2 Atmosphere and weather

2.1 Diurnal energy budgets

An energy budget refers to the amount of energy entering a system, the amount leaving the system and the transfer of energy within the system. Energy budgets are commonly considered at a global scale (macro-scale) and at a local scale (micro-scale). However, the term microclimate is sometimes used to describe regional climates, such as those associated with large urban areas, coastal areas and mountainous regions.

Figure 2.1 shows a classification of climate and weather phenomena at a variety of spatial and temporal scales. Phenomena vary from small-scale turbulence and eddying (such as dust devils) that cover a small area and last for a very short time, to large-scale anticyclones (high-pressure zones) and jet streams that affect a large area and may last for weeks. The jet stream that carried volcanic dust from underneath the Eyjafjallajökull glacier in Iceland to northern Europe in 2010 is a good example of jet-stream activity (Figure 2.2).

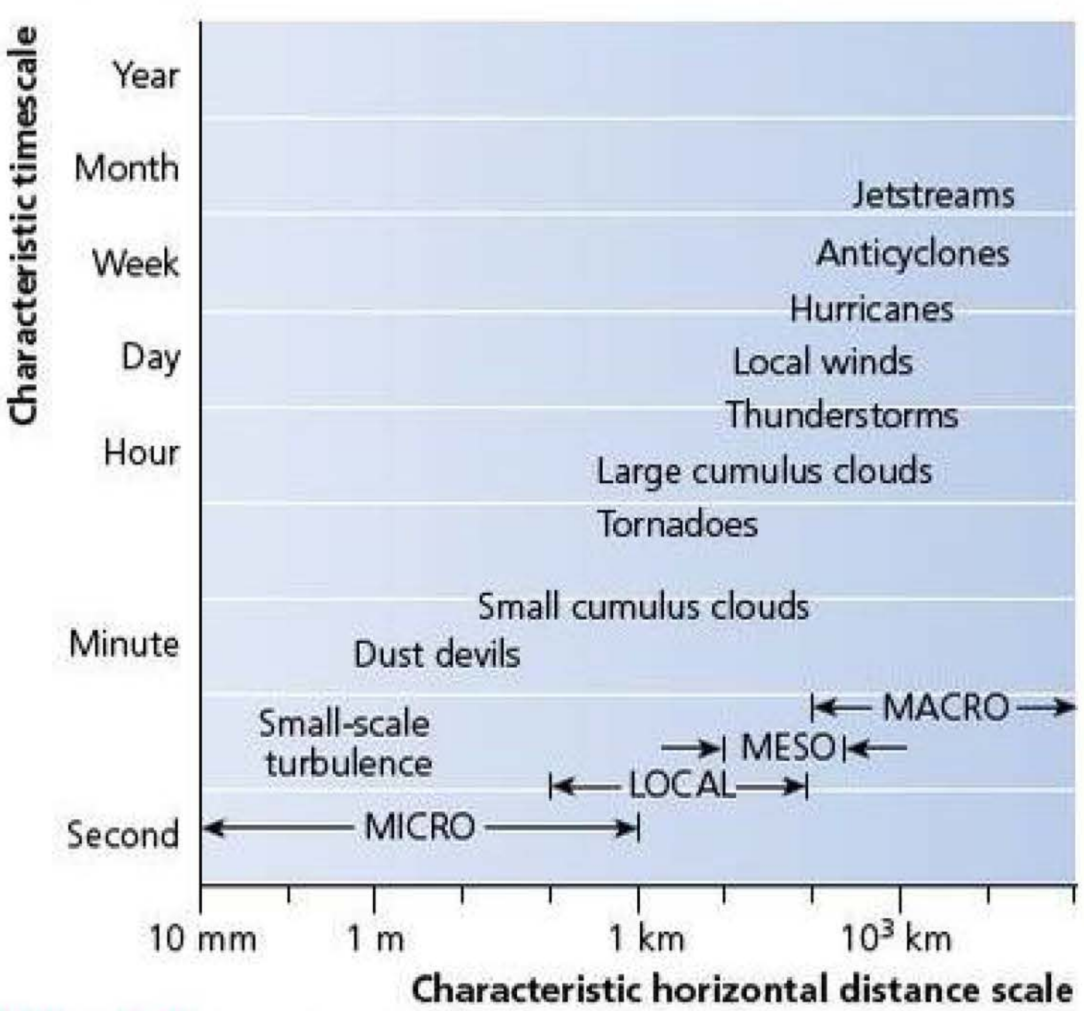


Figure 2.1 Classification of climate and weather phenomena at a variety of spatial and temporal scales

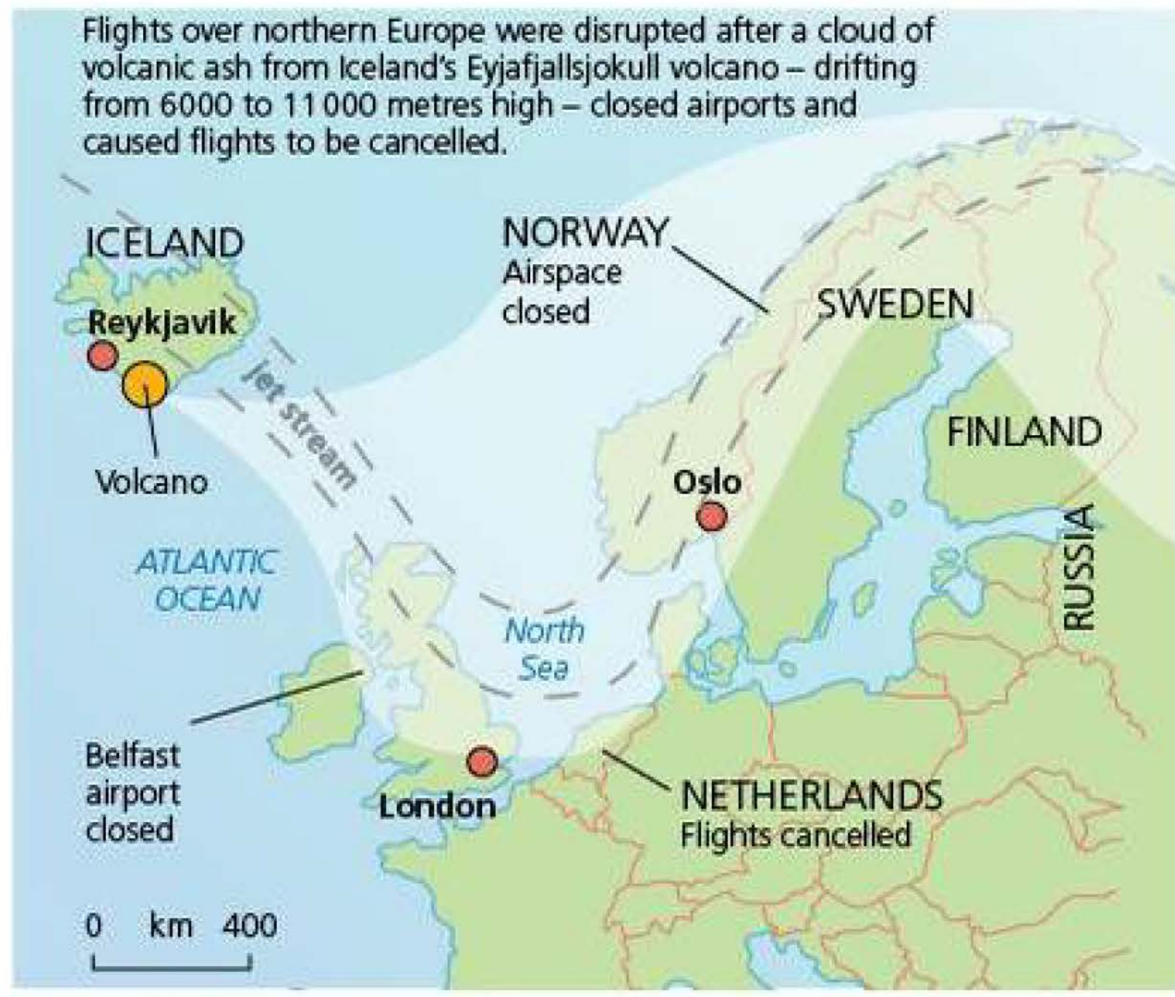


Figure 2.2 Jet-stream activity and the transfer of dust from Eyjafjallajökull, Iceland

These different scales should not be considered as separate scales but as a hierarchy of scales in which smaller phenomena may exist within larger ones. For example, the temperature surrounding a building will be affected by the nature of the building and processes that are taking place within the building. However, it will also be affected by the wider synoptic (weather) conditions, which are affected by latitude, altitude, cloud cover and season, for example.

