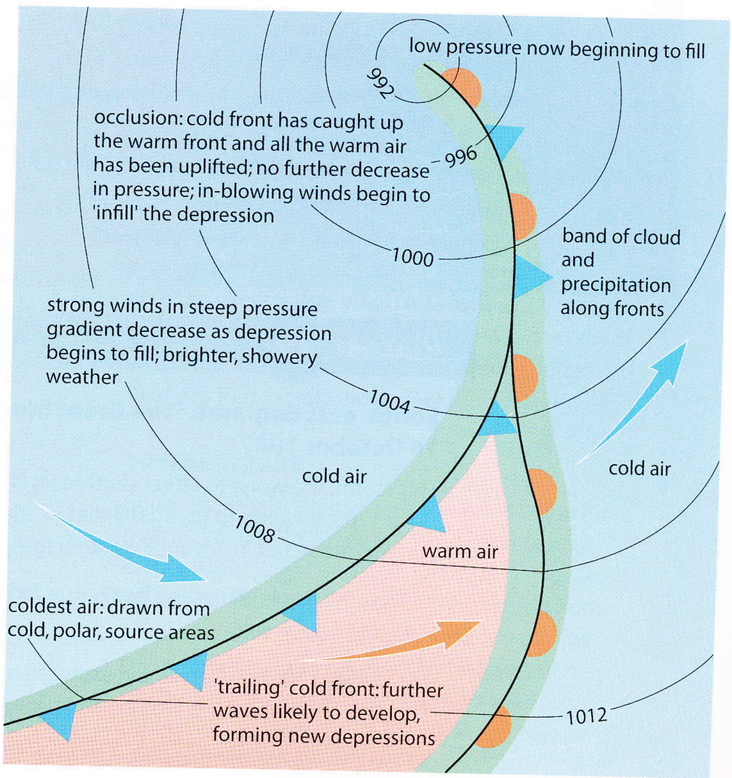
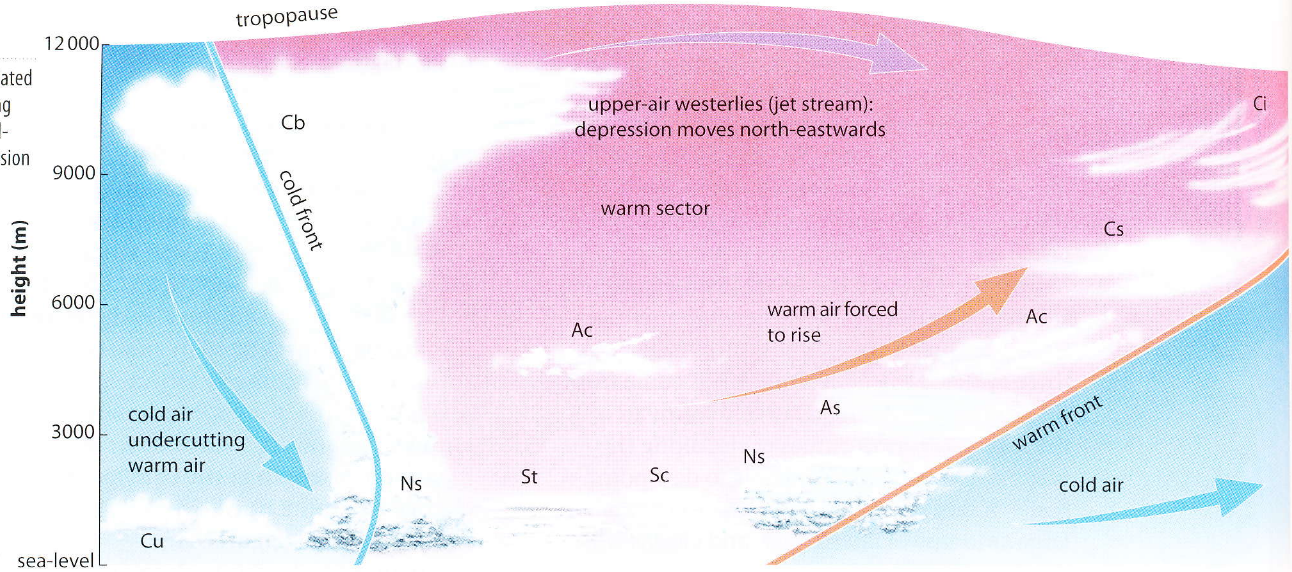


Figure 9.44

Life-cycle of a depression: Stage 3 – decay

Figure 9.45

Weather associated with the passing of a typical mid-latitude depression



read from right to left (i.e. from 1 to 5)

	5. Behind the cold front	4. Passing of the cold front	3. Warm sector	2. Passing of the warm front	1. Approach of depression
Pressure	rise continues more slowly	sudden rise	steady	fall ceases	steady fall
Wind direction	NW	veers from SW to NW	SW	veers from SSE to SW	SSE
Wind speed	squally; speed slowly decreases (e.g. force 3–6)	very strong to gale force (e.g. force 6–8)	decreases (e.g. force 2–4)	strong (e.g. force 5–6)	slowly increases (e.g. force 1–3)
Temperature (e.g. winter)	cold (e.g. 3°C)	sudden decrease	warm/mild (e.g. 10°C)	sudden rise	cool (e.g. 6°C)
Relative humidity	rapid fall	high during precipitation	steady and high	high during precipitation	slow rise
Cloud (Figure 9.20)	decreasing; in succession, Cb and Cu	very thick and towering Cb	low or may clear; St, Sc, Ac	low and thick Ns	high and thin; in succession, Ci, Cs, Ac, As
Precipitation	heavy showers	short period of heavy rain or hail	drizzle or stops raining	continuous rainfall, steady and quite heavy	none
Visibility	very good; poor in showers	poor	often poor	decreases rapidly	good but beginning to decrease

Places 29 Storms in southern England

South-east England: 'The Great Storm', 16 October 1987

This storm, the worst to affect south-east England since 1703, developed so rapidly that its severity was not predicted in advance weather forecasts.

***11 October*:** High winds and heavy rain forecast for the end of the week.

***15 October* 1200 hrs:** Depression expected to move along the English Channel with fresh to strong winds.

2130 hrs: TV weather forecast: strong winds gusting to 50 km/hr.

***16 October* 0030 hrs:** Radio weather forecast: warning of severe gales.

0130 hrs: Police and fire services alerted about extreme winds.

0500 hrs: Winds reached 94 km/hr at Heathrow and 100 km/hr on parts of the south coast.

0800 hrs: Centre of depression reached the North Sea. Winds over southern England dropped to 50–70 km/hr.

1200 hrs: 'The Great Storm' was over.

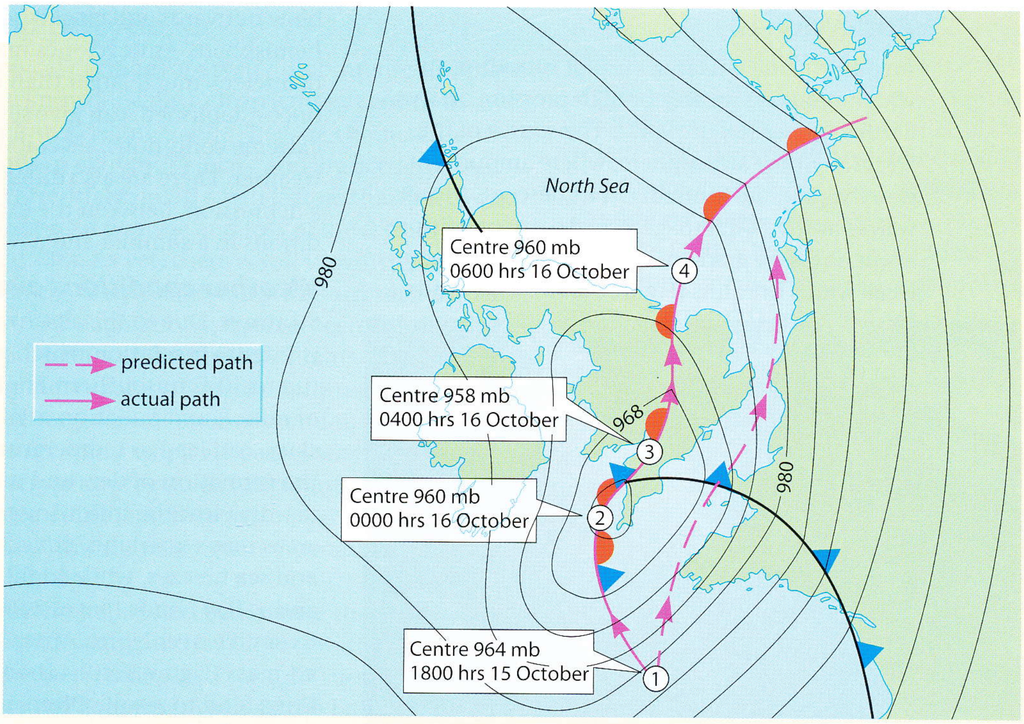


Figure 9.46

'The Great Storm', 16 October 1987

The storm began on 15 October as a small wave on a cold front in the Bay of Biscay, where the few weather ships give only limited information. It was caused by contact between very warm air from Africa and cold air from the North Atlantic. It appeared to be a 'typical' depression until, at about 1800 hrs on 15 October, it unexpectedly deepened giving a central pressure reading of 964 mb and creating an exceptionally steep pressure gradient. The exact cause of this is unknown but it was believed to result from a combination of an exceptionally strong jet stream (initiated on 13 October by air spiralling upwards along the east coast of North America in Hurricane Floyd) and extreme warming over the Bay of Biscay (see hurricanes, page 235). Together, these could have caused an excessive release of latent heat energy which North American meteorologists compare with the effect of detonating a bomb. It was this unpredicted deepening, combined with the change of direction from the English Channel towards the Midlands, which caught experts by surprise.

The depression moved rapidly across southern England, clearing the country in six hours (Figure 9.46). Winds remained light in and around the centre (Birmingham 13 km/hr), but the strong pressure gradient on its southern flank resulted in severe winds from Portland Bill (102 km/hr, gusting to 141 km/hr) to Dover (115 km/hr, gusting to 167 km/hr).

Although the storm passed within a few hours, and luckily during the night when most people were asleep, it left a trail of death and destruction. There were 16 deaths; several houses collapsed and many others lost walls, windows and roofs; an estimated 15 million trees were blown over, blocking railways and roads; one-third of the trees in Kew Gardens were destroyed; power lines were cut and, in some remote areas, not restored for several days; few commuters managed to reach London the next day; a ferry was blown ashore at Folkestone; and insurance claims set an all-time record.

Once every 50 years, winds exceeding 100 km/hr with gusts of over 165 km/hr can be expected north of a line from Cornwall to Durham, and even stronger winds, gusting to 185 km/hr, once in 20 years in western and northern Scotland. The winds associated with the Great Storm were remarkable not so much for their strength as for their occurrence over south-east England. Here, the predicted return period can be measured in centuries rather than decades.

10 March 2008

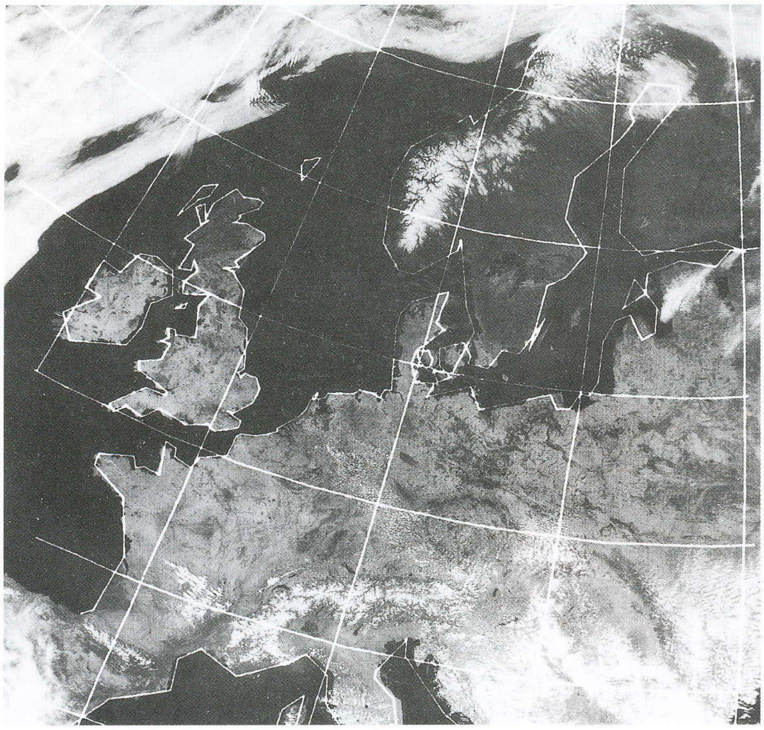
Southern Britain experienced the worst storm for over 20 years with winds of 150 km/hr recorded on the Isle of Wight and torrential rain falling over Wales and southern England. Flights to and from Heathrow were either cancelled or diverted and there were delays at other London airports. Cross-Channel ferries to France and Ireland were also cancelled and over 10 000 homes in south-west England lost their electricity.

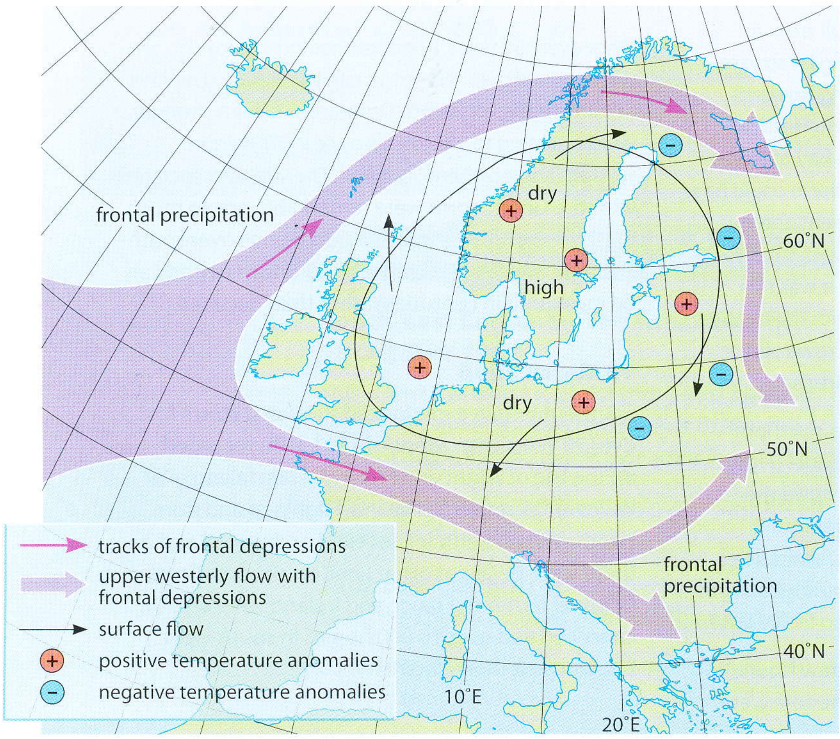
Anticyclones

An anticyclone is a large mass of subsiding air which produces an area of high pressure on the Earth's surface (Figure 9.47). The source of the air is the upper atmosphere, where amounts of water vapour are limited. On its descent, the air warms at the DALR (page 216), so dry conditions result. Pressure gradients are gentle, resulting in weak winds or calms (Figure 9.29b). The winds blow outwards and clockwise in the northern hemisphere. Anticyclones may be 3000 km in diameter – much larger than depressions – and, once established, can give several days or, under extreme conditions, several weeks, of settled weather. There are also differences, again unlike in a depression, between the expected weather conditions in a summer and a winter anticyclone.

Figure 9.47

Anticyclone over the British Isles





Weather conditions over Britain

Summer Due to the absence of cloud, there is intense insolation which gives hot, sunny days (up to 30°C in southern England) and an absence of rain. Rapid radiation at night, under clear skies, can lead to temperature inversions and the formation of dew and mist, although these rapidly clear the following morning. Coastal areas may experience advection fogs and land and sea breezes, while highlands have mountain and valley winds (pages 240–241). If the air has its source over North Africa – that is, if it is a *Tc* air mass (Figure 9.40) – then heatwave conditions tend to result. Often, after several days of increasing thermals, there is an increased risk of thunderstorms and the so-called Spanish plume (page 220).

Winter Although the sinking air again gives cloudless skies, there is little incoming radiation during the day due to the low angle of the sun. At night, the absence of clouds means low temperatures and the development of fog and frost. These may take a long time to disperse the next day in the weak sunshine. Polar continental (*Pc*) air (Figure 9.40), with its source in central Asia and a slow movement over the cold European land mass, is cold, dry and stable until it reaches the North Sea where its lower layers acquire some warmth and moisture. This can cause heavy snowfalls on the east coast (Figure 9.22).

Blocking anticyclones

These occur when cells of high pressure detach themselves from the major high pressure areas of the subtropics or poles (Figure 9.38b). Once created, they last for several days and 'block' eastward-moving depressions (Figure 9.48) to create anomalous conditions such as extremes of temperature, rainfall and sunshine – as in Britain in the summer of 1995 and the winter of 1987.

Figure 9.48

A blocking anticyclone over Scandinavia: the upper westerlies divide upwind of the block and flow around it with their associated rainfall; there are positive temperature anomalies within the southerly flow to the west of the block and negative anomalies to the east

Tropical cyclones

Tropical cyclones are systems of intense low pressure known locally as **hurricanes**, **typhoons** and **cyclones** (Figure 9.49). They are characterised by winds of extreme velocity and are accompanied by torrential rainfall – two factors that can cause widespread damage and loss of life (Places 31, page 238). As yet, there is still insufficient conclusive evidence as to the process of their formation, although knowledge has been considerably improved recently due to airflights through and over individual systems, and the use of weather satellites. Tropical cyclones tend to develop:

- over warm tropical oceans, where sea temperatures exceed 26°C and where there is a considerable depth of warm water
- in autumn, when sea temperatures are at their highest
- in the trade wind belt, where the surface winds warm as they blow towards the Equator
- between latitudes 5° and 20° north or south of the Equator (nearer to the Equator the Coriolis force is insufficient to enable the feature to 'spin' – page 225).

Once formed, they move westwards – often on erratic, unpredictable courses – swinging poleward on reaching land, where their energy is rapidly dissipated (Figure 9.49). They are another mechanism by which surplus energy is transferred away from the tropics (Figure 9.6).

Hurricanes

Hurricanes are the tropical cyclones of the Atlantic. They form after the ITCZ has moved to its most northerly extent enabling air to converge at low levels, and can have a diameter of up to 650 km. Unlike depressions, hurricanes occur when temperatures, pressure and humidity are uniform over a wide area in the lower troposphere for a lengthy period, and anticyclonic conditions exist in the upper troposphere. These conditions are essential for the development, near the Earth's surface, of intense low pressure and strong winds. To enable the hurricane to move, there must be a continuous source of heat to maintain the rising air currents. There must also be a large supply of moisture to provide the latent heat, released by condensation, to drive the storm and to provide the heavy rainfall. It is estimated that in a single day a hurricane can release an amount of energy equivalent to that released by 500 000 atomic bombs the size of the one dropped on Hiroshima in the Second World War. Only when the storm has reached maturity does the central **eye** develop. This is an area of subsiding air, some 30–50 km in diameter, with light winds, clear skies and anomalous high temperatures (Figure 9.50). The descending air increases instability by warming and exaggerates the storm's intensity.

The hurricane rapidly declines once the source of heat is removed, i.e. when it moves over colder water or a land surface; these increase friction and cannot supply sufficient moisture. The average lifespan of a tropical cyclone is 7 to 14 days. The characteristic weather conditions associated with the passage of a typical hurricane are shown diagrammatically in Figure 9.50, and from space in Figure 9.51.

Figure 9.49

Global location and mean frequency of tropical cyclones

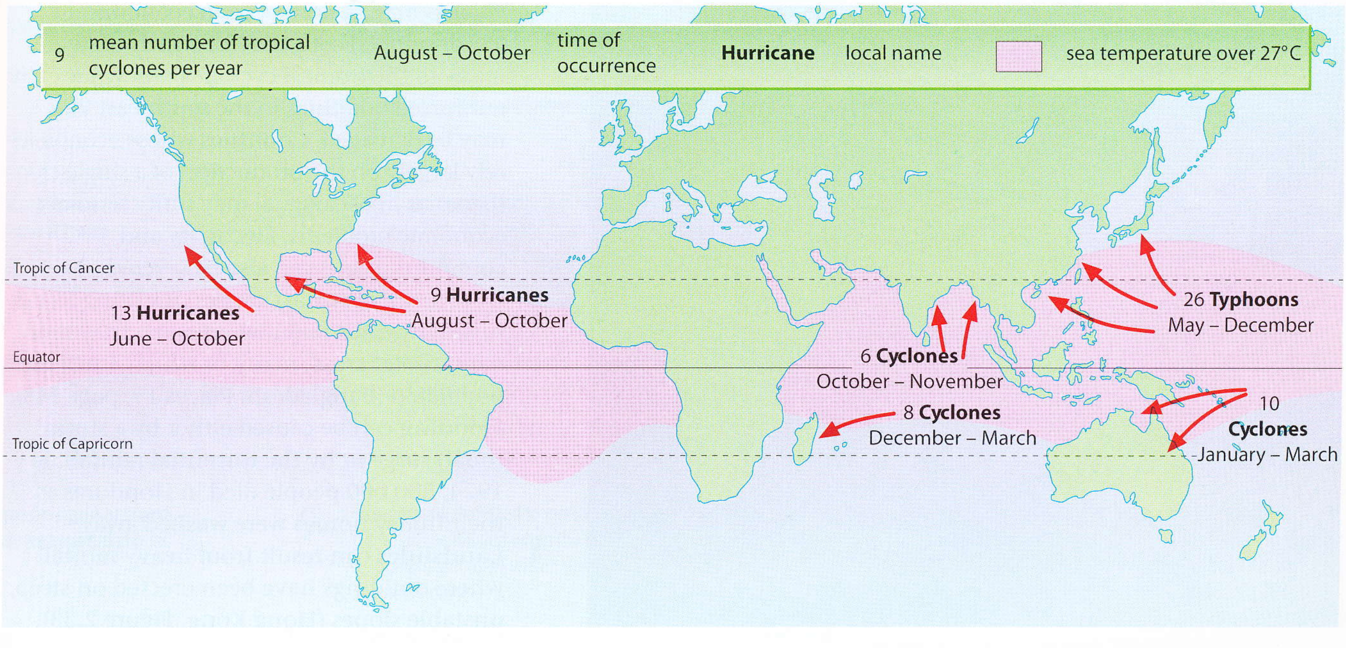
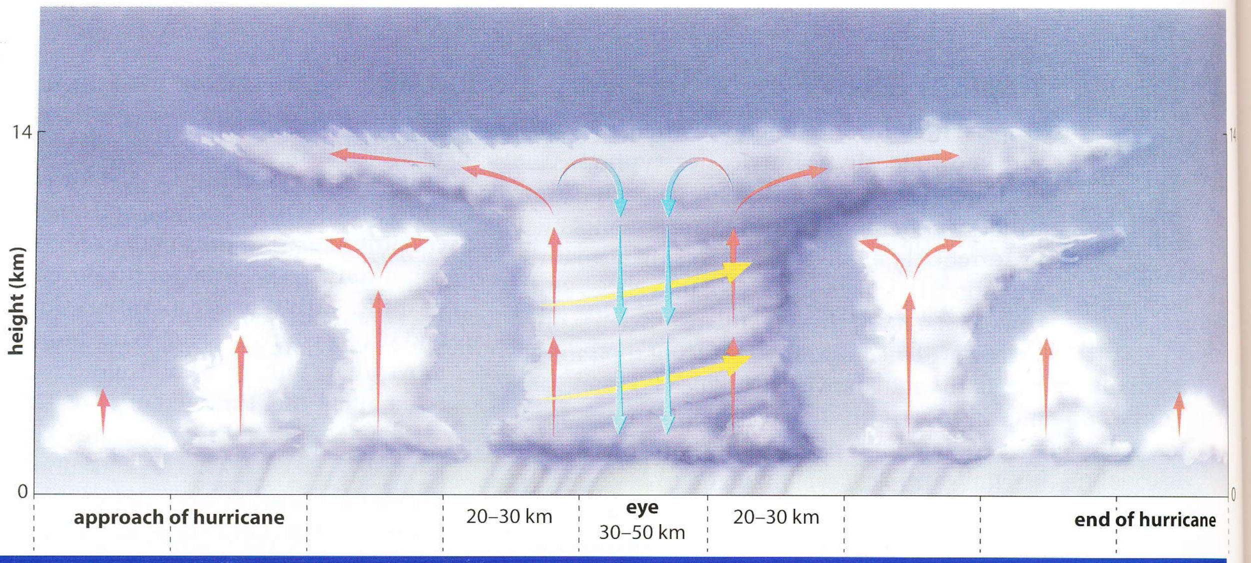


Figure 9.50

Weather associated with the passage of a hurricane or tropical cyclone



Vertical movement	updraughts increasing →		updraughts increasing		subsiding air	spiral uplift		updraughts decreasing →	
Clouds (Figure 9.20)	few Cu	Cu	Cu and some Cb	giant CB and Ci	none	giant Cb and Ci	Cu and some Cb	Cu	small Cu
Precipitation	none	showers	heavy showers	torrential rain 250 mm/day	none	torrential rain 250 mm/day	heavy showers	showers	none
Wind speed	gentle	fresh, gusty	locally very strong	hurricane force 160 km/hr	calm	hurricane force 160 km/hr	locally very strong	fresh gusty	gentle
Wind direction	NNW	NW	WNW	WNW	calm	SSE	SSE	SE	ESE
Temperatures (plus examples)	high (30°C)	still high (30°C)	falling (26°C)	low (24°C)	high (32°C)	low (24°C)	rising (26°C)	high (28°C)	high (30°C)
Pressure	average, 1012 mb	steady, 1010 mb	slowly falling, 1006 mb	rapid fall	low, 960 mb	rapid rise	slowly rising, 1004 mb	steady, 1010 mb	average, 1012 mb

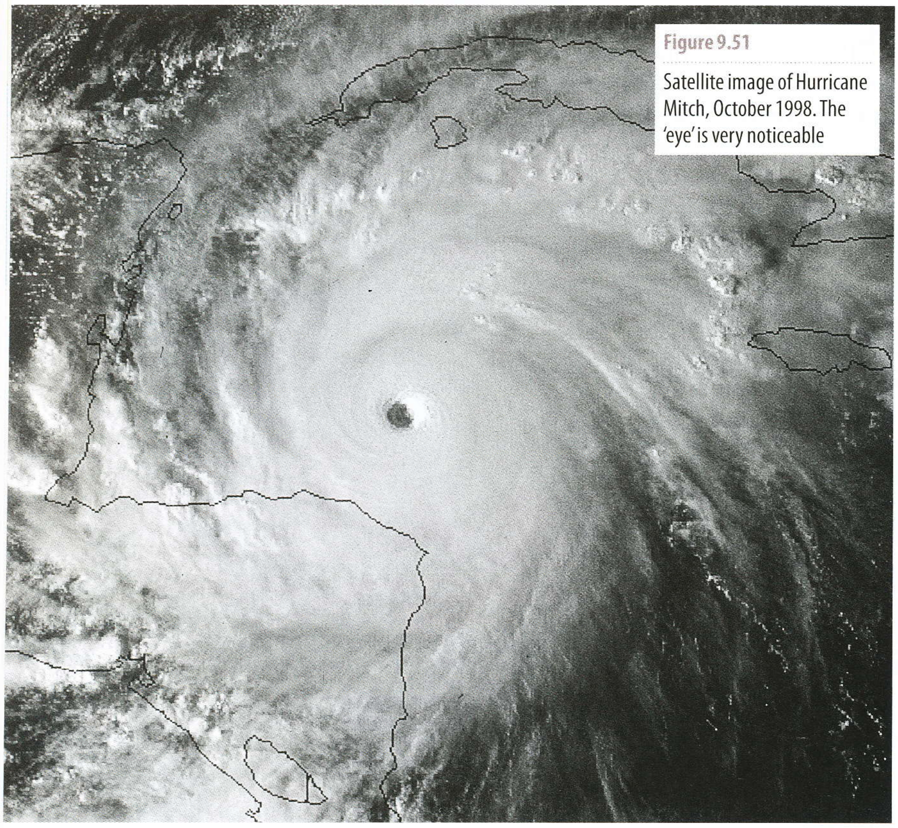


Figure 9.51

Satellite image of Hurricane Mitch, October 1998. The 'eye' is very noticeable

Tropical cyclones are a major natural hazard which often cause considerable loss of life and damage to property and crops (Places 31). There are four main causes of damage.

1 **High winds**, which often exceed 160 km/hr and, in extreme cases, 300 km/hr. Whole villages may be destroyed in economically less developed countries (of which there are many in the tropical cyclone belt), while even reinforced buildings in the south-east USA may be damaged. Countries whose economies rely largely on the production of a single crop (bananas in Nicaragua) may suffer serious economic problems. Electricity and communications can also be severed.
2 **Ocean storm (tidal) surges**, resulting from the high winds and low pressure, may inundate coastal areas, many of which are densely populated (Bangladesh, Places 19, page 148).
3 **Flooding** can be caused either by a storm (tidal) surge or by the torrential rainfall. In 1974, 800 000 people died in Honduras as their flimsy homes were washed away.
4 **Landslides** can result from heavy rainfall where buildings have been erected on steep, unstable slopes (Hong Kong, Figure 2.33).

Places 30 Hong Kong: typhoon warning, 1 May 1999

'The Number 8 signal may be raised today as Typhoon Leo moves closer to Hong Kong. Its approach forced the Hong Kong observatory to hoist the strong wind signal Number 3 yesterday afternoon [Figure 9.52] – the first time it had ever been raised in April [Figure 9.55]. Leo intensified into a typhoon yesterday, with central wind speeds of up to 130 km/hr. At midnight, it was 310 km south-south-east of Hong Kong, and was moving at about 8 km/hr [Figure 9.53]. The typhoon is expected to be closest to Hong Kong early tomorrow morning, by which time weather will deteriorate further and average rainfall could exceed 500 mm [Figure 9.54].

Figure 9.52

Typhoon warning system, Hong Kong

The Education Department has ordered kindergartens, schools for the mentally and physically handicapped, and nursing schools to remain closed. The Home Affairs Department's temporary shelters will open if Signal 8 goes up. People in need of shelter can make enquiries by calling the hotline.'

Source: South China Morning Post

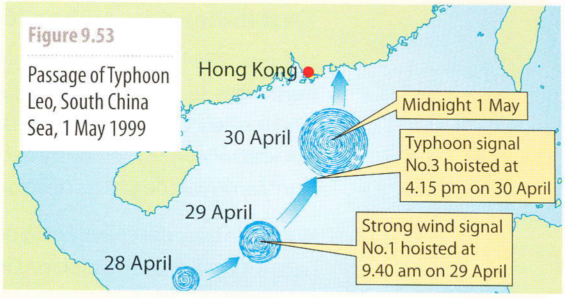


Figure 9.53

Passage of Typhoon Leo, South China Sea, 1 May 1999

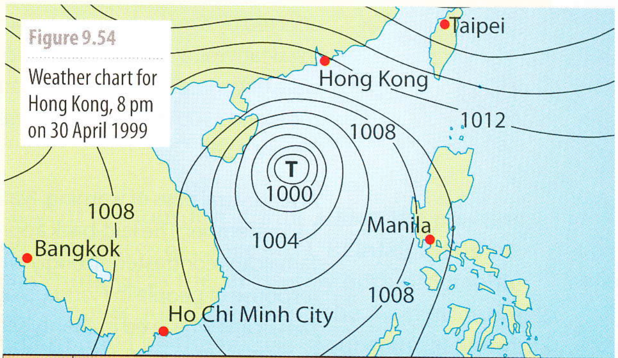


Figure 9.54

Weather chart for Hong Kong, 8 pm on 30 April 1999

°C	
	Saturday: Overcast with rain, heavy at times.
21 18	Wind: east to northeast force 6 to 7, becoming force 7 to 8 later.
	Sunday: Overcast with frequent heavy rain.
22 20	Wind: east to southeast force 7 to 8.
Coastal waters	
Hong Kong adjacent waters: East to northeast force 7 to 8, up to force 10 later today in the south. Occasional heavy rain. Rough to very rough seas becoming high.	

Figure 9.55

Typhoon warning signals, Hong Kong

Signal (four categories of tropical cyclone based on wind speed)		Meaning of the signal	What you should do Specific advice is contained in weather broadcasts, but the following general precautions can be taken
Stand by	1	A tropical cyclone is centred within about 800 km of Hong Kong. Hong Kong is placed on a state of alert because the tropical cyclone is a potential threat and may cause destructive winds later.	**Listen to weather broadcasts**. Some preliminary precautions are desirable and you should take the existence of the tropical cyclone into account in planning your activities.
Strong wind – A Tropical depression	3	Strong wind expected or blowing, with a sustained speed of 41–62 km/hr and gusts that may exceed 110 km/hr. The timing of the hoisting of the signal is aimed to give about 12 hours' advance warning of a strong wind in Victoria Harbour but the warning period may be shorter for more exposed waters.	**Take all necessary precautions**. Secure all loose objects, particularly on balconies and rooftops. Secure hoardings, scaffolding and temporary structures. Clear gutters and drains. Take full precautions for the safety of boats. Ships in port normally leave for typhoon anchorages or buoys. Ferry services may soon be affected by wind or waves. Even at this stage heavy rain accompanied by violet squalls may occur.
Gale or storm – B Tropical storm	4–8	Gale or storm expected or blowing, with a sustained wind speed of 63–117 km/hr from the quarter indicated and gusts that may exceed 180 km/hr. The timing of the replacement of the Strong Wind Signal No.3 by the appropriate one of these four signals, is aimed to give about 12 hours' advance warning of a gale in Victoria Harbour, but the sustained wind speed may reach 63 km/hr within a shorter period over more exposed waters. Expected changes in the direction of the wind will be indicated by corresponding changes of these signals.	**Complete all precautions** as soon as possible. It is extremely dangerous to delay precautions until the hoisting of No.9 or No.10 signals as these are signals of great urgency. Windows and doors should be bolted and shuttered. **Stay indoors** when the winds increase to avoid flying debris, but if you must go out, keep well clear of overhead wires and hoardings. All schools and law courts close and ferries will probably stop running at short notice. The sea-level will probably be higher than normal, particularly in narrow inlets. If this happens near the time of normal high tide then low-lying areas may have to be evacuated very quickly. Heavy rain may cause flooding, rockfalls and mudslips.
Increasing gale or storm – C Severe tropical storm	9	Gale or storm expected to increase significantly in strength. This signal will be hoisted when the sustained wind speed is expected to increase and come within the range 88–117 km/hr during the next few hours.	**Stay where you are** if reasonably protected and **away from exposed windows and doors**. These signals imply that the centre of a severe tropical storm or a typhoon will come close to Hong Kong. If the eye passes over there will be a lull lasting from a few minutes to some hours, but be prepared for a sudden resumption of destructive winds from a different direction.
Hurricane – D Typhoon	10	Hurricane-force winds expected or blowing, with a sustained wind speed reaching upwards from 118 km/hr and with gusts that may exceed 220 km/hr.	