Daytime and night-time energy budgets

There are six components to the daytime energy budget:

- incoming (shortwave) solar radiation (insolation)
- reflected solar radiation
- surface absorption
- sensible heat transfer
- long-wave radiation (Figure 2.3)
- latent heat (evaporation and condensation).

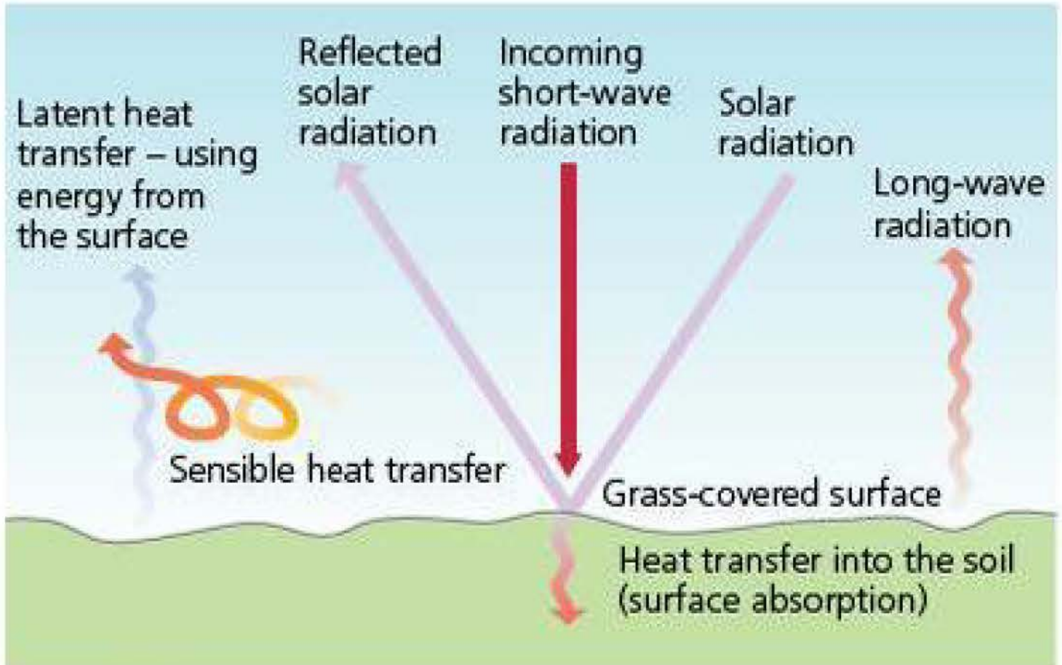


Figure 2.3 Local energy budget – daytime

These influence the gain or loss of energy for a point at the Earth's surface. The daytime energy budget assumes a horizontal surface with grass-covered soil and can be expressed by the formula:

energy available at the surface = incoming solar radiation – (reflected solar radiation + surface absorption + sensible heat transfer + long-wave radiation + latent heat transfers)

In contrast, the night-time energy budget consists of four components:

- long-wave Earth radiation
- latent heat transfer (condensation)
- absorbed energy returned to Earth (sub-surface supply)
- sensible heat transfer (Figure 2.4).

Incoming (shortwave) solar radiation

Incoming solar radiation (insolation) is the main energy input and is affected by latitude, season and cloud cover (Section 2.2). Figure 2.5 shows how the amount of insolation received varies with the angle of the Sun and with cloud type. For example, with strato-cumulus clouds (like those in Figure 2.6) when the Sun is low in the sky, about 23 per cent of the total radiation transmitted is received at the Earth's surface – about 250 watts per m². When the Sun is high in the sky, about 40 per cent is received – just over 450 watts per m². The less cloud cover there is, and/or the higher the cloud, the more radiation reaches the Earth's surface.

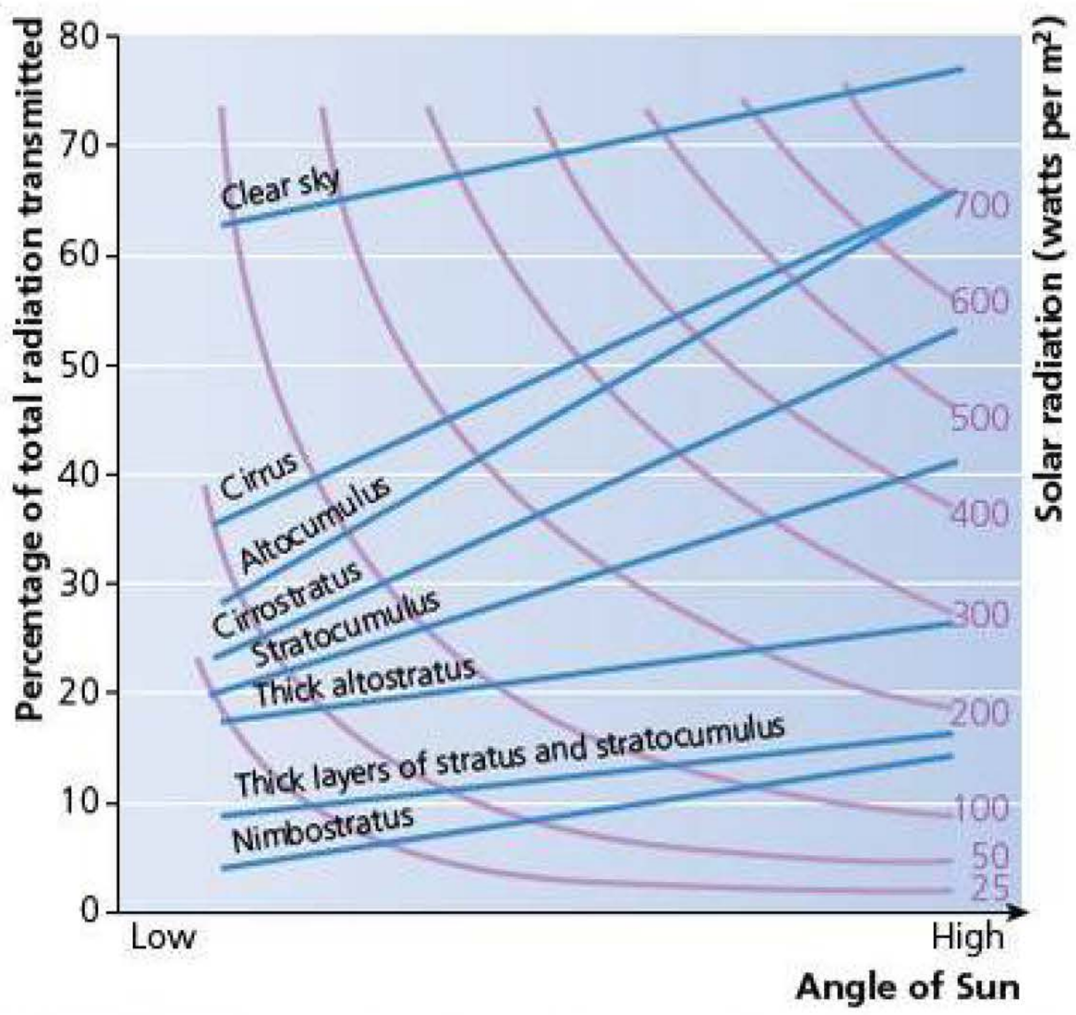


Figure 2.5 Energy, cloud cover/type and the angle of the Sun

Figure 2.6 Stratocumulus clouds

Reflected solar radiation

The proportion of energy that is reflected back to the atmosphere is known as the albedo. The albedo varies with colour – light materials are more reflective than dark materials (Table 2.1). Grass has an average albedo of 20–30 per cent, meaning that it reflects back about 20–30 per cent of the radiation it receives.