Places 31 The West Indies and Myanmar: tropical storms

West Indies, September 2004

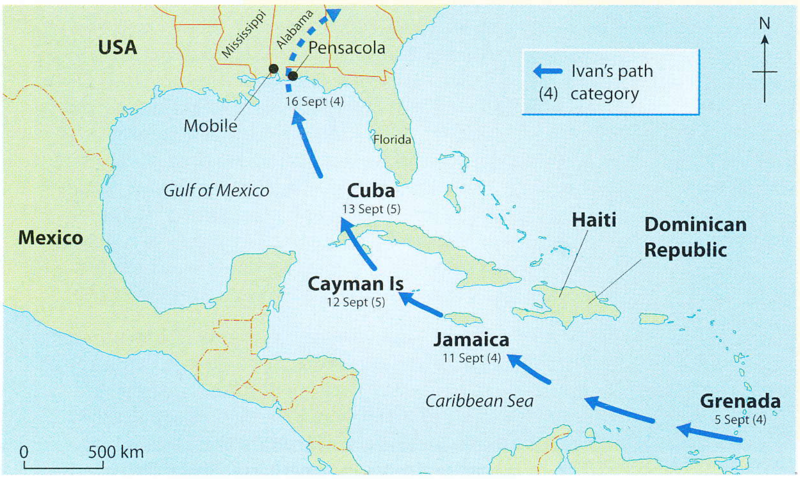
The year 2004 experienced the 'mother of hurricanes season'. Following hurricanes Charlie, which killed 16 people and caused damage in Florida only once previously exceeded, and Frances, Hurricane Ivan began its destructive course.

Hurricane Ivan, deservedly nicknamed 'the Terrible', began its trail of destruction on Grenada on 5 September – the first time the island had been affected by a major hurricane since 1955. Reports put the death toll at 34; water, electricity and air transport were disrupted for several days, and two-thirds of the island's 100 000 residents were made homeless (Figure 9.56).

After several days of warning, Ivan hit Jamaica on 11 September. The laid-back approach of many Jamaicans contrasted strongly with the well-practised response of people in Florida. Many of those Jamaicans who lived in shanty settlements refused to leave their flimsy, often makeshift homes, and only a few thousand of the half million ordered to evacuate heeded the government's warning, many preferring to protect what might be left of their possessions from post-hurricane looting. The resultant death toll was put at 20. By the time Ivan ravaged the Cayman Islands a day later, it had become a category 5 event – one of only a handful of that intensity in the last 100 years. Winds reached 260 km/hr while torrential rain and 6 m waves caused extensive flooding but, fortunately, no deaths were reported. In Cuba, next in Ivan's path, 2 million people were evacuated in advance of what was considered the most violent hurricane for over 50 years but at almost the last minute it veered sufficiently for the eye to pass just to the west of the island. Ivan, by now slightly reduced in strength, made landfall in the USA between Mobile (Alabama) and Pensacola (Florida) on 16 September, with wind speeds of 210 km/hr and a tidal surge of 4 m. Although 2 million people had been evacuated along a 675 km stretch of the Gulf coast, 12 deaths were reported. This might have been worse had Ivan veered westwards where parts of the Louisiana coast lie 3 m or more below sea-level and are protected by huge levées.

Figure 9.56

The path of Hurricane Ivan, September 2004



Myanmar, May 2008

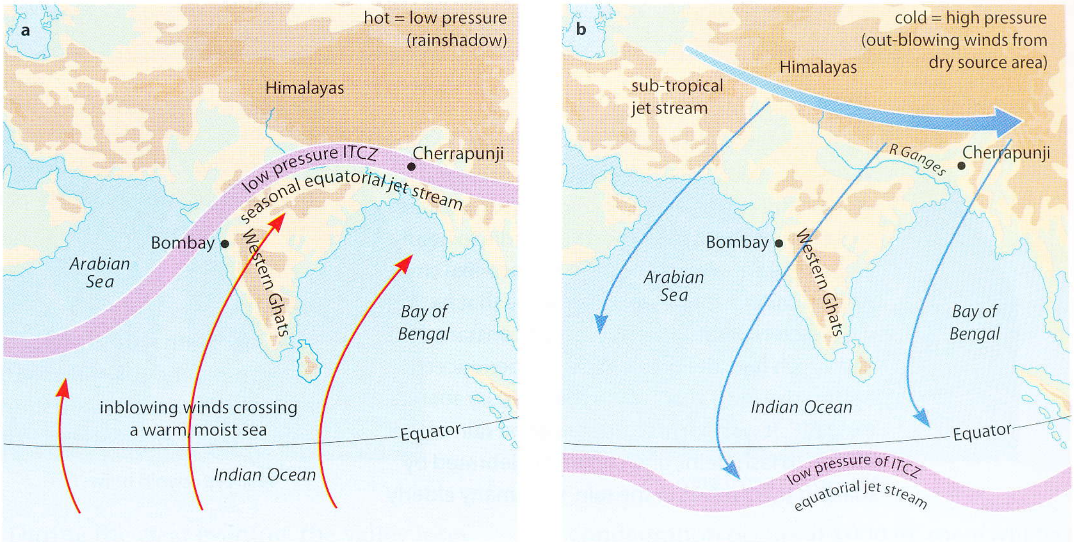
Bangladesh frequently experiences tropical cyclones which move northwards, accompanied by winds with speeds exceeding 200 km/hr, up the narrowing, shallowing Bay of Bengal. These cyclones can create storm surges of over 8 m that affect the flat delta region of the Ganges–Brahmaputra (Places 19, page 148). Improvements in coastal defences and early warning systems have reduced considerably the amount of damage and the number of deaths from 200 000 after the 1970 storm to 140 000 in 1990, 135 000 in 1991, 40 000 in 1994 and 10 000 in 1999. However, in 2008 tropical cyclone Nargis hit the still unprotected Irrawaddy delta lying to the south in Myanmar.

Little warning was given before Nargis, with wind speeds of 200 km/hr, swept over the flat Irrawaddy delta before affecting the former capital city of Rangoon. Unlike other recent catastrophes such as the Indian Ocean tsunami (Places 4) and the China earthquake (Places 2) where the world was immediately aware of the event, here, due to a lack of contact with the military regime, it was two days before news began to leak out of Myanmar and then only to admit to 350 deaths.

Later it became known that a tidal surge that followed the cyclone created devastation of tsunami proportions. Crops had been totally destroyed in the country's so-called rice bowl, as had coastal shrimp farms and fishing boats. Huge areas were left without fresh water, electricity or transport. Although the military junta made a rare appeal for help, outside aid workers were not to be allowed into the country and a week later many isolated areas had received no internal relief of any kind. By this time it was announced that the death toll was 22 000 with a further 40 000 missing in a declared disaster zone of 24 million people. Reports talked of flood waters receding to leave rotting, bloated bodies, both human and animal, reminiscent of the 2004 post-tsunami scenes. Indeed two weeks after Nargis hit the country and with overseas aid still being rejected, the UN suggested that up to 200 000 Burmese had either died at the time, afterwards through a failure to provide relief, or were unaccounted for – a figure close to that of the 2004 tsunami.

Figure 9.57

The monsoon in the Indian subcontinent



The monsoon

The word **monsoon** is derived from the Arabic word for 'a season', but the term is more commonly used in meteorology to denote a seasonal reversal of wind direction.

The major monsoon occurs in south-east Asia and results from three factors:

1. The extreme heating and cooling of large land masses in relation to the smaller heat changes over adjacent sea areas (page 210). This in turn affects pressure and winds.
2. The northward movement of the ITCZ (page 226) during the northern hemisphere summer.
3. The uplift of the Himalayas which, some 6 million years ago, became sufficiently high to interfere with the general circulation of the atmosphere (Places 5, page 20).

The south-west or summer monsoon

As the overhead sun appears to move northwards to the Tropic of Cancer in June, it draws with it the convergence zone associated with the ITCZ (Figure 9.57a). The increase in insolation over northern India, Pakistan and central Asia means that heated air rises, creating a large area of low pressure. Consequently, warm moist *Em* (equatorial maritime) and *Tm* air, from over the Indian Ocean, is drawn first northwards and then, because of the Coriolis force, is diverted north-eastwards (page 224). The air is humid, unstable and conducive to rainfall. Amounts of precipitation are most substantial on India's west coast, where the air rises over the Western Ghats, and on the windward slope of the Himalayas: Bombay has 2000 mm and Cherrapunji 13 000 mm in four summer months. The advent of monsoon storms allows the planting of rice (Places 67, page 481). Rainfall totals are accentuated as the air rises by both orographic and convectional uplift and the 'wet' monsoon is maintained by the release of substantial amounts of latent heat. The average arrival date is 10 May in Sri Lanka and 5 July at the Pakistan border – a time-lapse of seven weeks (Places 32).

The north-east or winter monsoon

During the northern winter, the overhead sun, the ITCZ and the subtropical jet stream all move southwards (Figure 9.57b). At the same time, central Asia experiences intense cooling which allows a large high pressure system to develop. Airstreams that move outwards from this high pressure area are dry because their source area is semi-desert. They become even drier as they cross the Himalayas and adiabatically warmer as they descend to the Indo-Gangetic plain. Bombay receives less than 100 mm of rain during these eight months. The south-west monsoon usually begins its retreat from the extreme north-west of India on 1 September and takes until 15 November, i.e. 11 weeks, to clear the southern tip.

The monsoon, which in reality is much more complex than the model described above, affects the lives of one-quarter of the world's population. Unfortunately, monsoon rainfall, especially in the Indian subcontinent, is unreliable (Figure 9.28). If the rains fail, then drought and famine ensue: 1987 was the ninth year in a decade when the monsoon failed in north-west India. If, on the other hand, there is excessive rainfall then large areas of land experience extreme flooding (Bangladesh in 1987, 1988 and 1998).

Places 32 Delhi: the monsoon climate

June

'Rain brought welcome relief to the Indian capital yesterday, a day after 18 people collapsed and died on the streets in the blistering heat, pushing the summer death toll in northern India to nearly 350. Heavy showers cooled the furnace-like city, reeling under a three-week heatwave that has kept daytime temperatures at an almost constant 45°C and which had, the previous day, experienced its hottest day in 50 years when the mercury soared to 42.6°C. It was the first pre-monsoon rain of the season to lash Delhi, and children celebrated by soaking themselves in the rain, with many elderly citizens joining them in the belief that monsoon rains help cure blisters and skin diseases caused by extreme heat. More thunderstorms are expected by the weekend, which should mark the onset of the summer monsoon.'

July

'The July death toll from relentless monsoon rains across India and Pakistan rose to more than 590 as several waves of severe storms passed across the subcontinent. Many streets in Delhi are still under water.'

Meso-scale: local winds

Of the three meso-scale circulations described here, two – **land and sea breezes** and **mountain and valley winds** – are caused by local temperature differences; the third – the **föhn** – results from pressure differences on either side of a mountain range.

The land and sea breeze

This is an example, on a diurnal timescale, of a circulation system resulting from differential heating and cooling between land surfaces and adjacent sea areas. The resultant pressure differences, although small and localised, produce gentle breezes which affect coastal areas during calm, clear anticyclonic conditions. When the land heats up rapidly each morning, lower pressure forms and a gentle breeze begins to blow from the sea to the land (Figure 9.58a). By early afternoon, this breeze has strengthened sufficiently to bring a freshness which, in the tropics particularly, is much appreciated by tourists at the beach resorts. Yet by sunset, the air and sea are both calm again.

Although the circulation cell rarely rises above 500 m in height or reaches more than 20 km inland in Britain, the sea breeze is capable of lowering coastal temperatures by 15°C and can produce advection fogs such as the 'sea-fret' or 'haar' of eastern Britain.

At night, when the sea retains heat longer than the land, there is a reversal of the pressure gradient and therefore of wind direction (Figure 9.58b). The land breeze, the gentler of the two, begins just after sunset and dies away by sunrise.

The mountain and valley wind

This wind is likely to blow in mountainous areas during times of calm, clear, settled weather. During the morning, valley sides are heated by the sun, especially if they are steep, south-facing (in the northern hemisphere) and lacking in vegetation cover. The air in contact with these slopes will heat, expand and rise (Figure 9.59a), creating a pressure gradient. By 1400 hours, the time of maximum heating, a strong uphill or **anabatic wind** blows up the valley and the valley sides – ideal conditions for hang-gliding! The air becomes conditionally unstable (Figure 9.19), often producing cumulus cloud and, under very warm conditions, cumulo-nimbus with the possibility of thunderstorms on the mountain ridges. A compensatory sinking of air leaves the centre of the valley cloud-free.

Figure 9.58

Land and sea breezes in Britain

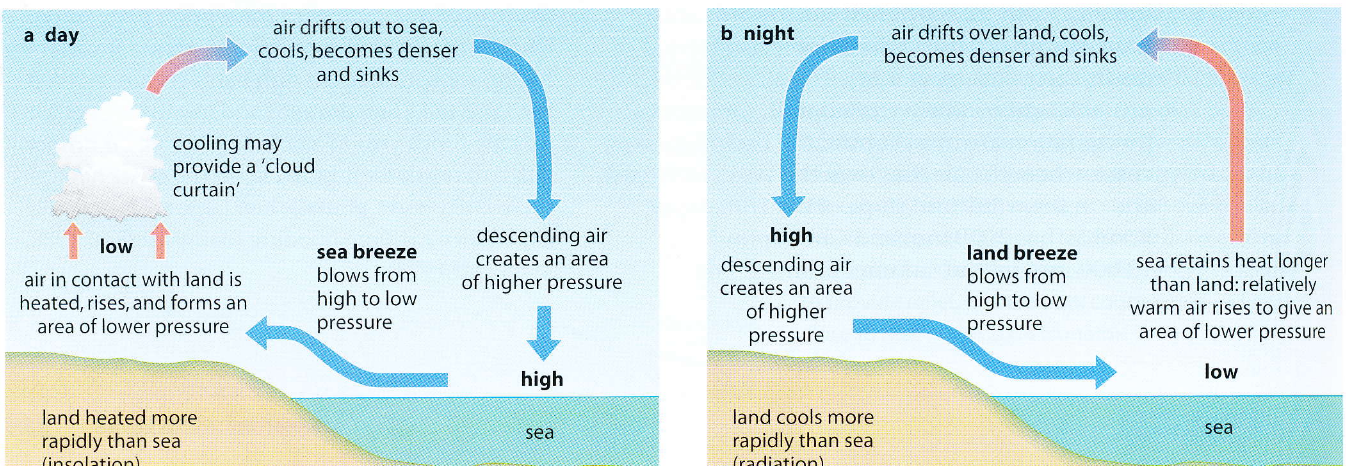
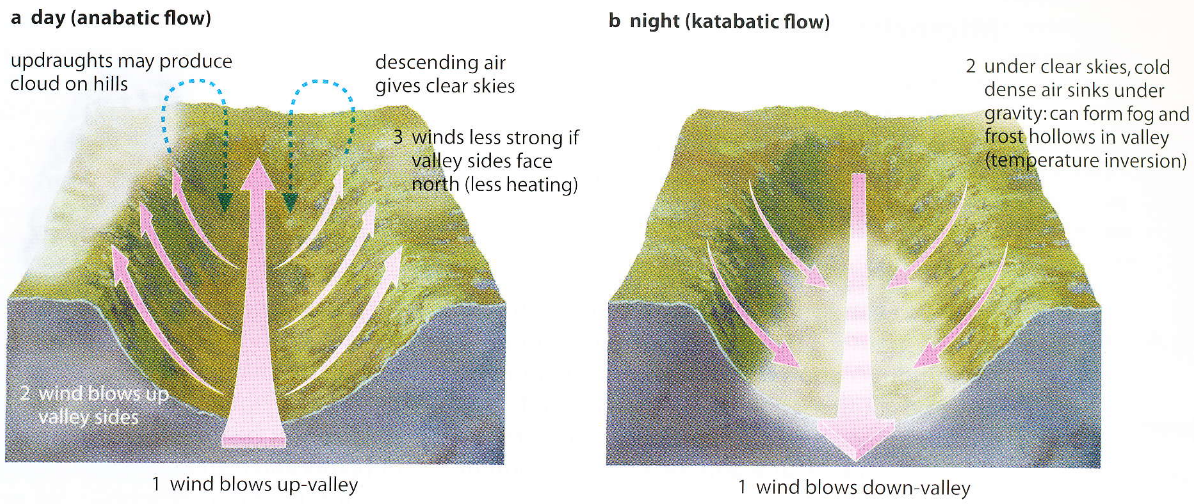


Figure 9.59

Mountain and valley winds



During the clear evening, the valley loses heat through radiation. The surrounding air now cools and becomes denser. It begins to drain, under gravity, down the valley sides and along the valley floor as a mountain wind or **katabatic wind** (Figure 9.59b). This gives rise to a temperature inversion (Figure 9.24) and, if the air is moist enough, in winter may create fog (Figure 9.23) or a **frost hollow**. Maximum wind speeds are generated just before dawn, normally the coldest time of the day. Katabatic winds are usually gentle in Britain, but are much stronger if they blow over glaciers or permanently snow-covered slopes. In Antarctica, they may reach hurricane force.

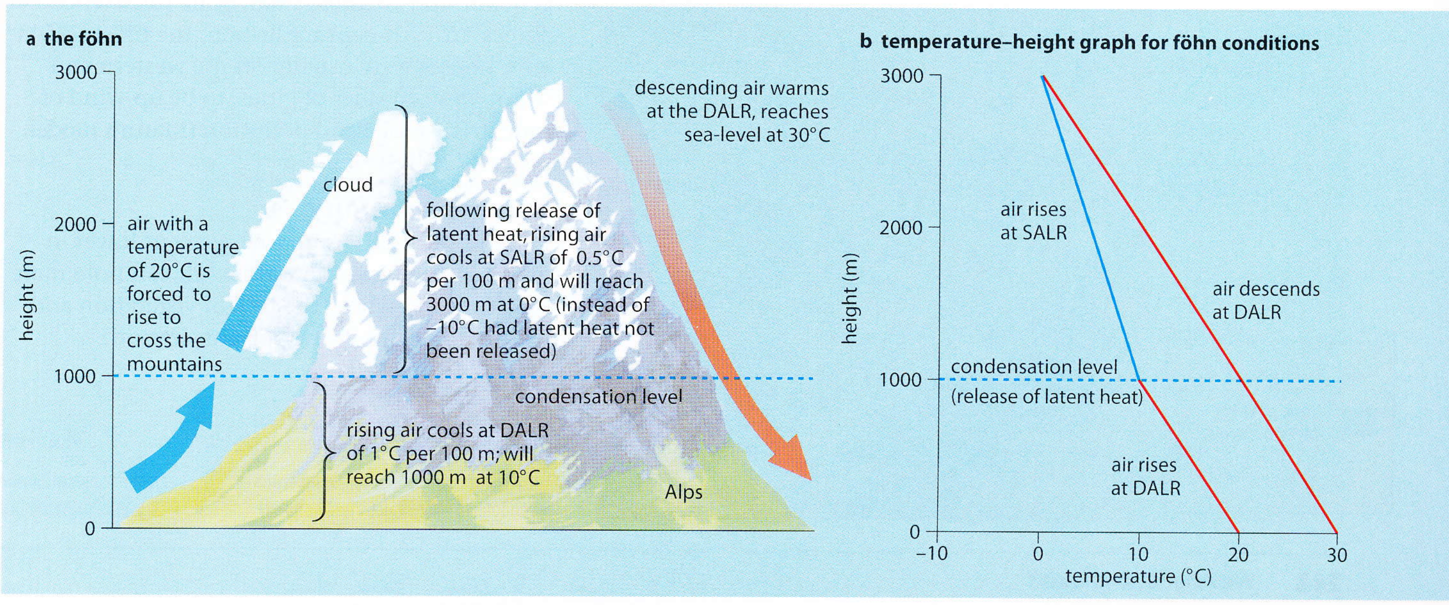
The föhn

The föhn is a strong, warm and dry wind which blows periodically to the lee of a mountain range. It occurs in the Alps when a depression passes to the north of the mountains and draws in warm, moist air from the Mediterranean. As the air rises (Figure 9.60), it cools at the DALR of 1°C per 100 m (page 216). If, as in Figure 9.60, condensation occurs at 1000 m, there will be a release of latent heat and the rising air will cool more slowly at the SALR of 0.5°C per 100 m. This means that when the air reaches 3000 m it will have a temperature of 0°C instead of the –10°C had latent heat not been released. Having crossed the Alps, the descending air is compressed and warmed at the DALR so that, if the land drops sufficiently, the air will reach sea-level at 30°C. This is 10°C warmer than when it left the Mediterranean. Temperatures may rise by 20°C within an hour and relative humidity can fall to 10 per cent.

This wind, also known as the **chinook** on the American Prairies, has considerable effects on human activity. In spring, when it is most likely to blow, it lives up to its Native American name of 'snow-eater' by melting snow and enabling wheat to be sown; and in Switzerland it clears the alpine pastures of snow. Conversely, its warmth can cause avalanches, forest fires and the premature budding of trees (Case Study 4a).

Figure 9.60

The föhn



Microclimates

Microclimatology is the study of climate over a small area. It includes changes resulting from the construction of large urban centres as well as those existing naturally between different types of land surface, e.g. forests and lakes.

Urban climates

Large cities and conurbations experience climatic conditions that differ from those of the surrounding countryside. They generate more dust and condensation nuclei than natural environments; they create heat; they alter the chemical composition and the moisture content of the air above them; and they affect both the albedo and the flow of air. Urban areas therefore have distinctive climates.

Temperature

Although tower blocks cast more shadow, normal building materials tend to be non-reflective and so absorb heat during the daytime. Dark-coloured roofs, concrete or brick walls and tarmac roads all have a high thermal capacity which means that they are capable of storing heat during the day and releasing it slowly during the night. Further heat is obtained from car fumes, factories, power stations, central heating and people themselves. The term **urban heat island** acknowledges that, under calm conditions, temperatures are highest in the more built-up city centre and decrease towards the suburbs and open countryside (Figure 9.61). In urban areas:

- daytime temperatures are, on average, 0.6°C higher
- night-time temperatures may be 3° or 4°C higher as dust and cloud act like a blanket to reduce radiation and buildings give out heat like storage radiators
- the mean winter temperature is 1° to 2°C higher (rural areas are even colder when snow-covered as this increases their albedo)
- the mean summer temperature may be 5°C higher
- the mean annual temperature is higher by between 0.6°C in Chicago and 1.3°C in London compared with that of the surrounding area.

Note how, in Figure 9.61, temperatures not only decrease towards London's boundary but also beside the Thames and Lea rivers. The urban heat island explains why large cities have less snow, fewer frosts, earlier budding and flowering of plants and a greater need, in summer, for air-conditioning than neighbouring rural areas.

Sunlight

Despite having higher mean temperatures, cities receive less sunshine and more cloud than their rural counterparts. Dust and other particles may absorb and reflect as much as 50 per cent of insolation in winter, when the sun is low in the sky and has to pass through more atmosphere, and 5 per cent in summer. High-rise buildings also block out light (Figure 9.62).

Wind

Wind velocity is reduced by buildings which create friction and act as windbreaks. Urban mean annual velocities may be up to 30 per cent lower than in rural areas and periods of calm may be 10–20 per cent more frequent. In contrast, high-rise buildings, such as the skyscrapers of New York and Hong Kong (Figure 9.62), form 'canyons' through which wind may be channelled. These winds may be strong enough to cause tall buildings to sway and pedestrians to be blown over and troubled by dust and litter. The heat island effect may cause local thermals and reduce the wind chill factor. It also tends to generate considerable small-scale turbulence and eddies. In 19th-century Britain, the most sought-after houses were usually on the western and south-western sides of cities, to be up-wind of industrial smoke and pollution (Mann's model, pages 422–423).

Relative humidity

Relative humidity is up to 6 per cent lower in urban areas where the warmer air can hold more moisture and where the lack of vegetation and water surface limits evapotranspiration.

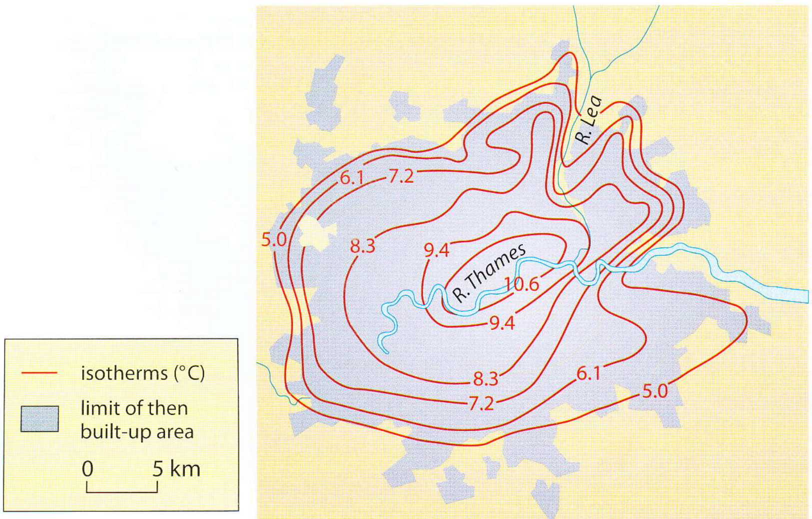


Figure 9.61

An urban heat island: minimum temperatures over London, 14 May 1959 (*after* Chandler)

Figure 9.62

Narrow streets with high-rise buildings are more likely to develop microclimates than those that are wider and have lower buildings; New York City

Cloud

Urban areas appear to receive thicker and up to 10 per cent more frequent cloud cover than rural areas. This may result from convection currents generated by the higher temperatures and the presence of a larger number of condensation nuclei.

Precipitation

The mean annual precipitation total and the number of days with less than 5 mm of rainfall are both between 5 and 15 per cent greater in major urban areas. Reasons for this are the same as for cloud formation. Strong thermals increase the likelihood of thunder by 25 per cent and the occurrence of hail by up to 400 per cent. The higher urban temperatures may turn the snow of rural areas into sleet and limit, by up to 15 per cent, the number of days with snow lying on the ground. On the other hand, the frequency, length and intensity of fog, especially under anticyclonic conditions, is much greater – there may up to 100 per cent more in winter and 25 per cent more in summer, caused by the concentration of condensation nuclei (Figure 9.63).

Thick fog is continuing to cause travel chaos among those looking forward to spending Christmas abroad. Over the last few days, thousands of passengers have experienced severe delays or cancellations of flights at numerous UK airports.

Yesterday 350 flights, 40 per cent of the total, were cancelled from Heathrow alone and, with fog set to remain today, British Airways has already decided to cancel all domestic flights to and from that airport. The problem with fog is that it means, for safety reasons, the distance between aircraft on approach to runways has to be doubled, thus reducing the number of landings.

22 December 2006

Figure 9.63

Fog causes Christmas chaos

Atmospheric composition

There may be three to seven times more dust particles over a city than in rural areas. Large quantities of gaseous and solid impurities are emitted into urban skies by the burning of fossil fuels, by industrial processes and from car exhausts. Urban areas may have up to 200 times more sulphur dioxide and 10 times more nitrogen oxide (the major components of acid rain) than rural areas, as well as 10 times more hydrocarbons and twice as much carbon dioxide. These pollutants tend to increase cloud cover and precipitation, cause smog (Figure 9.25), give higher temperatures and reduce sunlight.

Forest and lake microclimates

Different land surfaces produce distinctive local climates. Figure 9.64 summarises and compares some of the characteristics of microclimates found in forests and around lakes. As with urban climates, research and further information are still needed to confirm some of the statements.

Microclimate feature	Forest (coniferous and deciduous)	Water surface (lake, river)
Incoming radiation and albedo	Much incoming radiation is absorbed and trapped. Albedo for coniferous forest is 15%; deciduous 25% in summer and 35% in winter; and desert scrub 40%.	Less insolation absorbed and trapped. Albedo may be over 60%, i.e. higher than over seas/oceans (page 207). Higher on calm days.
Temperature	Small diurnal range due to blanket effect of canopy. Forest floor is protected from direct sunlight. Some heat lost by evapotranspiration.	Small diurnal range because water has a higher specific heat capacity. Cooler summers and milder winters. Lakesides have a longer growing season.
Relative humidity	Higher during daytime and in summer, especially in deciduous forest. Amount of evapotranspiration depends on length of day, leaf surface area, wind speed, etc.	Very high, especially in summer when evaporation rates are also high.
Precipitation	Heavy rain can be caused by high evapotranspiration rates, e.g. in tropical rainforests. On average, 30–35% of rain is intercepted: more in deciduous woodland in summer than in winter.	Air is humid. If forced to rise, air can be unstable and produce cloud and rain. Amounts may not be great due to fewer condensation nuclei. Fogs form in calm weather.
Wind speed and direction	Trees reduce wind speeds, especially at ground level. (They are often planted as windbreaks.) Trees can produce eddies.	Wind may be strong due to reduced friction. Large lakes (e.g. L. Victoria) can create land and sea breezes (page 240).

Figure 9.64

Microclimates of forests and water surfaces

Weather maps and forecasting in Britain

A weather map or **synoptic chart** shows the weather for a particular area at one specific time (Figures 9.67 and 9.68). It is the result of the collection and collation of a considerable amount of data at numerous weather stations, i.e. from a number of sample points (Framework 6, page 159). These data are then refined, usually as quickly as possible and now using computers, and are plotted using internationally accepted weather symbols. A selection of these symbols is shown in Figure 9.65. Weather maps are produced for different purposes and at various scales.

1 The daily weather map, as seen on television or in a national newspaper, aims to give a clear, but highly simplified, impression of the weather.
2 At a higher level, a synoptic map shows selected meteorological characteristics for specific **weather stations**. The **station model** in Figure 9.66 shows six elements: temperature, pressure, cloud cover, present weather (e.g. type of precipitation), wind direction and wind speed.
3 At the highest level, the Meteorological Office produces maps showing finite detail, e.g. amounts of various types of cloud at low, medium and high levels, dew point temperatures, barometric tendency (i.e. trends of pressure change), etc.

The role of the weather forecaster is to try to determine the speed and direction of movement of various air masses and any associated fronts, and to try to predict the type of weather these movements will bring. Forecasters now make considerable use of **satellite images** (Figures 9.67 and 9.68). Satellite images are photos taken by weather satellites as they continually orbit the Earth. These photos, which are relayed back to Earth, are invaluable in the prediction of short-term weather trends. Although forecasting is increasingly assisted by information from satellites, radar and computers, which show upper air as well as surface air conditions in a three-dimensional model, the complexity and unpredictability of the atmosphere can still catch the forecaster by surprise (Places 29, page 232). Part of this problem is related to the fact that meteorological information is a sample (Framework 6, page 159) rather than a total picture of the atmosphere, and so there is always a risk of the anomaly becoming the reality.

Figure 9.65

Weather symbols for cloud, precipitation, wind speed, temperature, pressure and wind direction

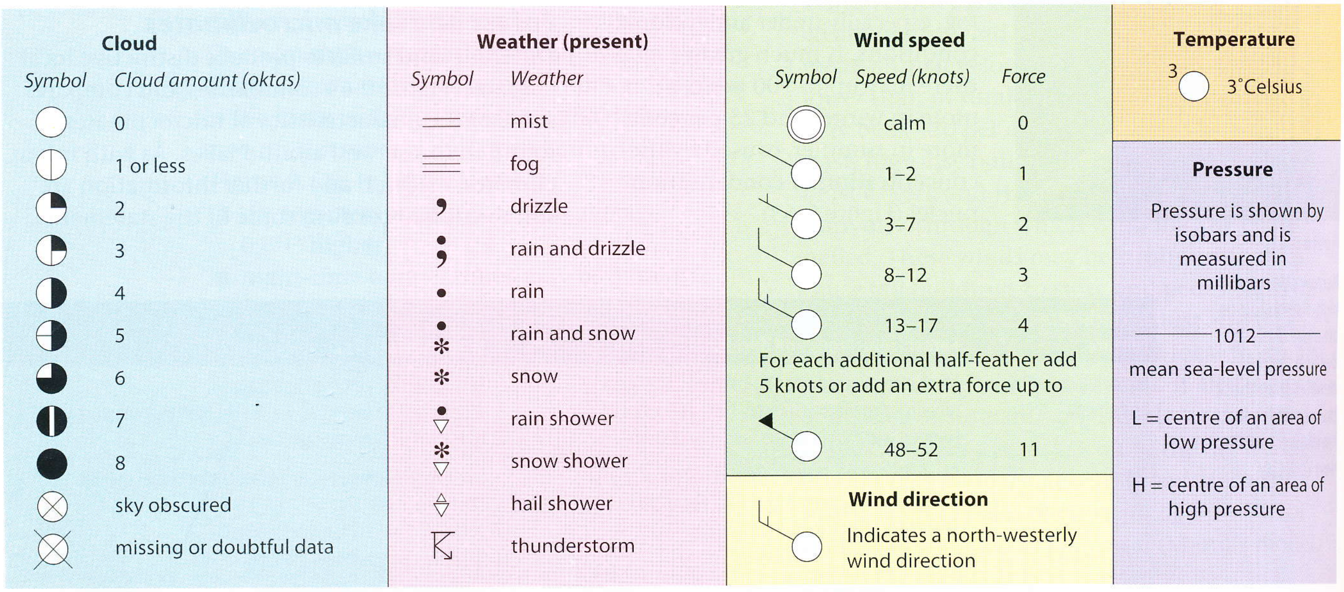


Figure 9.66

A weather station model and an example

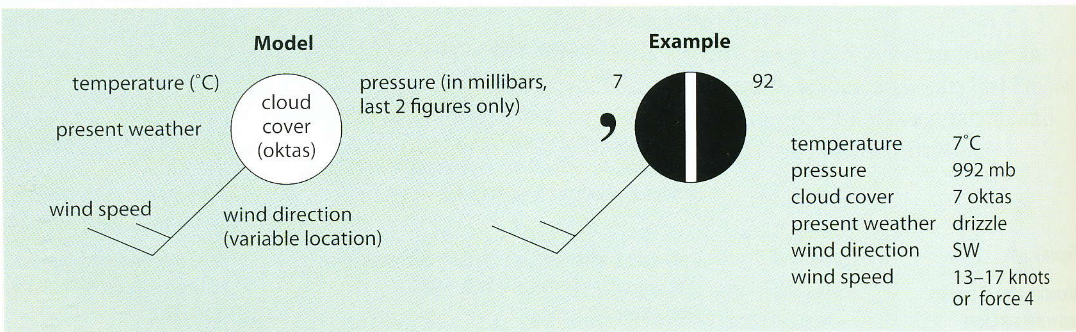
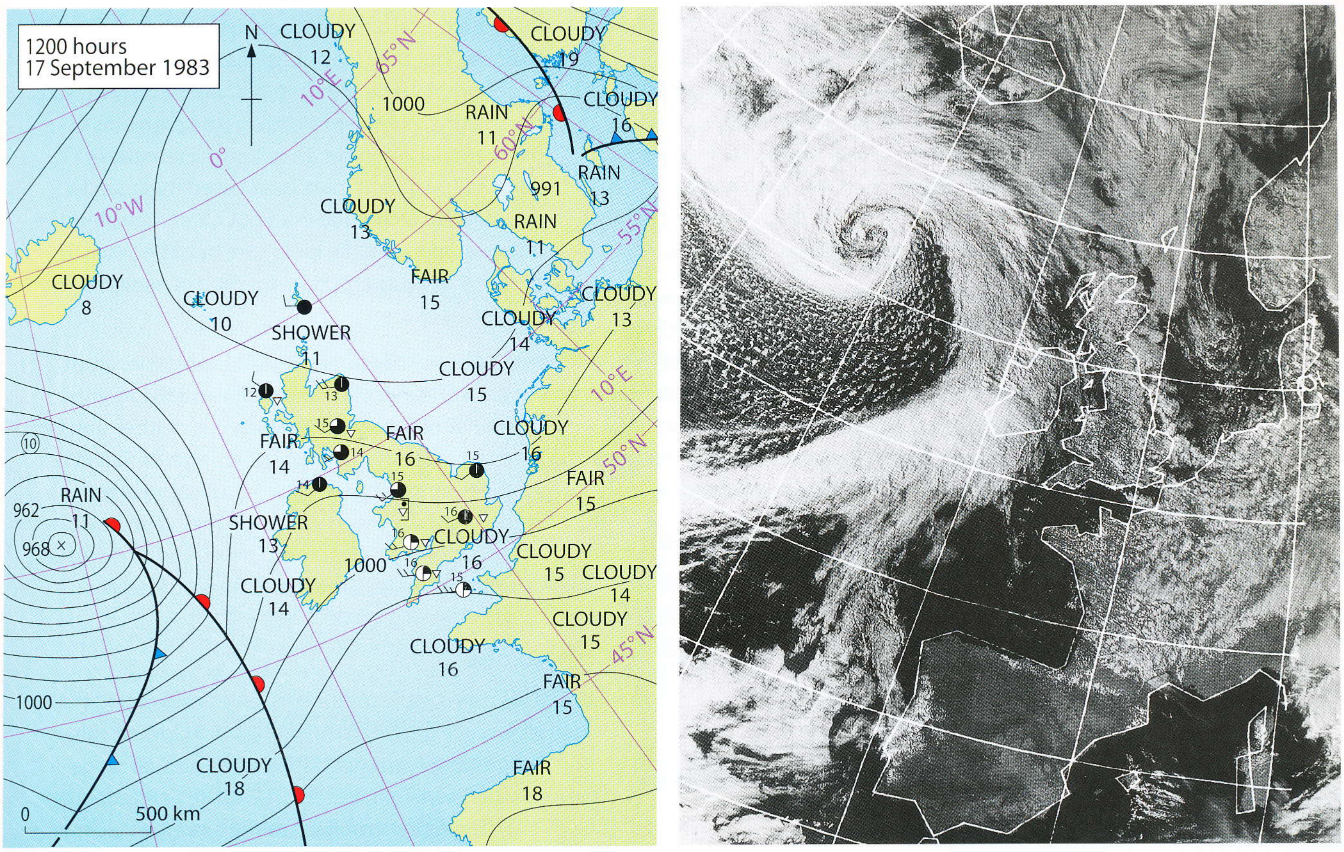


Figure 9.67

Synoptic chart and satellite image, 17 September 1983

Figure 9.67 shows the synoptic chart (weather map) and satellite (infra-red) image of a depression approaching the British Isles (compare Figure 9.43). Figure 9.68 shows the same depression 24 hours later, by which time it had passed over the British Isles (compare Figure 9.44).