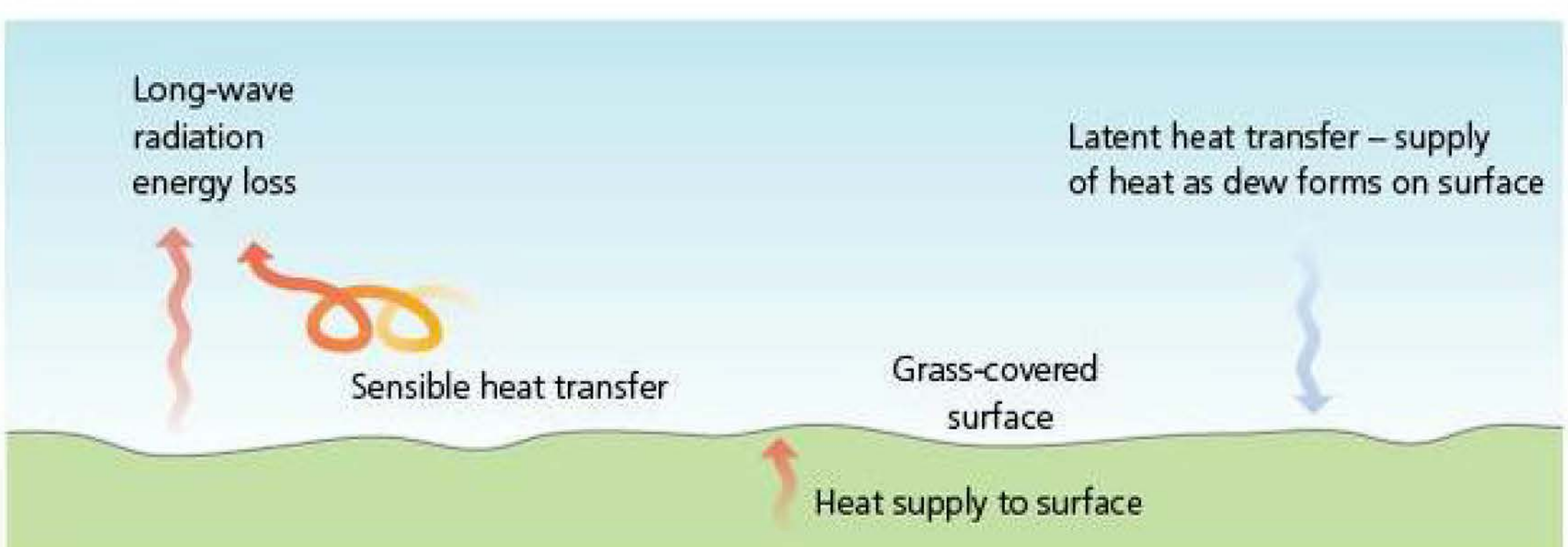


Figure 2.4 Night-time energy budget

Table 2.1 Selected albedo values

Surface	Albedo (%)
Water (Sun's angle over 40°)	2–4
Water (Sun's angle less than 40°)	6–80
Fresh snow	75–90
Old snow	40–70
Dry sand	35–45
Dark, wet soil	5–15
Dry concrete	17–27
Black road surface	5–10
Grass	20–30
Deciduous forest	10–20
Coniferous forest	5–15
Crops	15–25
Tundra	15–20

Section 2.1 Activities

1 The model for the daytime energy budget assumes a flat surface with grass-covered soil. Suggest reasons for this assumption.
2 Study Table 2.1.
   a What is meant by the term *albedo*?
   b Why is albedo important?

Surface and sub-surface absorption

Energy that reaches the Earth's surface has the potential to heat it. Much depends on the nature of the surface. For example, if the surface can conduct heat to lower layers, the surface will remain cool. If the energy is concentrated at the surface, the surface warms up.

The heat transferred to the soil and bedrock during the day may be released back to the surface at night. This can partly offset the night-time cooling at the surface.

Sensible heat transfer

Sensible heat transfer refers to the movement of parcels of air into and out of the area being studied. For example, air that is warmed by the surface may begin to rise (convection) and be replaced by cooler air. This is known as a convective transfer. It is very common in warm areas in the early afternoon. Sensible heat transfer is also part of the night-time energy budget: cold air moving into an area may reduce temperatures, whereas warm air may supply energy and raise temperatures.

Long-wave radiation

Long-wave radiation refers to the radiation of energy from the Earth (a cold body) into the atmosphere and, for some of it, eventually into space. There is, however, a downward movement of long-wave radiation from particles in the atmosphere. The difference between the two flows is known as the net long-wave radiation balance. During the day, the outgoing long-wave radiation transfer is greater than the incoming long-wave radiation transfer, so there is a net loss of energy from the surface.

During a cloudless night, there is a large loss of long-wave radiation from the Earth. There is very little return of long-wave radiation from the atmosphere, due to the lack of clouds. Hence there is a net loss of energy from the surface. In contrast, on a cloudy night the clouds return some long-wave radiation to the surface, hence the overall loss of energy is reduced. Thus in hot desert areas, where there is a lack of cloud cover, the loss of energy at night is maximised. In contrast, in cloudy areas the loss of energy (and change in daytime and night-time temperatures) is less noticeable.

Latent heat transfer (evaporation and condensation)

When liquid water is turned into water vapour, heat energy is used up. In contrast, when water vapour becomes a liquid, heat is released. Thus when water is present at a surface, a proportion of the energy available will be used to evaporate it, and less energy will be available to raise local energy levels and temperature.

During the night, water vapour in the air close to the surface can condense to form water, since the air has been cooled by the cold surface. When water condenses, latent heat is released. This affects the cooling process at the surface. In some cases, evaporation may occur at night, especially in areas where there are local sources of heat.

Dew

Dew refers to condensation on a surface. The air is saturated, generally because the temperature of the surface has dropped enough to cause condensation. Occasionally, condensation occurs because more moisture is introduced, for example by a sea breeze, while the temperature remains constant.

Absorbed energy returned to Earth

The insolation received by the Earth will be reradiated as long-wave radiation. Some of this will be absorbed by water vapour and other greenhouse gases, thereby raising the temperature.

Temperature changes close to the surface

Ground-surface temperatures can vary considerably between day and night. During the day, the ground heats the air by radiation, conduction (contact) and convection. The ground radiates energy and as the air receives more radiation than it emits, the air is warmed. Air close to the ground is also warmed through conduction. Air movement at the surface is slower due to friction with the surface, so there is more time for it to be heated. The combined effect of radiation and conduction is that the air becomes warmer, and rises as a result of convection.

At night, the ground is cooled as a result of radiation. Heat is transferred from the air to the ground.

Case Study: Annual surface energy budget of an Arctic site – Svalbard, Norway

The annual cycle of the surface energy budget at a high-arctic permafrost site on Svalbard shows that during summer, the net short-wave radiation is the dominant energy source (Figure 2.7). In addition, sensible heat transfers and surface absorption in the ground lead to a cooling of the surface. About 15 per cent of the net radiation is used up by the seasonal thawing of the active layer in July and August (the active layer is the layer at the top of the soil that freezes in winter and thaws in summer). During the polar night in winter, the net long-wave radiation is the dominant energy loss channel for the surface, which is mainly compensated by the sensible heat transfer and, to a lesser extent, by the ground heat transfer, which originates from the refreezing of the active layer. The average annual sensible heat transfer of –6.9 Wm^{-2} is composed of strong positive transfers in July and August, while negative transfers dominate during the rest of the year. With 6.8 Wm^{-2}, the latent heat transfer more or less compensates the sensible heat transfer in the annual average. Strong evaporation occurs during the snowmelt period and particularly during the snow-free period in summer and autumn. When the ground is covered by snow, latent heat fluxes through sublimation of snow are recorded, but are insignificant for the average surface energy budget.

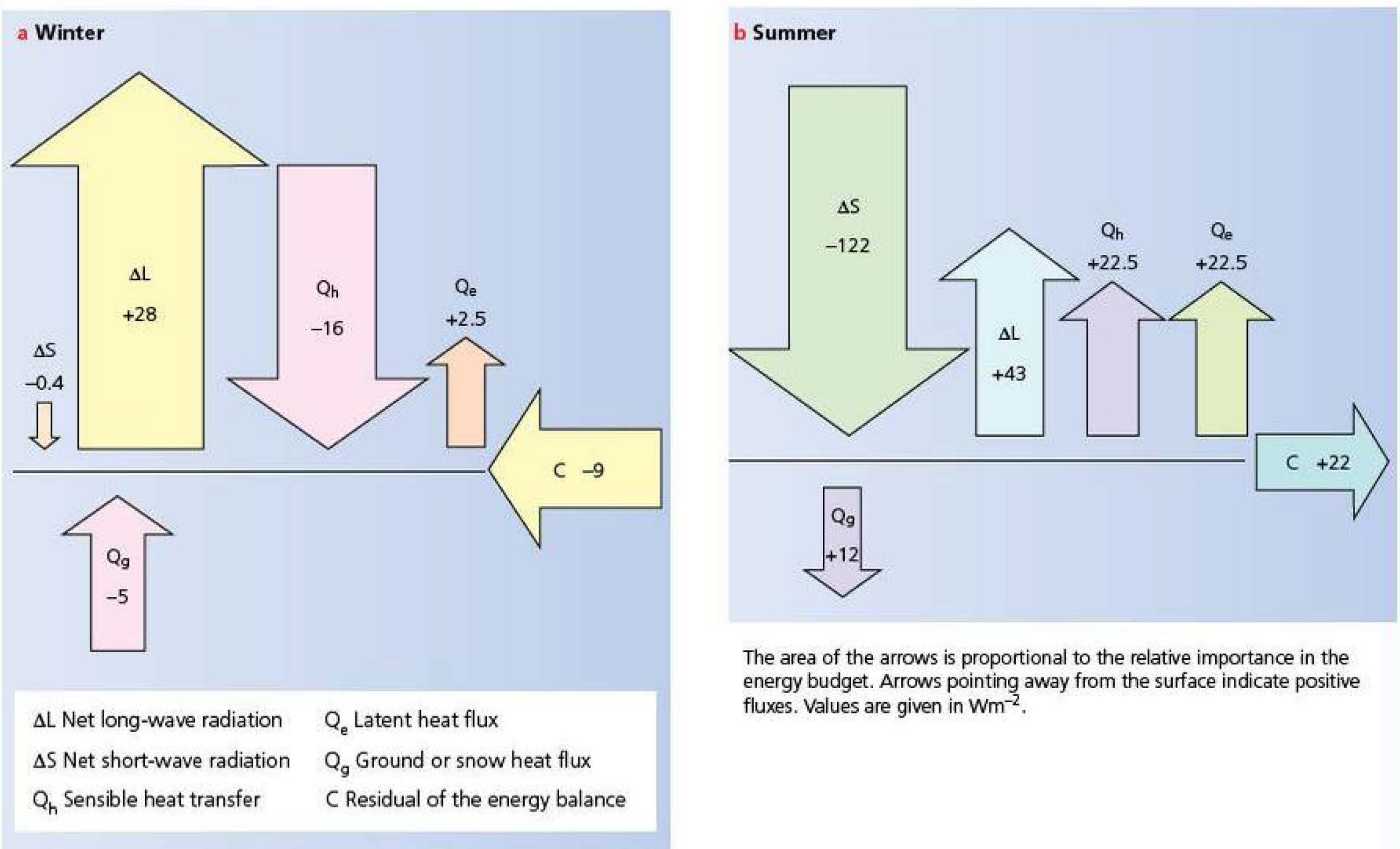


Figure 2.7 Energy budgets for Svalbard

Section 2.1 Activities

1 With reference to Figure 2.7, draw the likely *night-time* energy budgets for Svalbard in summer and in winter.

2 Figure 2.8 shows rural and urban energy budgets for Washington DC (USA) during daytime and night-time. The figures represent the proportions of the original 100 units of incoming solar radiation dispersed in different directions.

  a How does the amount of insolation received vary between the rural area and the urban area?

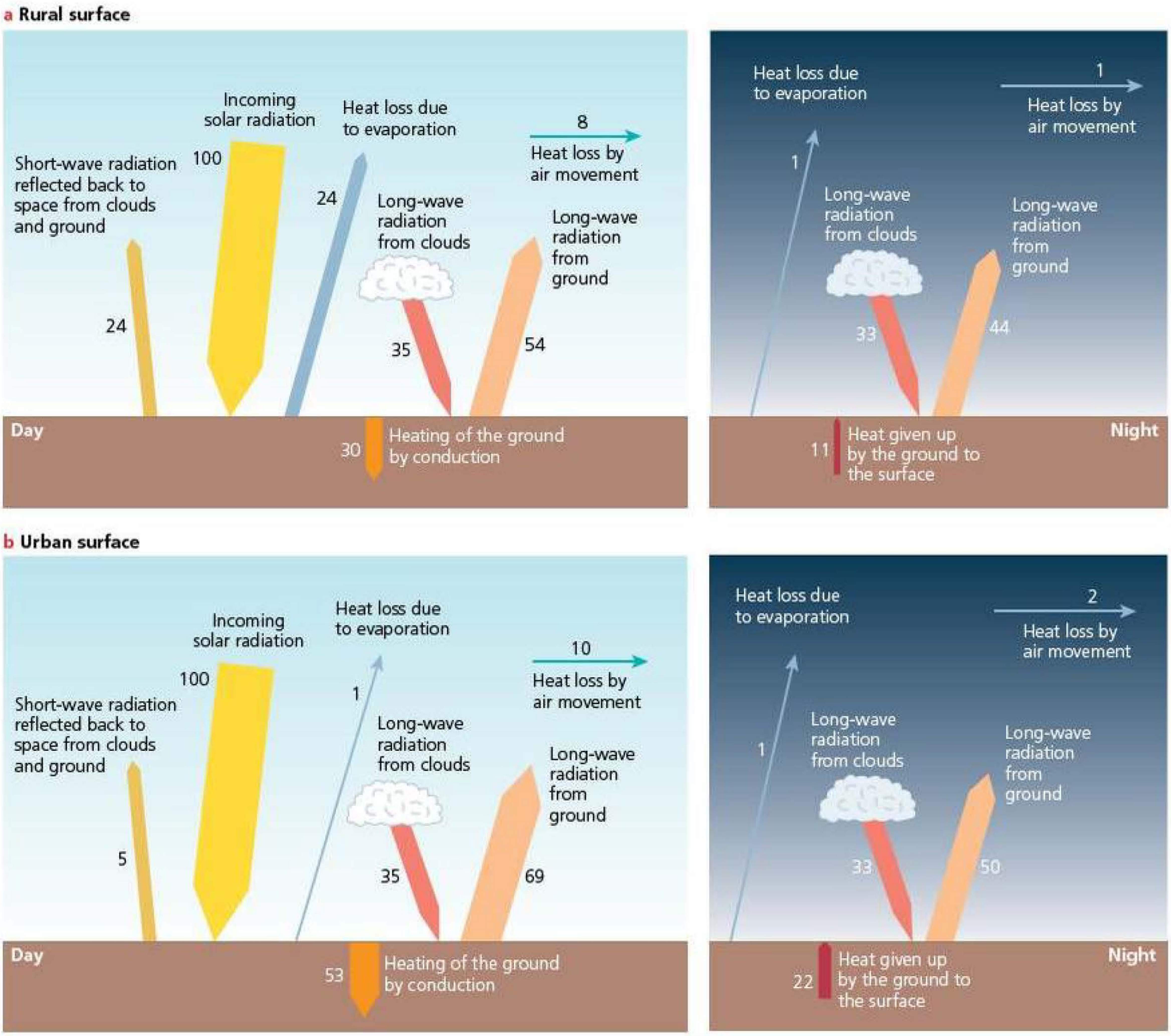
  b How does the amount of heat lost through evaporation vary between the areas? Justify your answer.

  c Explain the difference between the two areas in terms of short-wave radiation reflected to the atmosphere.

  d What are the implications of the answers to b and c for the heating of the ground by conduction?

  e Compare the amount of heat given up by the rural area and the urban area by night. Suggest two reasons for these differences.

  f Why is there more long-wave radiation by night from the urban area than from the rural area?