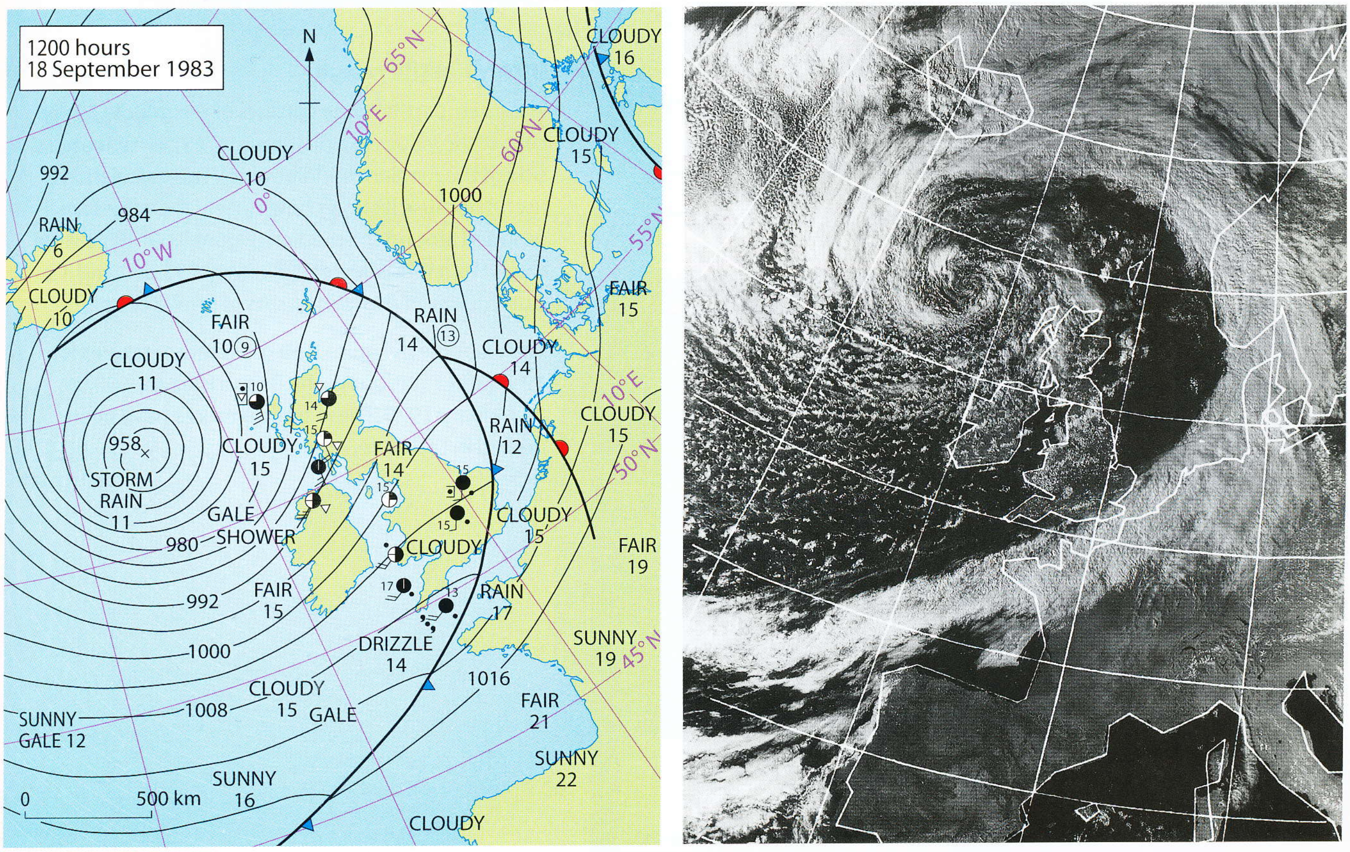


Figure 9.68

Synoptic chart and satellite

Framework 8 Measures of dispersion

Throughout this chapter on weather and climate, mean climatic figures have been quoted. To build up these pictures of global, regional and local climate patterns, statistics have been obtained by averaging readings, usually for temperature and precipitation, over a 30-year timescale. However, these averages themselves are often not as significant as the range or the degree to which they vary from, or are dispersed about, the mean.

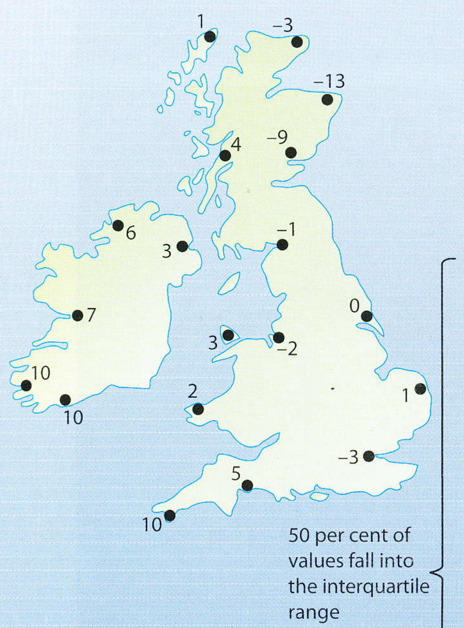
For example, two tropical weather stations may have equal annual rainfall totals when measured over 30 years. Station A may lie on the Equator and experience reliable rainfall with little variation from one year to the next. Station B may experience a monsoon climate where in some years the rains may fail entirely while in others they cause flooding.

The measure of dispersion from the mean can be obtained by using any one of three statistical techniques:

- the range
- the interquartile range, or
- the standard deviation.

Figure 9.69

The interquartile range



Rank	Temperatures 0°C (ranked)	
	10	
	10	
	10	
	7	
5	6	upper quartile
	5	
	4	
	3	
	3	
10	2	median (middle quartile)
	1	
	1	
	0	
	−1	
15	−2	lower quartile
	−3	
	−3	
	−9	
	−13	

These techniques are included here because meteorological data both require and benefit from their use, but they may be applied to most branches of geography where there is a danger that the mean, taken alone, may be misleading (the problems of overgeneralisation are discussed in Framework 11, page 347). Again, it must be stressed that use of a quantitative technique does not guarantee objective interpretation of data: great care must be taken to ensure that an appropriate method of manipulating the data is chosen.

It has already been seen how it is possible, given a data set, to calculate the mean and the median (Framework 5, page 112). However, neither statistic gives any idea of the spread, or range, of that data. As the example above of two tropical weather stations shows, mean values on their own give only part of the full picture. The spread of the data around the mean should also be considered.

Range

This very simple method involves calculating the difference between the highest and lowest values of the sample population, e.g. the annual range in temperature for London is 14°C (July 18°C, January 4°C). The range emphasises the extreme values and ignores the distribution of the remainder.

Interquartile range

The interquartile range consists of the middle 50 per cent of the values in a distribution, 25 per cent each side of the median (middle value). This calculation is useful because it shows how closely the values are grouped around the median (Figure 9.69). It is easy to calculate; it is unaffected by extreme values; and it is a useful way of comparing sets of similar data.

The example in Figure 9.69 gives temperatures for 19 weather stations in the British Isles at 0600 on 14 January 1979. These temperatures have been ranked in the table.

The upper quartile (*UQ*) is obtained by using the formula:

$$UQ = \left(\frac{n+1}{4}\right) \qquad \text{i.e.} \quad \left(\frac{19+1}{4}\right) = 5$$

This means that the *UQ* is the fifth figure from the top of the ranking order, i.e. 6°C. The lower quartile (*LQ*) is found by using a slightly different formula:

$$LQ = \left(\frac{n+1}{4}\right) \times 3 \qquad \text{i.e.} \quad \left(\frac{19+1}{4}\right) \times 3 = 15$$

This shows the *LQ* to be the 15th figure in the ranking order, i.e. –2°C. You will notice that the middle quartile is the same as the median. The interquartile range is the difference between the upper and lower quartiles, i.e. 6°C – –2°C = 8°C.

Another measure of dispersion, the quartile deviation, is obtained by dividing the interquartile range by two, i.e. 8°C ÷ 2 = 4°C

The smaller the interquartile range, or quartile deviation, the greater the grouping around the median and the smaller the dispersion or spread.

Standard deviation

This is the most commonly used method of measuring dispersion and although it may involve lengthy calculations it can be used with the arithmetic mean and it removes extreme values. The formula for the standard deviation is:

$$\sigma = \sqrt{\frac{\Sigma (x - \bar{x})^2}{n}}$$

where: σ = standard deviation

x = each value in the data set

$\bar{x}$ = mean of all values in the data set, and

n = number of values in the data set.

Let us suppose that the minimum temperatures for 10 weather stations in Britain on a winter's day were, in °C, 5, 8, 3, 2, 7, 9, 8, 2, 2 and 4. The standard deviation of this data set is worked out in Figure 9.70, proceeding as follows:

1 Find the mean ($\bar{x}$).

2 Subtract the mean from each value in the set: $x - \bar{x}$.

3 Calculate the square of each value in 2, to remove any minus signs: $(x - \bar{x})^2$.

4 Add together all the values obtained in 3: $\Sigma (x - \bar{x})^2$.

5 Divide the sum of the values in 4 by *n*: $\frac{\Sigma (x - \bar{x})^2}{n}$

6 Take the square root of the value obtained in 5 to obtain the standard deviation: $\sqrt{\frac{\Sigma (x - \bar{x})^2}{n}}$

The resulting standard deviation of $\sigma = 2.65$ is a low value, indicating that the data are closely grouped around the mean.

Figure 9.70

Finding the standard deviation

Minimum temperatures for 10 weather stations in Britain on a winter's day

The mean of 5, 8, 3, 2, 7, 9, 8, 2, 2, 4:

$$\bar{x} = \frac{50}{10} = 5$$

Weather station	Temperature at each station (x)	$x - \bar{x}$	$(x - \bar{x})^2$
1	5	5 – 5 = 0	0
2	8	8 – 5 = 3	9
3	3	3 – 5 = –2	4
4	2	2 – 5 = –3	9
5	7	7 – 5 = 2	4
6	9	9 – 5 = 4	16
7	8	8 – 5 = 3	9
8	2	2 – 5 = –3	9
9	2	2 – 5 = –3	9
10	4	4 – 5 = –1	1
			$\Sigma(x - \bar{x})^2 = 70$

$$\sigma = \sqrt{\frac{70}{10}}$$

$$\sigma = \sqrt{7} \qquad \therefore \text{ standard deviation} = 2.65$$

Climatic change

Climates have changed and still are constantly changing at all scales, from local to global, and over varying timespans, both long-term and short-term (Case Studies 9A and 9B). However, there have been surges of change over time which meteorologists and earth scientists are continually trying to clarify and explain.

Evidence of past climatic changes

- **Rocks** are found today which were formed under climatic conditions and in environments that no longer exist (Figure 1.1). In Britain, for example, coal was formed under hot, wet tropical conditions; sandstones were laid down during arid times; various limestones accumulated on the floors of warm seas; and glacial deposits were left behind by retreating ice sheets.
- **Fossil landscapes** exist, produced by certain geomorphological processes which no longer operate. Examples include glacially eroded highlands in north and west Britain (Chapter 4), granite tors on Dartmoor (page 202) and wadis formed during wetter periods (pluvials) in deserts (Places 25, page 188).
- Evidence exists of **changes in sea-level** (both isostatic as on Arran – Places 23, page 166) and eustatic (as at present in the Maldives – page 169) and changes in **lake levels** (Sahara, Figure 7.27).
- **Vegetation belts** have shifted through some 10° of latitude, e.g. changes in the Sahara Desert (Figure 7.27).
- **Pollen analysis** shows which plants were dominant at a given time. Each plant species has a distinctively-shaped pollen grain. If these grains land in an oxygen-free environment, such as a peat bog, they resist decay. Although pollen can be transported considerable distances by the wind and by wildlife, it is assumed that grains trapped in peat form a representative sample of the vegetation that was growing in the surrounding area at a given time; also, that this vegetation was a response to the climatic conditions prevailing at that time. Vertical sections made through peat show changes in pollen (i.e. vegetation), and these changes can be used as evidence of climatic change (the vegetation–climatic timescale in Figures 11.18 and 11.19).
- **Dendrochronology**, or tree-ring dating, is the technique of obtaining a core from a tree-trunk and using it to determine the age of the tree. Tree growth is rapid in spring, slower by the autumn and, in temperate latitudes, stops in winter. Each year's growth is shown by a single ring. However, when the year is warm and wet, the ring will be larger because the tree grows more quickly than when the year is cold and dry. Tree-rings therefore reflect climatic changes. Recent work in Europe has shown that tree growth is greatest under intense cyclonic activity and is more a response to moisture than to temperature. Tree-ring timescales are being established by using the remains of oak trees, some nearly 10 000 years old, found in river terraces in south-central Europe. Bristlecone pines, still alive after 5000 years, give a very accurate measure in California (page 294).
- **Chemical methods** include the study of oxygen and carbon isotopes. An isotope is one of two or more forms of an element which differ from each other in atomic weight (i.e. they have the same number of protons in the nucleus, but a different number of neutrons). For example, two isotopes in oxygen are O-16 and O-18. The O-16 isotope, which is slightly lighter, vaporises more readily; whereas O-18, being heavier, condenses more easily. During warm, dry periods, the evaporation of O-16 will leave water enriched with O-18 which, if it freezes into polar ice, will be preserved as a later record (Places 14, page 104). Colder, wetter periods will be indicated by ice with a higher level of O-16. The most accurate form of dating is based on C-14, a radioactive isotope of carbon. Carbon is taken in by plants during the carbon cycle (Figure 11.25). Carbon-14 decays radioactively at a known rate and can be compared with C-12, which does not decay. Using C-12 and C-14 from a dead plant, scientists can determine the date of death to a standard error of ± 5 per cent. This method can accurately date organic matter up to 50 000 years old.
- **Historical records** of climatic change include:
  - cave paintings of elephants in central Sahara (Figure 7.27) and giraffes in Jordan (Figure 7.7)
  - vines growing successfully in southern England between AD 1000 and 1300
  - graves for human burial in Greenland which were dug to a depth of 2 m in the 13th century, but only 1 m in the 14th century, and could not be dug at all in the 15th century due to the extension of permafrost – in contrast to its retreat in the 2000s (Case Study 5)
  - fairs held on the frozen River Thames in Tudor times
  - the measurement of recent advances and retreats of alpine glaciers and polar sea-ice.

Causes of climatic change

Several suggestions have been advanced to try to explain climatic change over different timescales (Figure 4.2) and epochs (Figure 1.1). Most climatologists now accept that each of the causes of climatic change described below has a role to play in explaining change in the past, whether over long or short periods of time.

1 **Variations in solar energy** Although it was initially believed that solar energy output did not vary over time (hence the term 'solar constant' in Figure 9.3), increasing evidence suggests that sunspot activity, which occurs in cycles, may significantly affect our climate – times of high annual temperatures on Earth appear to correspond to periods of maximum sunspot activity.

2 **Astronomical relationships between the sun and the Earth** There is increasing evidence supporting Milankovitch's cycles of change in the Earth's orbit, tilt and wobble (Figure 4.6), which would account for changes in the amounts of solar radiation reaching the Earth's surface. This evidence is mainly from cores that have been drilled through undisturbed ocean-floor sediment which has accumulated over thousands of years (compare Places 14, page 104).

3 **Changes in oceanic circulation** Changes in oceanic circulation affect the exchange of heat between the oceans and the atmosphere. This can have both long-term effects on world climate (where currents at the onset of the Quaternary ice age flowed in opposite directions to those at the end of the ice age) and short-term effects (El Niño, Case Study 9A). The latest theory compares the North Atlantic Drift with a conveyor belt that brings water to north-west Europe. Should this conveyor belt be closed down, possibly by a huge influx of fresh water into the sea, then the climate will become dramatically colder.

4 **Meteorites** A major extinction event, which included the dinosaurs, took place about 65 million years ago. This event was believed to have been caused by one or more meteors colliding with the Earth. This seems to have caused a reduction in incoming radiation, a depletion of the ozone layer and a lowering of global temperatures.

5 **Volcanic activity** It has been accepted for some time that volcanic activity has influenced climate in the past, and continues to do so. World temperatures are lowered after any large single eruption, e.g. Mount Pinatubo (Case Study 1) and Krakatoa (Figure 1.29 and Places 35, page 289) or after a series of volcanic eruptions. This is due to the increase in dust particles in the lower atmosphere which will absorb and scatter more of the incoming radiation (Figure 9.4). Evidence suggests that these major eruptions may temporarily offset the greenhouse effect. Precipitation also increases due to the greater number of hygroscopic nuclei (dust particles) in the atmosphere (page 215).