The figures represent the proportions of the original 100 units of incoming solar radiation dispersed in different directions.

Source: University of Oxford, 1989, Entrance examination for Geography

Figure 2.8 Daytime and night-time energy budgets for Washington DC

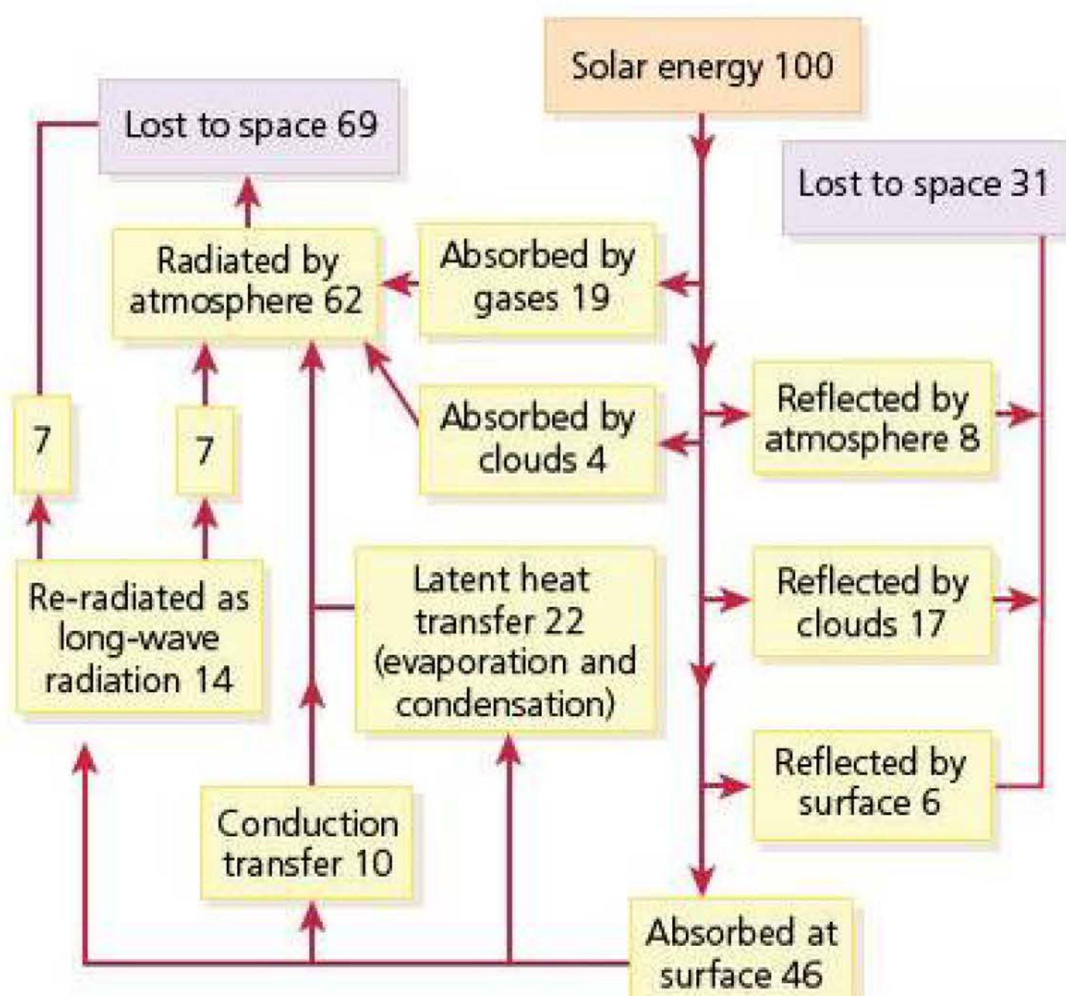
2.2 The global energy budget

The latitudinal pattern of radiation: excesses and deficits

The atmosphere is an open energy system, receiving energy from both Sun and Earth. Although the latter is very small, it has an important local effect, as in the case of urban climates. **In**coming **sol**ar radi**ation** is referred to as insolation.

The atmosphere constantly receives solar energy, yet until recently the atmosphere was not getting any hotter. Therefore there has been a balance between inputs (insolation) and outputs (re-radiation) (Figure 2.9). Under 'natural' conditions the balance is achieved in three main ways:

- **radiation** – the emission of electromagnetic waves such as X-ray, short- and long-wave; as the Sun is a very hot body, radiating at a temperature of about 5700°C, most of its radiation is in the form of very short wavelengths such as ultraviolet and visible light
- **convection** – the transfer of heat by the movement of a gas or liquid
- **conduction** – the transfer of heat by contact.



Source: *Access to Geography – Climate and Society* by G. Nagle (Hodder Education, 2002) p.15

Figure 2.9 The Earth's energy budget

Of incoming radiation, 19 per cent is absorbed by atmospheric gases, especially oxygen and ozone at high altitudes, and carbon dioxide and water vapour at low altitudes. Reflection by the atmosphere accounts for a net loss of 8 per cent, and clouds and water droplets reflect 23 per cent. Reflection from the Earth's surface (known as the planetary albedo) is generally about 6 per cent. About 36 per cent of insolation is reflected back to space and a further 19 per cent is absorbed by atmospheric gases. So only about 46 per cent of the insolation at the top of the atmosphere actually gets through to the Earth's surface.

Energy received by the Earth is re-radiated at long wavelength. (Very hot bodies such as the Sun emit short-wave radiation, whereas cold bodies such as the Earth emit long-wave radiation.) Of this, 8 per cent is lost to space. Some energy is absorbed by clouds and re-radiated back to Earth. Evaporation and condensation account for a loss of heat of 22 per cent. There is also a small amount of condensation (carried up by turbulence). Thus heat gained by the atmosphere from the ground amounts to 32 per cent of incoming radiation.

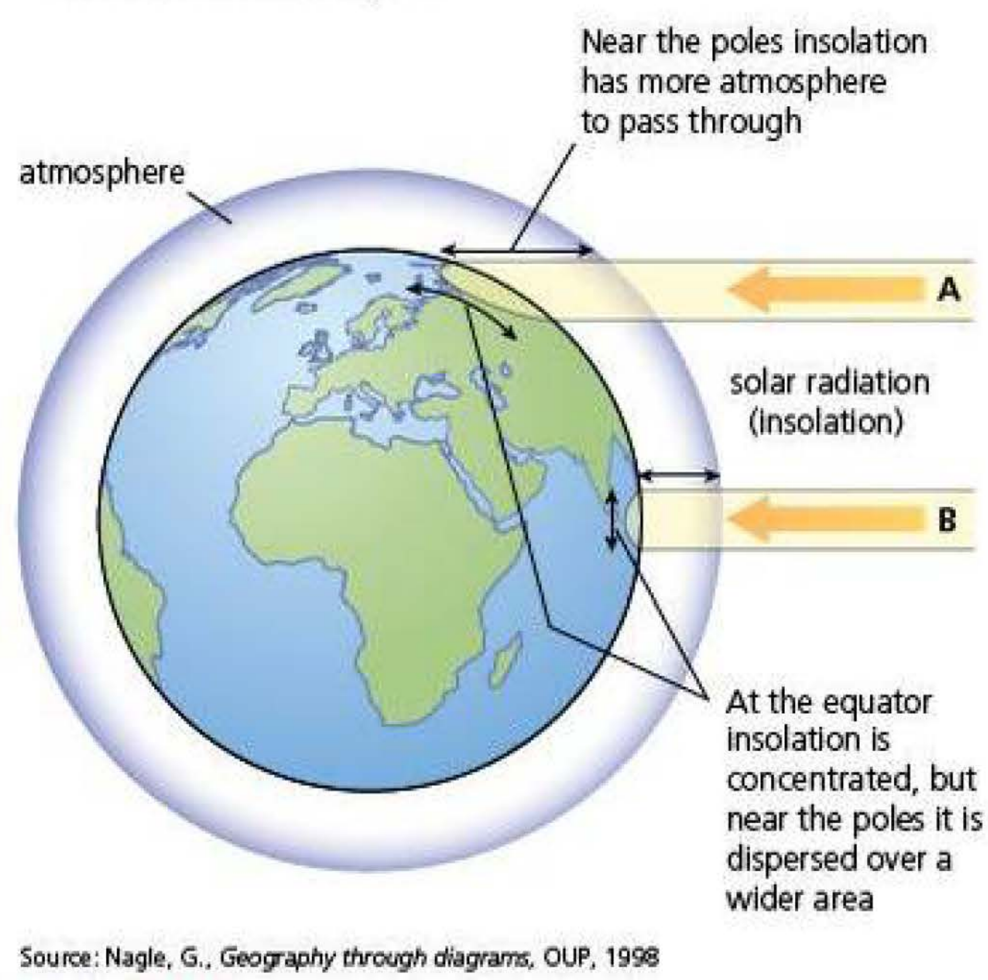
The atmosphere is largely heated from below. Most of the incoming short-wave radiation is let through, but some outgoing long-wave radiation is trapped by greenhouse gases. This is known as the greenhouse principle or greenhouse effect.

There are important variations in the receipt of solar radiation with latitude and season (Figure 2.10). The result is an imbalance: an excess of radiation (positive budget) in the tropics; a deficit of radiation (negative balance) at higher latitudes (Figure 2.11). However, neither region is getting progressively hotter or colder. To achieve this balance, the horizontal transfer of energy from the equator to the poles takes place by winds and ocean currents. This gives rise to an important second energy budget in the atmosphere: the horizontal transfer between low latitudes and high latitudes to compensate for differences in global insolation.

Latitude

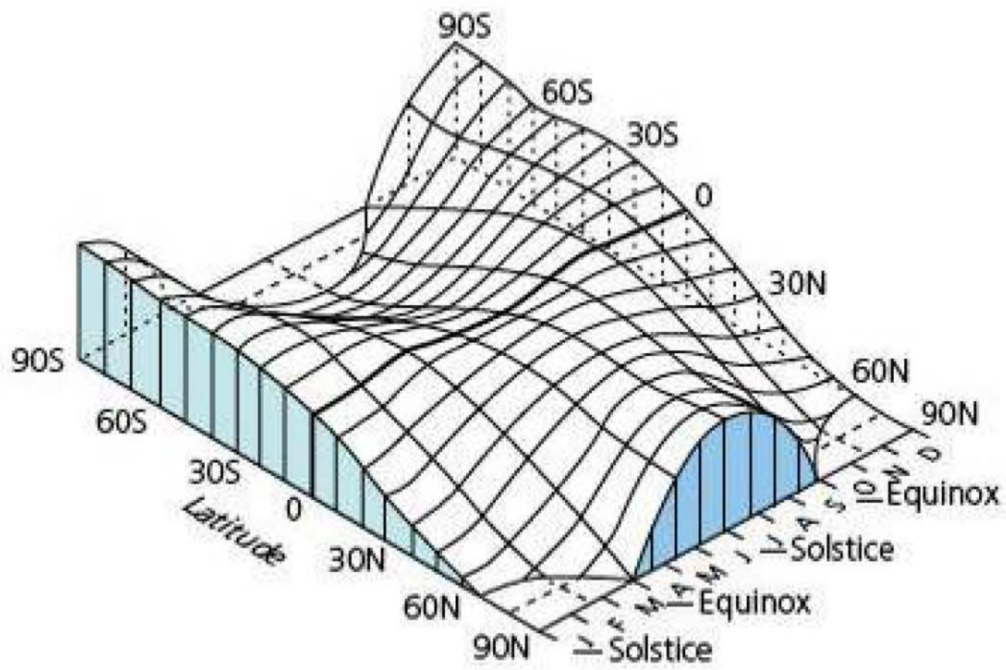
Areas that are close to the equator receive more heat than areas that are close to the poles. This is due to two reasons:

1 Incoming solar radiation (insolation) is concentrated near the equator, but dispersed near the poles.
2 Insolation near the poles has to pass through a greater amount of atmosphere and there is more chance of it being reflected back out to space.



Source: Nagle, G., *Geography through diagrams*, OUP, 1998

Figure 2.10 Latitudinal contrasts in insolation



The variations of solar radiation with latitude and season for the whole globe, assuming no atmosphere. This assumption explains the abnormally high amounts of radiation received at the poles in summer, when daylight lasts for 24 hours each day.

Source: Barry, R. and Chorley, R., *Atmosphere, Weather and Climate*, Routledge, 1998

Figure 2.11 Contrasts in insolation by season and latitude

Section 2.2 Activities

1 Outline the main thermal differences between short-wave and long-wave radiation.
2 Study Figures 2.10 and 2.11. Comment on latitudinal differences in the receipt of solar radiation.

Annual temperature patterns

There are important large-scale north–south temperature zones (Figure 2.12). For example, in January highest temperatures over land (above 30°C) are found in Australia and southern Africa. By contrast, the lowest temperatures (less than –40°C) are found over parts of Siberia, Greenland and the Canadian Arctic. In general, there is a decline in temperatures northwards from the Tropic of Capricorn, although there are important anomalies, such as the effect of the Andes in South America, and the effect of the cold current off the coast of Namibia. In July, maximum temperatures are found over the Sahara, Near East, northern India and parts of southern USA and Mexico. By contrast, areas in the southern hemisphere are cooler than in January.

These patterns reflect the general decrease of insolation from the equator to the poles. There is little seasonal variation at the equator, but in mid or high latitudes large seasonal differences occur due to the decrease in insolation from the equator to the poles, and changes in the length of day. There is also a time lag between the overhead Sun and the period of maximum insolation – up to two months in some places – largely because the air is heated from below, not above. The coolest period is after the winter solstice (the shortest day), since the ground continues to lose heat even after insolation has resumed. Over oceans, the lag time is greater than over the land, due to differences in their specific heat capacities.

Section 2.2 Activities

Describe the differences in temperature as shown in Figure 2.12. Suggest reasons for these contrasts.

Atmospheric transfers

There are two main influences on atmospheric transfer: pressure variations and ocean currents. Air blows from high pressure to low pressure, and is important in redistributing heat around the Earth. In addition, the atmosphere is influenced by ocean currents – warm currents raise the temperature of overlying air, while cold currents cool the air above them (see pages 39–40).