6 **Plate tectonics** Plate movements have led to redistributions of land masses and to long-term effects on climate. These effects may result from a land mass 'drifting' into different latitudes (British Isles, page 22); or from the seabed being pushed upwards to form high fold mountains (page 19). The presence of fold mountains can lead to a colder climate (a suggested cause of the Quaternary ice age, page 103) and can act as a barrier to atmospheric circulation – the Asian monsoon was established by the creation of the Tibetan Plateau (page 239).

7 **Composition of the atmosphere** Gases in the atmosphere can be increased and altered following volcanic eruptions. At present there is increasing concern at the build-up of CO_2 and other greenhouse gases in the atmosphere (Case Study 9B), together with the use of aerosols and the release of CFCs (Places 27, page 209), which are blamed for the depletion of ozone in the upper atmosphere.

Climatic change in Britain

Britain's climate has undergone changes in the longest term (page 22 and Figure 1.1); during and since the onset of the Quaternary (Figure 4.2); and in the more recent short term (Figure 11.18). Following the 'little ice age' (which lasted from about AD 1540 to 1700), temperatures generally increased to reach a peak in about 1940. After that time, there was a tendency for summers to become cooler and wetter, springs to be later, autumns milder and winters more unpredictable. However, since the onset of the 1980s there appears to have been a considerable warming, with eight of the ten warmest years on record being in the last decade. This, together with the apparent increase in variations from the norm for Britain's expected autumn, winter, spring and even, since 2005, summer weather, tends to add further evidence to the concept of global warming (Case Study 9B).

Short-term and long-term climatic changes

A Short-term change: El Niño and La Niña

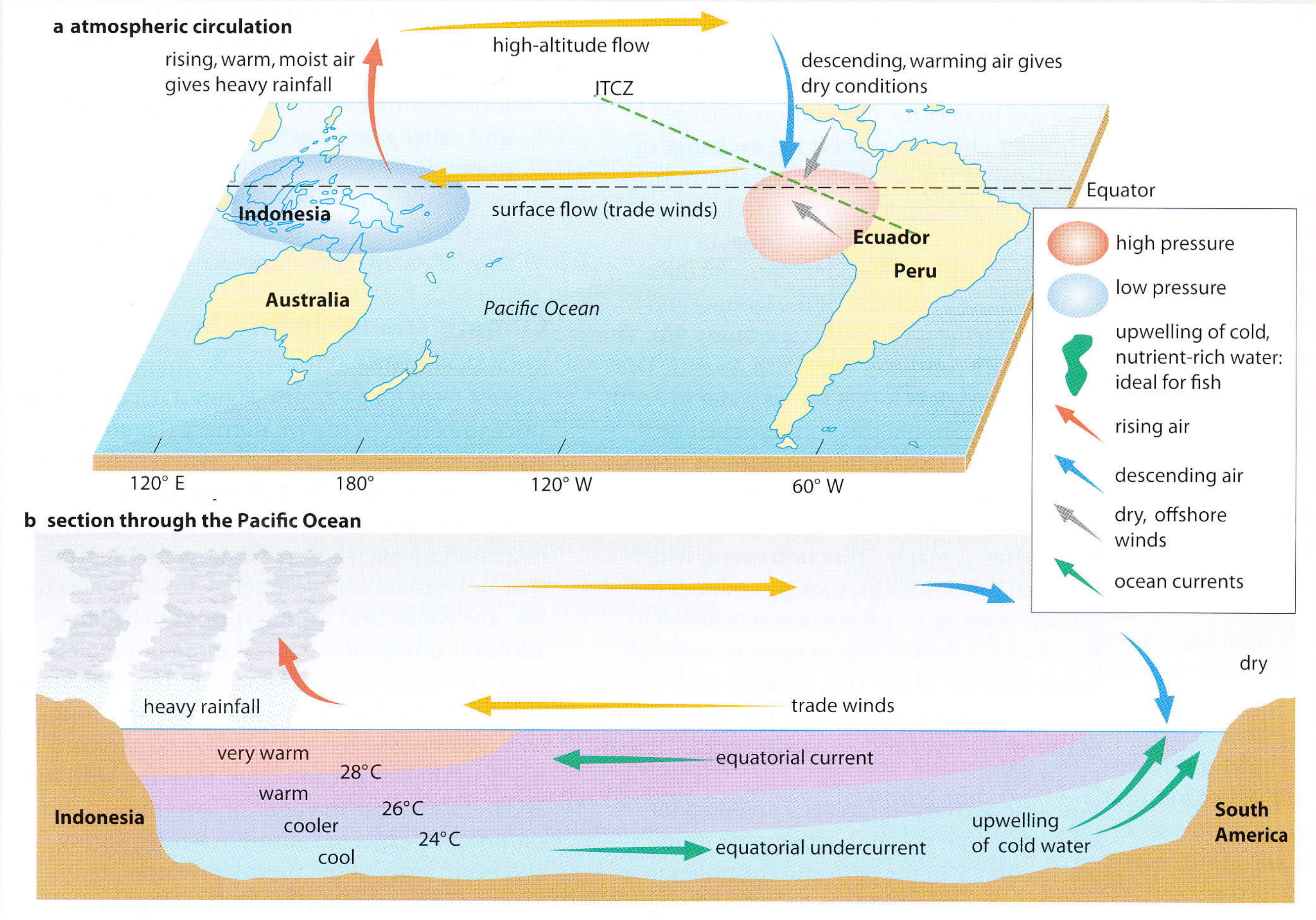
The oceans, as we have seen, have a considerable heat storage capacity which makes them a major influence on world climates. If ocean temperatures change, this will have a considerable effect upon weather patterns in adjacent land masses. Interactions between the ocean and the atmosphere have become, recently, a major scientific study.

The most important and interesting example of the ocean–atmosphere interrelationship is provided by the El Niño and La Niña events which occur periodically in the Pacific Ocean. Under normal atmospheric conditions, pressure rises over the eastern Pacific Ocean (off the coast of South America) and falls over the western Pacific Ocean (towards Indonesia and the Philippines). The descending air over the eastern Pacific gives the clear, dry conditions that create the Atacama Desert in Peru (Figure 7.2 and Places 24, page 180), while the warm, moist ascending air over the western Pacific gives that region its heavy convectional rainfall (page 226). This movement of air creates a circulation cell, named after Walker who first described it, in which the upper air moves from west to east, and the surface air from east to west as the trade winds (Figure 9.71). The trade winds:

- push surface water westwards so that sea-level in the Philippines is normally 60 cm higher than in Panama and Colombia
- allow water, flowing westward as the equatorial current, to remain near to the ocean surface where it can gradually heat. This gives the western Pacific the world's highest ocean temperature, usually above 28°C. In contrast, as warm water is pushed away from South America, it is replaced by an upwelling of colder, nutrient-rich water. This colder water lowers temperatures, sometimes to below 20°C, but does provide a plentiful supply of plankton which forms the basis of Peru's fishing industry.

Figure 9.71

The Walker circulation cell



El Niño

An El Niño event, scientifically referred to as an El Niño Southern Oscillation (ENSO), occurs periodically – on average every three to four years. It is called 'El Niño', which means 'little child' in Spanish, because, in those years that it does occur, it appears just after Christmas. An El Niño event usually lasts for 12–18 months.

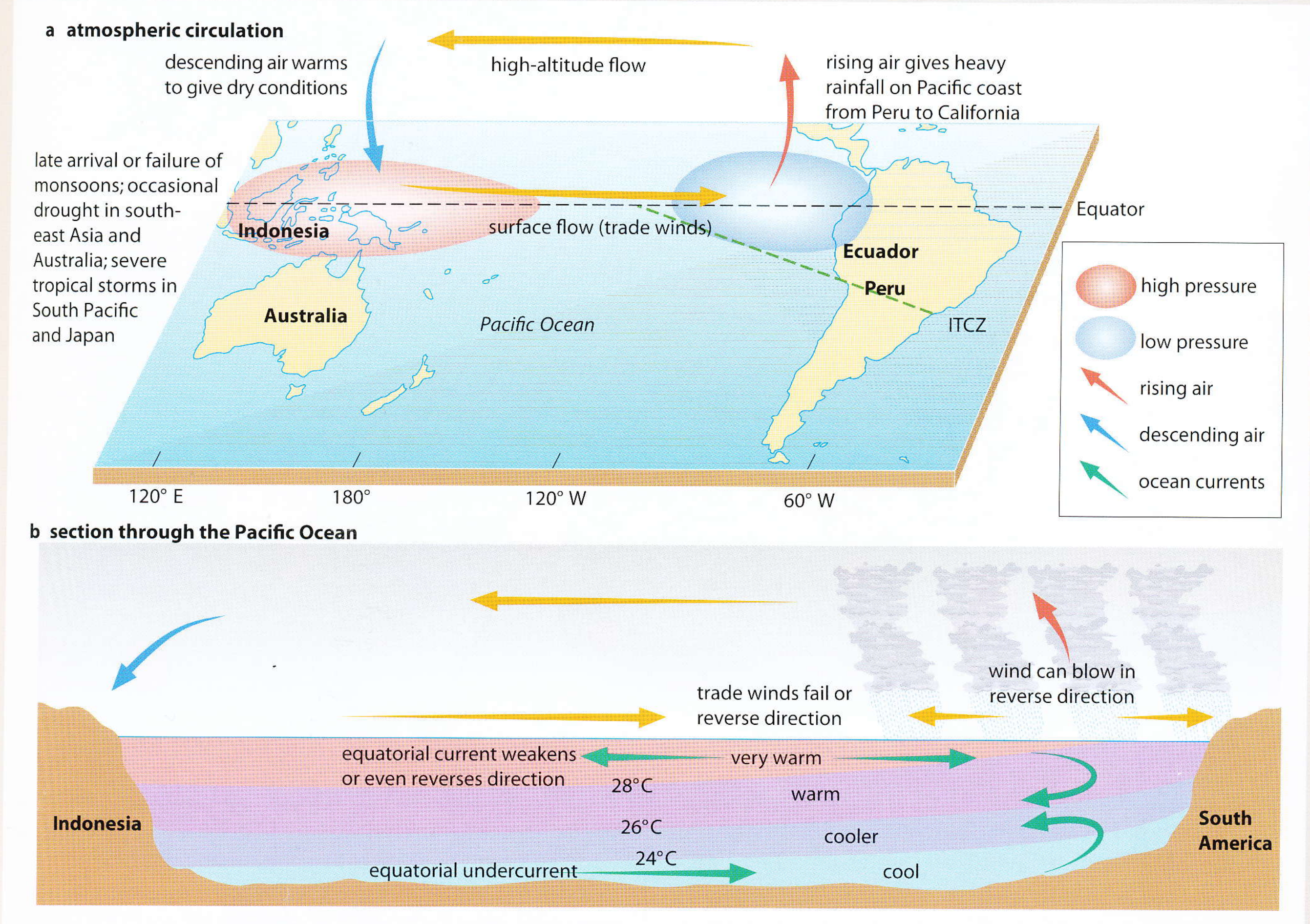
In contrast to normal conditions (Figure 9.71) there is a reversal, in the equatorial Pacific region, in pressure, precipitation and, often, winds and ocean currents (Figure 9.72). Pressure rises over the western Pacific and falls over the eastern Pacific. This allows the ITCZ (Figure 9.34) to migrate southwards and causes the trade winds to weaken in strength, or, sometimes, even to be reversed in their direction. The descending air, now over South-east Asia, gives that region much drier conditions than it usually experiences and, on extreme occasions, even causing drought. In contrast the air over the eastern Pacific is now rising, giving much wetter conditions in places, like Peru, that normally experience desert conditions. The change in the direction of the trade winds means that:

- surface water tends to be pushed eastwards so that sea-level in South-east Asia falls, while it rises in tropical South America
- surface water temperatures in excess of 28°C extend much further eastwards and the upwelling of cold water off South America is reduced, allowing sea temperatures to rise by up to 6°C. The warmer water in the eastern Pacific lacks oxygen, nutrients and, therefore, plankton and so has an adverse effect on Peru's fishing industry.

NASA-Mir astronauts were able, during the record-breaking 1997–98 El Niño, to observe, photograph and document the global impacts of the event. These, together with ground observations and recordings, are summarised in Figure 9.73.

Figure 9.72

An El Niño event



Evidence collected during the El Niño events of 1982–83 (at the time the biggest ever recorded), 1986 and 1992–93, increasingly suggested that the ENSO had a major effect on places far beyond the Pacific margins as well as on those bordering the ocean itself in its low latitudes. Apart from the drier conditions in South-east Asia and the wetter conditions in South America:

- severe droughts were experienced in the Sahel (Case Study 7) and southern Africa as well as across the Indian subcontinent
- there were extremely cold winters in central North America, and stormy conditions with floods in California
- exceptionally wet, mild and windy winters were experienced in Britain and north-west Europe.

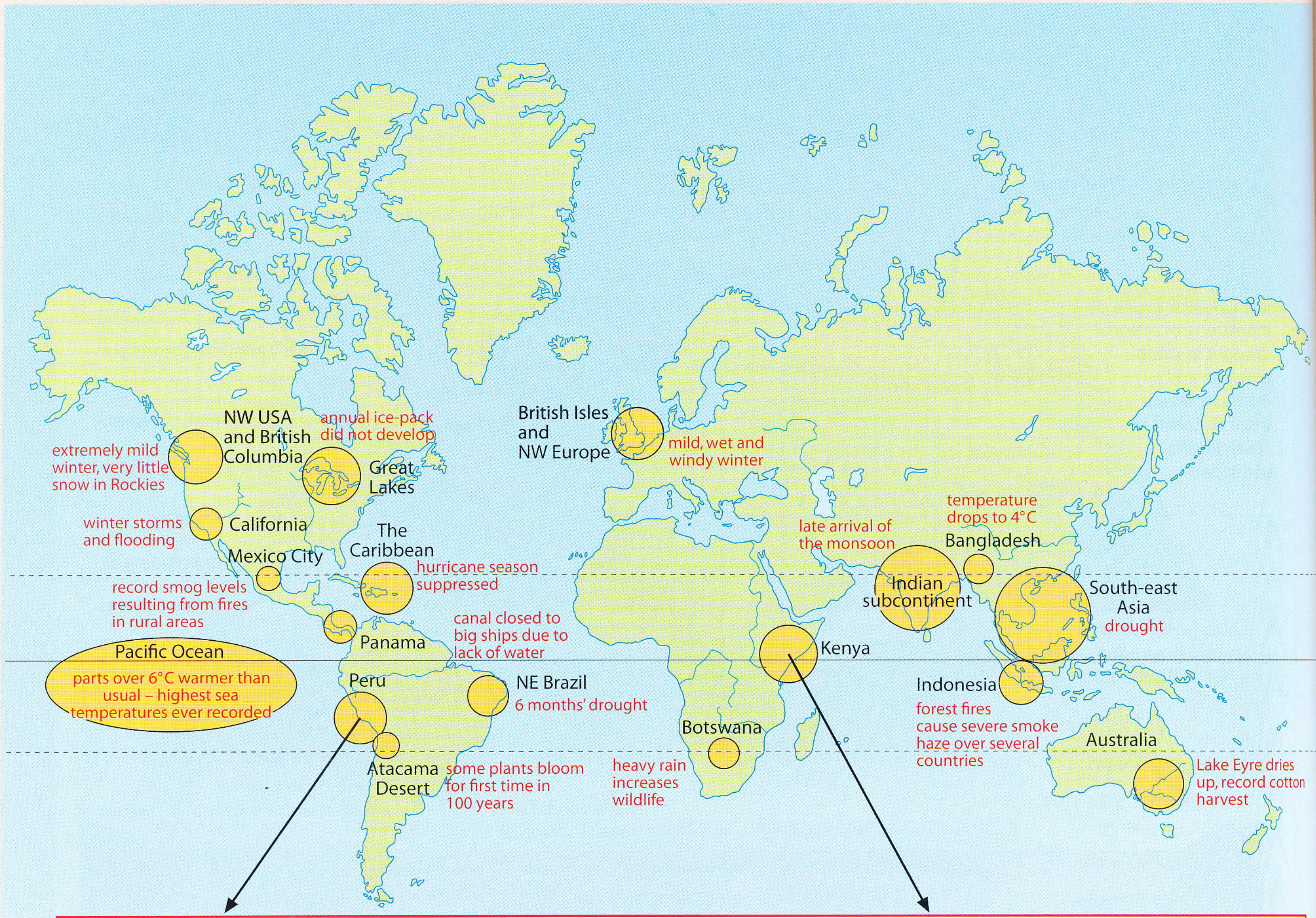
Short-term and long-term climatic changes

The 1997–98 event: the biggest yet experienced

Early 1997	Evidence of a rapid rise in sea temperatures in the eastern Pacific.
July	El Niño conditions intense.
September	Over 24 million km^2 of warm water (size of North and Central America) extended from the International Dateline to South America.
1998 April	Evidence of El Niño weakening.
June	NASA satellite surveillance showed a significant drop in sea temperatures in the eastern Pacific.
Autumn	Signs of a La Niña event (page 253).

Figure 9.73

The effects of the 1997–98 El Niño event



PERU For *each* of 12 days in early March, Peru received the equivalent of six months of normal rain. Over several months, flash flooding caused 292 deaths, injured more than 16 000 people, left 400 missing, destroyed 13 200 houses, wrecked 250 000 km of roads, swept away bridges, damaged crops and schools and disrupted the lives of up to half a million Peruvians.

KENYA Parts of Kenya received over 1000 mm of rainfall during six months (up to 50 times more than the average) at a time normally considered to be the 'dry season'. Roads and the mainline railway were swept away, the latter causing the derailment of the Nairobi–Mombasa train. Later, more than 500 people died of malaria as the receding floodwaters created ideal mosquito-spawning pools.