Pressure variations

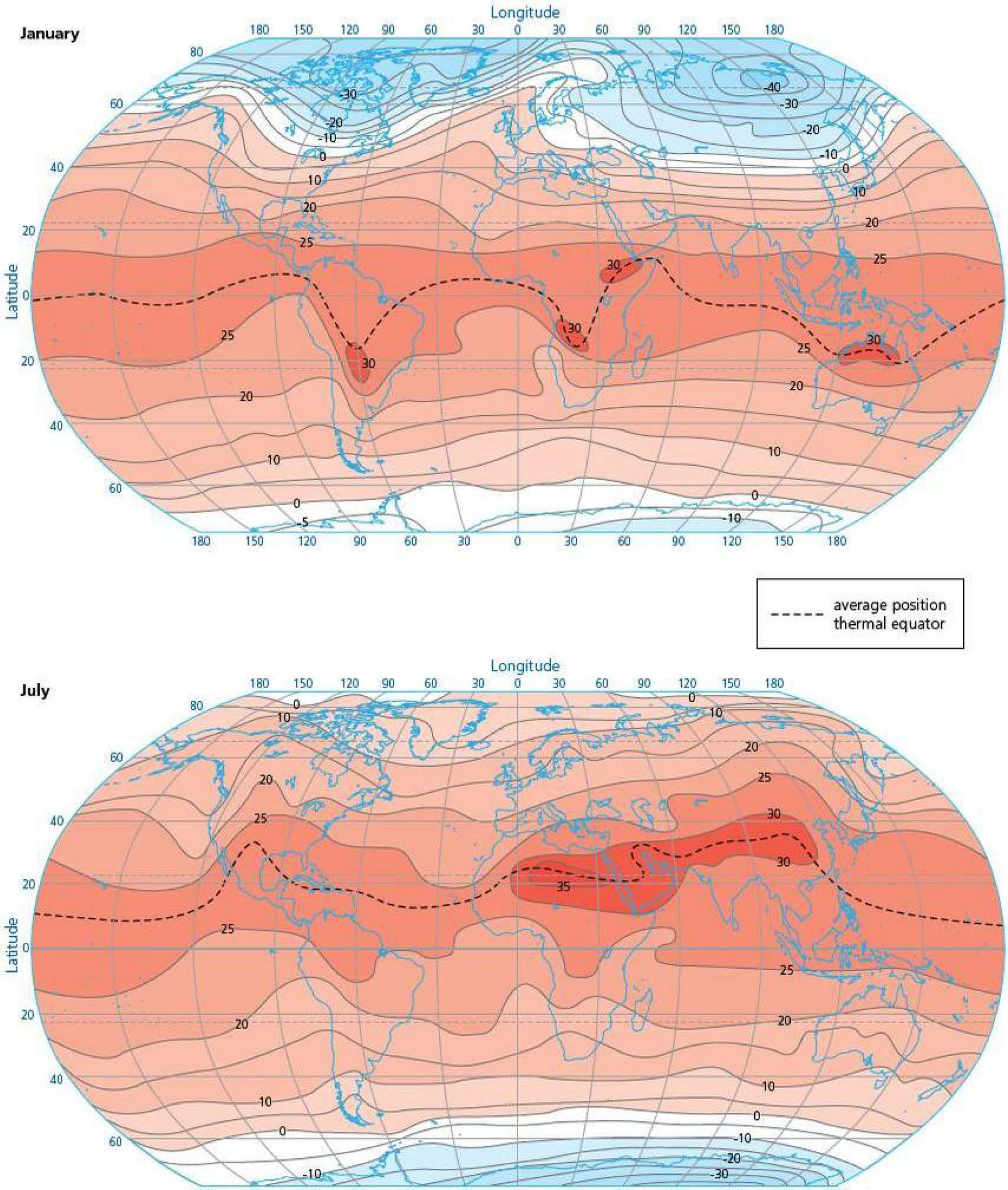
Pressure is measured in millibars (mb) and is represented by isobars, which are lines of equal pressure. On maps, pressure is adjusted to mean sea level (MSL), therefore eliminating elevation as a factor. MSL pressure is 1013mb, although the mean range is from 1060mb in the Siberian winter high-pressure system to 940mb (although some intense low pressure storms may be much lower). The trend of pressure change is more important than the actual reading itself. Decline in pressure indicates poorer weather, and rising pressure better weather.

Surface pressure belts

Sea-level pressure conditions show marked differences between the hemispheres. In the northern hemisphere there are greater seasonal contrasts, whereas in the southern hemisphere much simpler average conditions exist (see Figure 2.13). Over Antarctica there is generally high pressure over the 3–4kilometre-high eastern Antarctic Plateau, but the high pressure is reduced by altitude. The differences are largely related to unequal distribution of land and sea, because ocean areas are much more equable in terms of temperature and pressure variations.

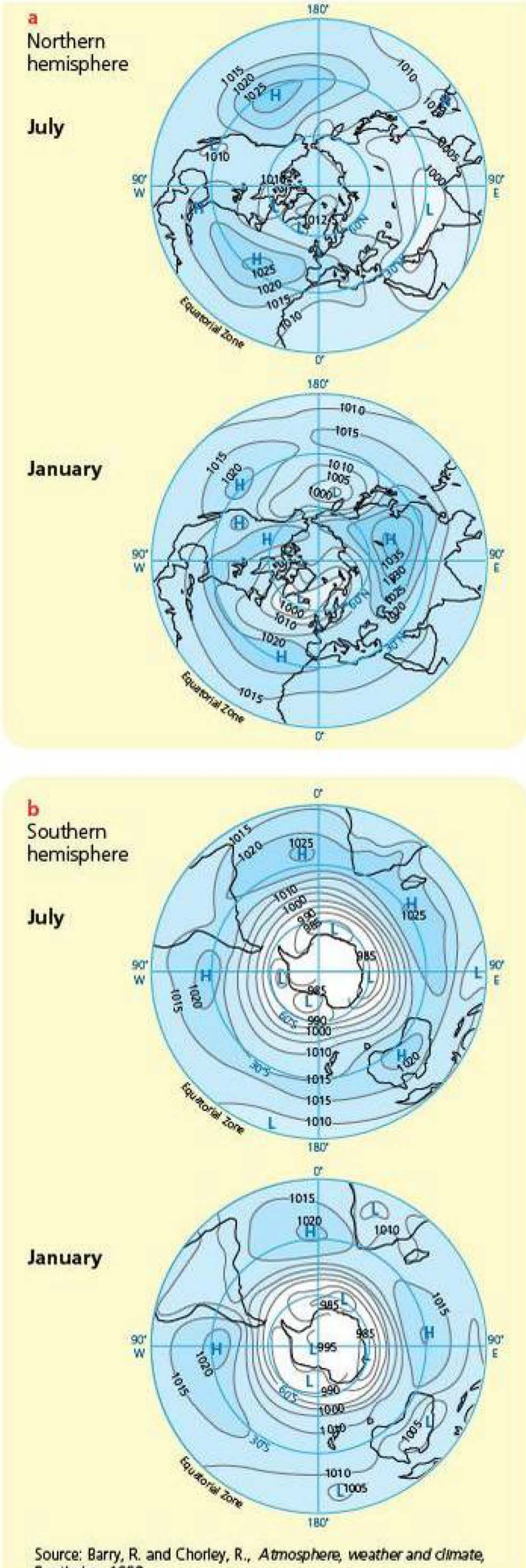
One of the more permanent features is the subtropical high-pressure (STHP) belts, especially over ocean areas. In the southern hemisphere these are almost continuous at about 30° latitude, although in summer over South Africa and Australia they tend to be broken. Generally pressure is about 1026mb. In the northern hemisphere, by contrast, at 30° the belt is much more discontinuous because of the land. High pressure only occurs over the ocean as discrete cells such as the Azores and Pacific highs. Over continental areas such as south-west USA, southern Asia and the Sahara, major fluctuations occur: high pressure in winter, and summer lows because of overheating.

Over the equatorial trough, pressure is low: 1008–1010mb. The trough coincides with the zone of maximum insolation. In the northern hemisphere (in July) it is well north of the equator (25°C over India), whereas in the southern hemisphere (in January) it is just south of the equator because land masses in the southern hemisphere are not



Source: Briggs, D et al., *Fundamentals of the physical environment*, Routledge, 1997

Figure 2.12 Seasonal temperature patterns of sufficient size to displace it southwards. The 'doldrums' refers to the equatorial trough over sea areas, where slack pressure gradients have a becalming effect on sailing ships.