A mild El Niño episode: 2006–07

In September 2006, NASA's Jason altimetric satellite detected a rise in the sea-level of the Pacific Ocean which indicated the return of El Niño. However, the rise was slight, suggesting that the event might be short-lived and, being far less intense than the 1997–98 El Niño episode, unlikely to have a great effect on global weather patterns. It declined within six months without ending the drought in the south-west of the USA.

La Niña

Just as El Niño was ending in June 1998, forecasters were predicting – based on an 8°C fall in sea temperatures in the eastern Pacific in May – the arrival that winter of a La Niña event. La Niña, or 'little girl', has climatic conditions that are the reverse of those of El Niño. However, although when La Niña does appear it is just before or just after El Niño, its occurrence has been less frequent (the last was between June 1988 and February 1989) and, consequently, it is less easy to predict its possible effects because there is less evidence.

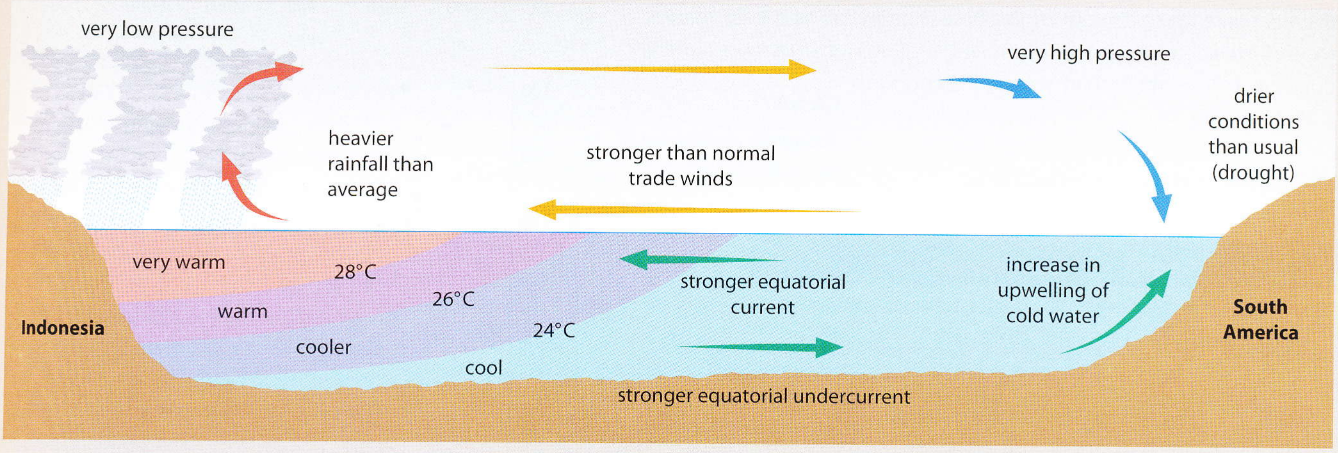
In a La Niña event, in contrast to normal conditions in the Pacific Ocean (Figure 9.71), the low pressure over the western Pacific becomes even lower and the high pressure over the eastern Pacific even higher (Figure 9.74). This means that rainfall increases over South-east Asia (was the La Niña event of 1988 responsible for the severe flooding at that time in Bangladesh?), there are drought conditions in South America and, due to the increased difference in pressure between the two places, the trade winds strengthen. The stronger trade winds:

- push large amounts of water westwards, giving a higher than normal sea-level in Indonesia and the Philippines
- increase the equatorial undercurrent and significantly enhance the upwelling of cold water off the Peruvian coast.

Scientists suggest that La Niña can be linked with increased hurricane activity in the Caribbean (Places 31) and that it can interrupt the jet stream over Britain to give stormier (Places 29), wetter (Case Study 3C) and cooler conditions.

Figure 9.74

A La Niña event



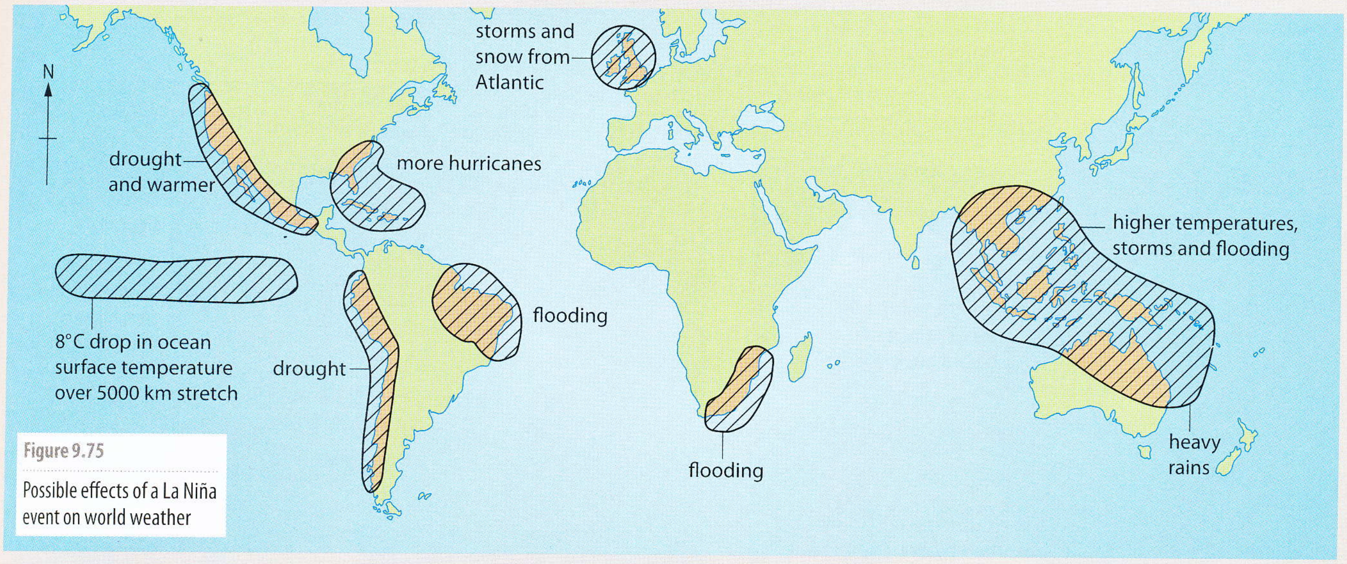


Figure 9.75

Possible effects of a La Niña event on world weather

A La Niña episode: 2007–08

The Jason altimetric satellite noted, in February 2007, a transition from the warm El Niño to the cool La Niña, a change not welcomed by the parched south-west of the USA. This La Niña episode, the strongest for several years, lasted for over 12 months until it began to weaken in April 2008. By then, it had caused torrential rain in Australia, breaking a long crop-ruining drought, and had given central China an exceptionally cold, snow-covered winter.

B Long-term change: global warming – an update

2005 and 2007: the warmest two years on record

Scientists claimed it was clear that temperatures around the world were continuing their upward climb. The global average for these years was 14.76°C in 2005 and 14.73°C in 2007 – the two warmest since reliable instrumental records began 126 years earlier and, according to palaeo-climatologists using evidence from ancient tree-rings (page 248), probably the highest in over 1200 years. Records collected by NASA GISS also showed that eight of the ten warmest years have been in the last decade and that 2007 was the 31st consecutive year when the global mean surface temperature exceeded the long-term average (Figure 9.76). More alarmingly, whereas the global mean rose by only 0.23°C in the 100 years between 1880 and 1979, in the 27 years since then it has increased by 0.62°C. Although the main reason for the rise in global temperature (Figure 9.76) is the longer-term effect of the continued release of greenhouse gases into the atmosphere (Figures 9.77 and 9.78), there is increasing evidence suggesting that temperatures increase more rapidly during an El Niño rather than in a La Niña episode (Case Study 9A).

Figure 9.76

Average global temperatures, 1880–2007

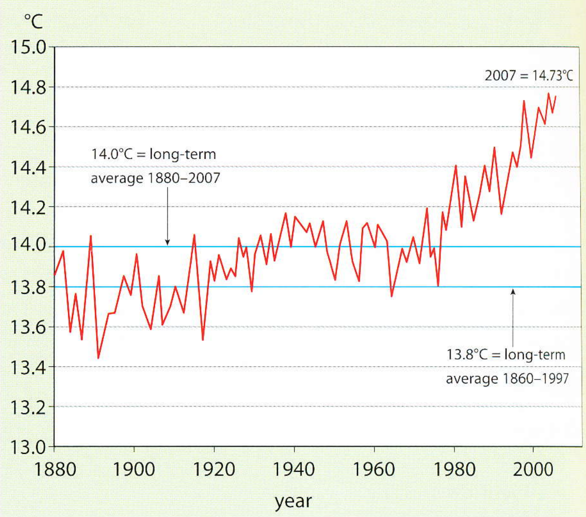


Figure 9.77

Atmospheric concentration of carbon dioxide, 1000–2007

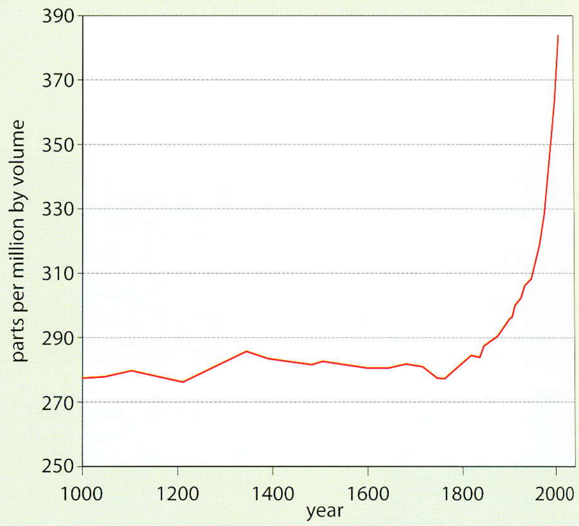


Figure 9.78

The major greenhouse gases

Gas	Sources (natural and manmade)
water vapour	evaporation from the ocean, evapotranspiration from land
carbon dioxide	burning of fossil fuels (power houses, industry, transport), burning rainforests, respiration
methane	decaying vegetation (peat and in swamps), farming (fermenting animal dung and rice-growing), sewage disposal and landfill sites
nitrous oxide	vehicle exhausts, fertiliser, nylon manufacture, power stations
CFCs	refrigerators, aerosol sprays, solvents and foams

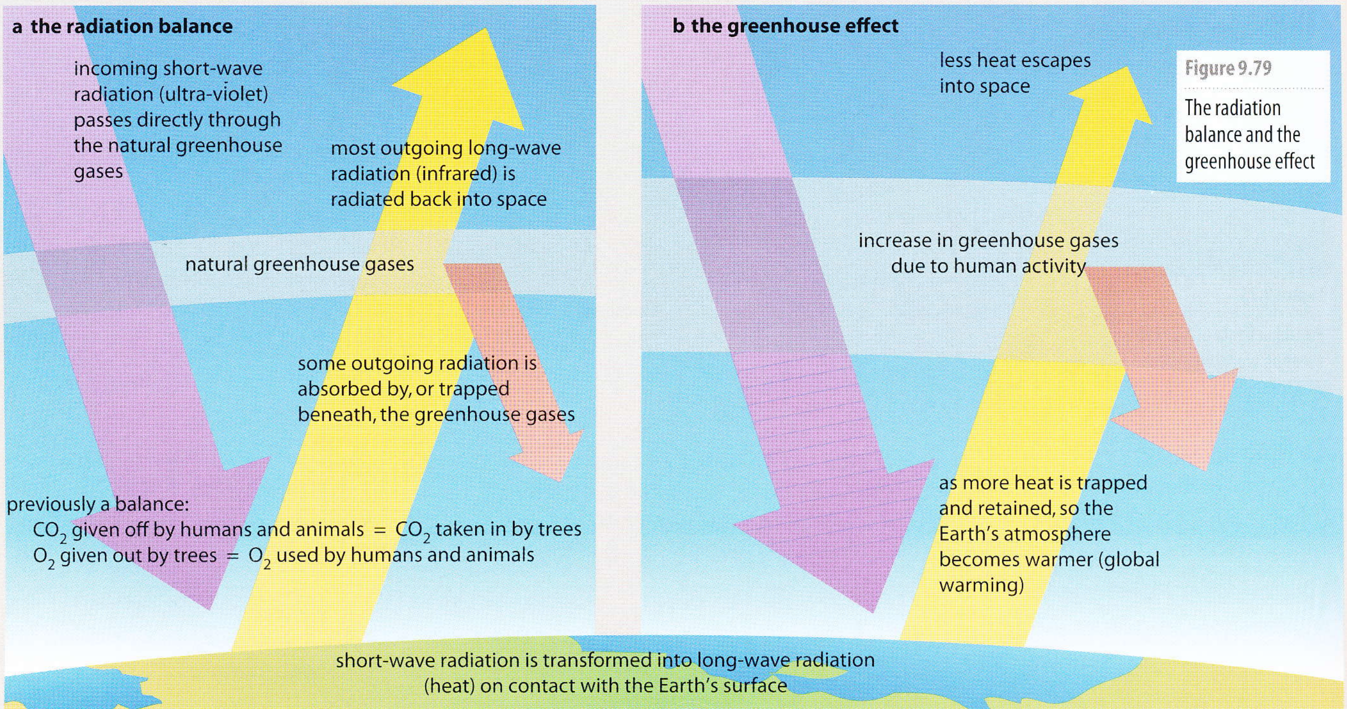


Figure 9.79

The radiation balance and the greenhouse effect

The Earth is warmed during the day by incoming, short-wave radiation (insolation) from the sun and cooled at night by out-going, longer-wave, infra-red radiation (page 207). As, over a lengthy period of time, the Earth is neither warming up nor cooling down, there must be a balance between incoming and outgoing radiation (page 209). While incoming radiation is able to pass through the atmosphere (which is 99 per cent nitrogen and oxygen, Figure 9.2), some of the outgoing radiation is trapped by a blanket of trace gases. Because they trap heat as in a greenhouse, these are referred to as **greenhouse gases** (Figure 9.79). Without these natural greenhouse gases, the Earth's average temperature would be 33°C lower than it is today – far too cold for life in any form. (During the last ice age, temperatures were only 4°C lower.) Water vapour provides the majority of the natural greenhouse effect, with lesser contributions from carbon dioxide, methane, nitrous oxide and ozone.

During the last 150 years there has been, with the exception of water vapour which remains a constant in the system, a rise in greenhouse gas concentrations (Figure 9.78). This has been due largely to the increase in world population and a corresponding growth in human activity, especially agricultural and industrial activities.

By adding these gases to the atmosphere, we are increasing its ability to trap heat (Figure 9.79). Most scientists now accept that the greenhouse effect is causing global warming. World temperatures have risen by 0.9°C in the last 100 years. Latest predictions suggest that they are likely to increase by between 1°C and 6°C by the year 2100. Some of the predicted global effects of this climate change are shown in Figure 9.81.

Britain's weather forecast for the 2080s

The latest government report predicts, in general, an increasingly grim forecast for the next 70 years. Heavy winter rains, up to 30 per cent in excess of today, will lead to more frequent flooding, as was seen in the English Midlands in 2007 (Case Study 3C) and destructive gales will be more frequent and severe. With a predicted rise in sea-level of between 2 and 10 cm, storm surges and higher tides will threaten coastal areas (Case Study 6). However, the chances of extremely cold winters, and the risk of fog and heavy snowfalls, will decrease. Days with more than 25 mm of rain, at present an extreme event, could occur three or four times a year. Summers will be drier with a decrease in rain of up to 30 per cent in the south-east where drought will become more common. With a predicted increase in summer temperatures of over 3°C, heat waves will become a more regular occurrence and there will be many more days when thermometers exceed 25°C. Changes in the weather will be greater in the south-east than in the north-west.

However, some computer predictions are suggesting that Britain's climate could, over a long period of time, get colder. This could happen if the release of fresh water from Greenland's melting ice-cap pushed the North Atlantic Drift further south so that it no longer affected all, or certainly parts, of Britain.

Effects of climate change in the UK

DEFRA's claims, based on the predicted forecast of milder, wetter, stormier winters and warmer, drier summers, are summarised in Figure 9.80. Its two main concerns are:

- the potential effects of changing rainfall patterns on hydrology and ecosystems
- rising sea-levels and more frequent storms in coastal areas where there is a large proportion of Britain's population, its manufacturing industry, energy production, mineral extraction, valued natural environments and recreational amenities.

Soils	Higher temperatures could reduce water-holding capacities and increase soil moisture deficits, affecting the types of crops and trees. Less organic matter due to drier summers (less produced) and wetter winters (more lost).
Flora/fauna	Higher temperatures and increased water deficit could mean loss of several native species. Warmer climate would allow plants to grow further north and at higher altitudes. Earlier flowering plants and arrival of migrant birds.
Agriculture	Grasses helped by longer growing season (extra 15 days) but cereals hit by drier summers. Increase in number of pests. Maize and vines in the south. Need for irrigation in summer.
Forestry	Certain trees able to grow at higher altitudes. New species could be introduced from warmer climates. Threats from fires, diseases and pests.
Coastal regions	Rise in sea-level plus increase in frequency/number of gales and frequency/height of storm surges would mean more flooding, especially around estuaries, and increased erosion. Major impact on housing, industry, farming, energy, transport and wildlife, including marine ecosystems.
Water resources	Water resources would benefit from wetter winters, but hotter, drier summers would increase demands/pressures. Need for irrigation in summer in south-east. More frequent river flooding.
Energy	Space heating demand would fall in winter but need for air-conditioning would rise in summer. Probable overall fall in demand. Many power stations are in threatened coastal areas.
Manufacturing/construction	Problem for coastal industries. Fewer days lost in construction due to less snow/frost.
Transport	Many types of transport are sensitive to extreme weather conditions. Benefit of less snow, ice and perhaps fog. Loss due to more frequent and severe storms and flooding, including flash floods.
Recreation/tourism	Tourism would benefit from longer, warmer, drier summers, but insufficient snow for skiing in Scotland.

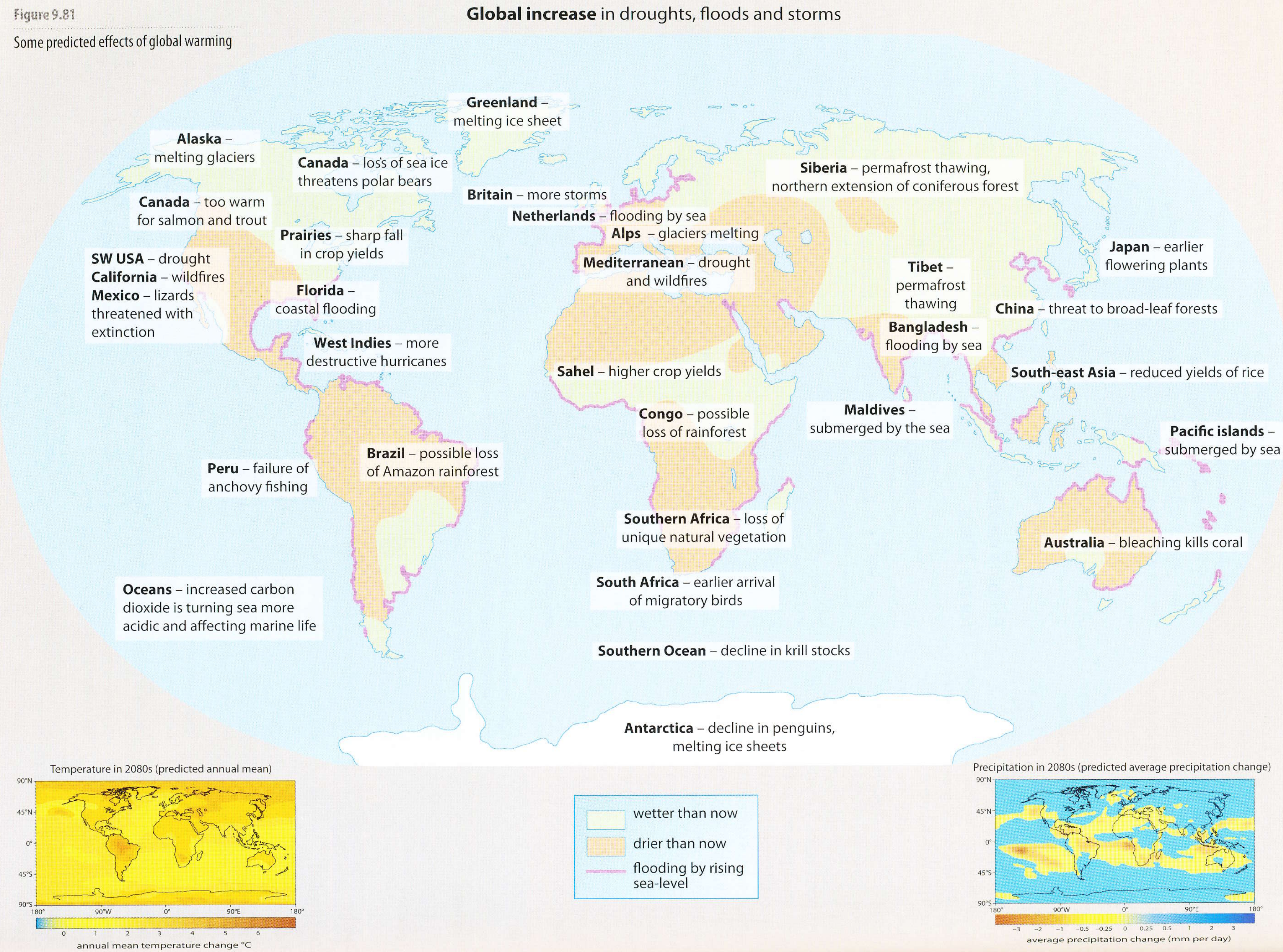
Source: DEFRA

Figure 9.80

Specific effects of climate change in the UK

Figure 9.81

Some predicted effects of global warming



Further reference

Barry, R.G. and Chorley, R.J. (2003) *Atmosphere, Weather and Climate*, Routledge.

Bentley, M. (2005) 'Antarctic ozone hole', *Geography Review* Vol 18 No 3 (January).

Burt, T. (2005) 'Rain in the hills', *Geography Review* Vol 18 No 4 (March).

Digby, B. (2005) 'El Niño Part 1', *Geography Review* Vol 19 No 2 (November).

Digby, B. (2006) 'El Niño Part 2' *Geography Review* Vol 19 No 3 (January).

Goudie, A.S. (2001) *The Nature of the Environment*, WileyBlackwell.

Holden, J. (2008) *Introduction to Physical Geography and the Environment*, Prentice Hall.

Middleton, N. 'Acid shock', *Geography Review* Vol 18 No 4 (March).

O'Hara, G., Sweeney, J. and O'Hare, G. (1986) *The Atmospheric System*, Oliver & Boyd.

Smithson, P., Addison, K. and Atkinson, K. (2008) *Fundamentals of Physical Geography*, Routledge.

Center for Ocean-Atmospheric Prediction Studies, resources:
http://coaps.fsu.edu/lib/elninolinks/themes

Earth Space Research Group, Indian monsoons:
www.icess.ucsb.edu/esrg/IOM2/Start2_IOM.html

Jet Propulsion Laboratory, NASA, El Niño:
www.jpl.nasa.gov/earth/ocean_motion/el_nino_index.cfm

UK Climate Impacts Programme:
www.ukcip.org.uk/index.php

UK Meterological Office (Met Office):
www.metoffice.gov.uk/

UK Met Office weather charts:
www.meto.gov.uk/education/data/charts.html

Union of Concerned Scientists (UCS), global warming:
http://ucsusa.org/warming/index.html

Union of Concerned Scientists (UCS), ozone depletion:
www.ucsusa.org/global_warming/science_and_impacts/science/faq-about-ozone-depletion-and.html

US Environmental Protection Agency, global warming:
www.epa.gov/climatechange/index.html

US Environmental Protection Agency, glossary of climate change terms:
http://yosemite.epa.gov/oar/globalwarming.nsf/content/glossary.html

US Environmental Protection Agency, ozone science:
www.epa.gov/ozone/strathome.html

US National Oceanographic and Atmospheric Administration (NOAA) Climate Prediction Center:
www.cpc.noaa.gov/

US NOAA, El Niño / La Niña:
www.cpc.noaa.gov/products/analysis_monitoring/ensostuff/

www.elnino.noaa.gov/

US NOAA, hurricanes:
http://hurricanes.noaa.gov/

US NOAA, research:
www.cdc.noaa.gov/ENSO

Questions & Activities

Activities

1 **a** What is the 'atmosphere' of the Earth? *(3 marks)*

b What is the *difference* between 'weather' and 'climate'? *(4 marks)*

c Describe the 'solar cascade of energy' to the Earth. *(4 marks)*

d What is the importance of **i** carbon dioxide and **ii** clouds in the energy balance of the Earth? *(4 marks)*

e Ozone in the troposphere is a danger to health. Why is there concern that ozone in the stratosphere is being depleted? *(5 marks)*

f What measures can be taken to restrict the potential damage due to ozone depletion? *(5 marks)*

2 **a** How does a meteorologist get information to forecast the weather? *(4 marks)*

b Use Places 29 (page 232) to answer the following questions:

i What was the weather forecast on 11–15 October 1987? *(3 marks)*

ii Describe the meteorological conditions over the Western Approaches and Bay of Biscay at 6.00 pm on 15 October. *(3 marks)*

iii Describe the track of the storm over the next 12 hours. *(4 marks)*

iv What happened to the weather over southern England during this 12-hour period? *(4 marks)*

v Describe three effects of the storm on people. *(3 marks)*

c Explain two reasons why meteorologists failed to forecast the very strong winds of 15 October. *(4 marks)*

Exam practice: basic structured questions

3 **a** Explain how each of the following factors affects the winds that cross them:

i a large body of water (e.g. a sea) *(4 marks)*

ii a mountain range. *(6 marks)*

b On a field course in Switzerland a geography student noted: 'On the north-facing side of the valley the forests came close to the valley floor while the settlement huddled at the foot of the south-facing slope and here there were ploughed fields. There were forests but they started higher up the slope.'

Suggest the cause of these differences in land use. *(6 marks)*

c A January weather forecast for the UK stated: 'Although it will be cool today, temperatures will stay above freezing tonight because of the cloud cover'.

Explain the effect of cloud on temperature. *(4 marks)*

d Why is it warmer in summer than in winter? *(5 marks)*

4 **a** **i** What is 'stratus' cloud? *(2 marks)*

ii What is 'cumulo-nimbus' cloud? *(2 marks)*

b Making good use of diagrams, explain why rain falls when an onshore wind blows over an upland area. *(7 marks)*

c Why does fog often form over a coastal area in the autumn? *(6 marks)*

d Explain the formation of smog over an urban area. *(8 marks)*

5 **a** Describe the causes of the ITCZ. *(5 marks)*

b What weather conditions are associated with the ITCZ? *(10 marks)*

c Why does the ITCZ move with the seasons? *(10 marks)*

6 Study Figure 9.82 and answer the following questions.

a What is the name of the pressure system shown? *(2 marks)*

b What is the weather like at place A (Doncaster)? *(4 marks)*

c What is the red line with half circles on it? *(5 marks)*

d Locate the warmest and the coolest place in the British Isles. *(2 marks)*

e **i** Over the next 12 hours the pressure system moves so that it is in the North Sea.

Give a weather forecast for place A (Doncaster) over this period. *(6 marks)*

ii Why would you expect this to happen? *(6 marks)*

Exam practice: structured questions

7 **a** Study Figures 9.82 and 9.83. Describe the changes in the weather being experienced at Limerick (place C) over this 24-hour period. *(8 marks)*

b Explain what has happened to the frontal system over this period of time. *(8 marks)*

c Describe, and explain the causes of, the types and distribution of the precipitation shown in Figure 9.83. *(9 marks)*

8 **a** Describe **three** mechanisms that are likely to trigger upward movement of a parcel of air from sea level. *(6 marks)*

b Study Figure 9.84.

i What is meant by the term 'ELR'? *(4 marks)*

ii Identify the height of the base of clouds. *(1 mark)*

iii Explain why this height is the cloud base. *(4 marks)*

iv Identify the air stream(s) (A, B, C) that would have cloud cover. State why this is so. *(2 marks)*

v At what height would condensation in a cloud be in the form of ice? *(2 marks)*

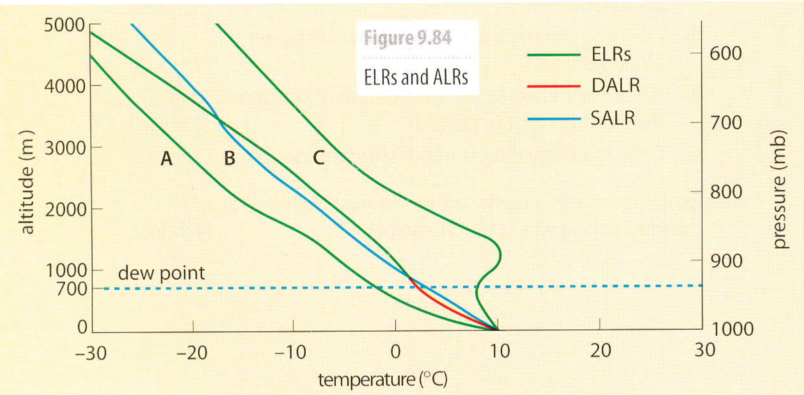


Figure 9.84 ELRs and ALRs

c Choose **either** stability **or** instability. Describe and explain the weather conditions normally associated with that atmospheric condition. *(6 marks)*

9 **a** **i** Using an annotated diagram only, illustrate the variation of temperature and pressure with altitude in the atmosphere. *(6 marks)*

ii Explain the variations in temperature with altitude in the atmosphere. *(6 marks)*

b **i** Study Figure 9.5 (page 209). Making good use of the data, explain why there is a general trend of movement of heat energy from the Equator to the poles. *(6 marks)*

ii Describe how heat is transferred from the tropics towards the poles. *(7 marks)*

10 **a** Describe and explain what happens to *incoming solar radiation* (**insolation**) once it reaches the edge of the Earth's atmosphere. *(10 marks)*

b Explain the importance of **each** of the following in relation to heat energy in the atmosphere:

i latitude

ii altitude

iii land and sea. *(10 marks)*

c The greatest amount of insolation is experienced close to the Equator. Why does this area not become increasingly hot? *(5 marks)*

11 **a** Suggest **one** way you could test the hypothesis that the temperatures in an urban area are different from those in the surrounding countryside. Describe the method you would use to collect and record the data to carry out the proposed test. *(7 marks)*

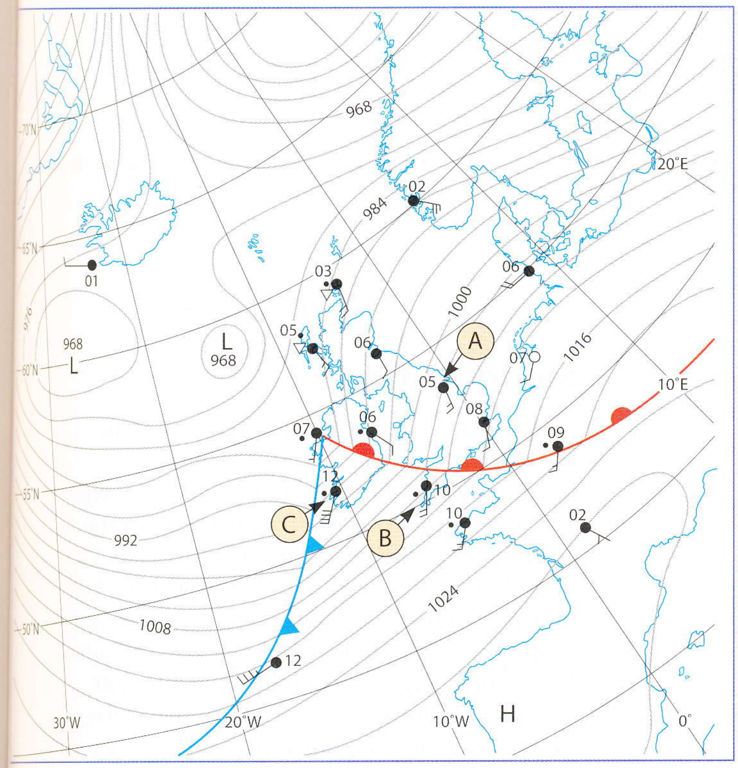


Figure 9.82

Weather map for 1200 hrs,
12 January 1984

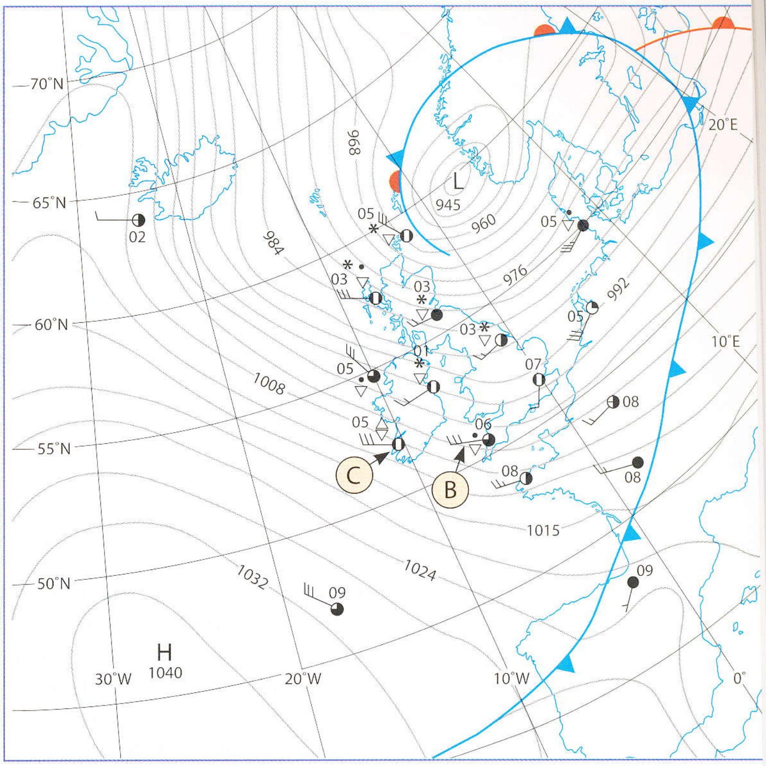


Figure 9.83

Weather map for 1200 hrs,
13 January 1984

b Explain **two** reasons why temperatures in urban areas may be higher than those in surrounding rural areas. *(10 marks)*

c Suggest **two** ways in which planning policies can reduce the problems caused by microclimatic features of urban areas. *(8 marks)*

12 **a** Explain the **difference** between absolute humidity and relative humidity. *(8 marks)*

b Making good use of diagrams, show how condensation occurs as air rises through the atmosphere. *(10 marks)*

c Explain the cause of low-level clouds (mist) as shown in Figure 9.23 (page 221). *(7 marks)*

13 The following are meteorological conditions that develop a range of weather conditions over the British Isles:

a an anticyclone centred over the English Midlands in winter

b a mature depression with its centre over the Central Valley of Scotland in summer

c a depression centred over Paris and an anticyclone to the north of Scotland in January.

Choose two of the situations a–c and, in both cases, describe how weather conditions would vary in two contrasting locations in the British Isles.

Explain these variations. *(12 + 13 marks)*

14 **a** Study Figure 9.49 (page 235). Describe the major distribution of tropical storms as shown on the map. *(6 marks)*

b Choose any **one** type of tropical storm. Describe and explain the sequence of weather associated with the passage of the storm. *(10 marks)*

c Explain how people respond to the hazard posed by tropical storms. In your answer refer to countries at different stages of economic development. *(9 marks)*

Exam practice: essays

15 'The polar front jet stream is one of the most important influences on the climate of the British Isles.'

Discuss this statement. *(25 marks)*

16 The passage of a depression over the British Isles leads to predictable changes in the weather over a period of time. Describe and explain the sequence of weather experienced in Liverpool over a 12-hour period as a mature depression passes from west to east. *(25 marks)*

17 'There is now overwhelming scientific evidence that human activity is causing major changes to the global climate.'

Is this statement true? Justify your answer. *(25 marks)*