Source: Barry, R. and Chorley, R., *Atmosphere, weather and climate*, Routledge, 1998

Figure 2.13 Variations in pressure

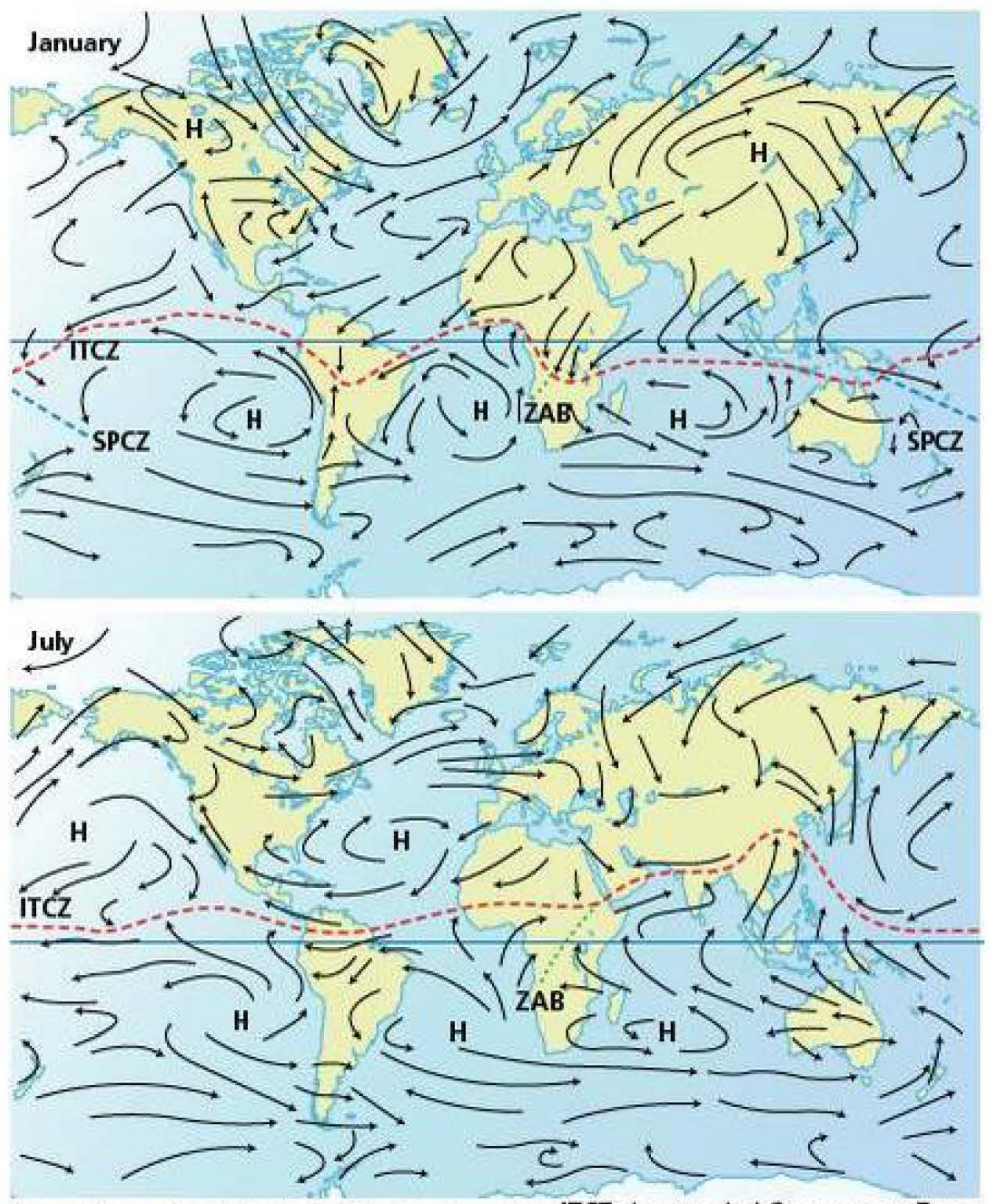
Section 2.2 Activities

Describe the variations in pressure as shown on Figure 2.13.

In temperate latitudes, pressure is generally less than in subtropical areas. The most unique feature is the large number of depressions (low pressure) and anticyclones (high pressure), which do not show up on a map of mean pressure. In the northern hemisphere, there are strong winter low-pressure zones over Icelandic and oceanic areas, but over Canada and Siberia there is high pressure, due to the coldness of the land. In summer, high pressure is reduced. In polar areas pressure is relatively high throughout the year, especially over Antarctica, owing to the coldness of the land mass.

Surface wind belts

Winds between the Tropics converge on a line known as the intertropical convergence zone (ITCZ) or equatorial trough (Figure 2.14). This convergence zone is a few hundred kilometres wide, into which winds blow inwards and subsequently rise (thereby forming an area of low pressure). The rising air releases vast quantities of latent heat, which in turn stimulates convection.



Source: Linacre, E. and Geerts, B., *Climates and Weather Explained*, Routledge, 1997

Figure 2.14 Surface winds

Latitudinal variations in the ITCZ occur as a result of the movement of the overhead Sun. In June the ITCZ lies further north, whereas in December it lies in the southern hemisphere. The seasonal variation in the ITCZ is greatest over Asia, owing to its large land mass. By contrast, over the Atlantic and Pacific Oceans its movement is far less. Winds at the ITCZ are generally light (the doldrums), occasionally broken by strong westerlies, generally in the summer months.

Low-latitude winds between 10° and 30° are mostly easterlies; that is, they flow towards the west. These are the reliable trade winds; they blow over 30 per cent of the world's surface. The weather in this zone is fairly predictable: warm, dry mornings and showery afternoons, caused by the continuous evaporation from tropical seas. Showers are heavier and more frequent in the warmer summer season.

Occasionally there are disruptions to the pattern; easterly waves are small-scale systems in the easterly flow of air. The flow is greatest not at ground level but at the 700 mb level. Ahead of the easterly wave, air is subsiding; hence there is surface divergence. At the easterly wave, there is convergence of air, and ascent – as in a typical low pressure system. Easterly waves are important for the development of tropical cyclones (Section 9.3).

Westerly winds dominate between 35° and 60° of latitude, which accounts for about a quarter of the world's surface. However, unlike the steady trade winds, these contain rapidly evolving and decaying depressions.

The word 'monsoon' means 'reverse'; the monsoon is reversing wind systems. For example, the south-east trades from the southern hemisphere cross the equator in July. Owing to the Coriolis force, these south-east trades are deflected to the right in the northern hemisphere and become south-west winds. The monsoon is induced by Asia – the world's largest continent – which causes winds to blow outwards from high pressure in winter, but pulls the southern trades into low pressure in the summer.

The monsoon is therefore influenced by the reversal of land and sea temperatures between Asia and the Pacific during the summer and winter. In winter, surface temperatures in Asia may be as low as –20 °C. By contrast, the surrounding oceans have temperatures of 20 °C. During the summer, the land heats up quickly and may reach 40 °C. By contrast, the sea remains cooler at about 27 °C. This initiates a land–sea breeze blowing from the cooler sea (high pressure) in summer to the warmer land (low pressure), whereas in winter air flows out of the cold land mass (high pressure) to the warm water (low pressure). The presence of the Himalayan Plateau also disrupts the strong winds of the upper atmosphere, forcing winds either to the north or south and consequently deflecting surface winds.

The uneven pattern shown in Figure 2.14 is the result of seasonal variations in the overhead Sun. Summer in the southern hemisphere means that there is a cooling in the northern hemisphere, thereby increasing the differences between polar and equatorial air. Consequently, high-level westerlies are stronger in the northern hemisphere in winter.

Section 2.2 Activities

Describe the main global wind systems shown in Figure 2.14.

Explaining variations in temperature, pressure and winds

Latitude

On a global scale, latitude is the most important factor determining temperature (Figure 2.10). Two factors affect the temperature: the angle of the overhead Sun and the thickness of the atmosphere. At the equator, the overhead Sun is high in the sky, so the insolation received is of a greater quality or intensity. At the poles, the overhead Sun is low in the sky, so the quality of energy received is poor. Secondly, the thickness of the atmosphere affects temperature. Energy has more atmosphere to pass through at point A on Figure 2.10, so more energy is lost, scattered or reflected by the atmosphere than at B – therefore temperatures are lower at A than at B. In addition, the albedo (reflectivity) is higher in polar regions. This is because snow and ice are very reflective, and low-angle sunlight is easily reflected from water surfaces. However, variations in length of day and season partly offset the lack of intensity in polar and arctic regions. The longer the Sun shines, the greater the amount of insolation received, which may overcome in part the lack of intensity of insolation in polar regions. (On the other hand, the long polar nights in winter lose vast amounts of energy.)

Land–sea distribution

There are important differences in the distribution of land and sea in the northern hemisphere and southern hemisphere. There is much more land in the northern hemisphere. Oceans cover about 50 per cent of the Earth's surface in the northern hemisphere but about 90 per cent of the southern hemisphere (Figure 2.15). This is not always clear when looking at conventional map projections such as the Mercator projection.

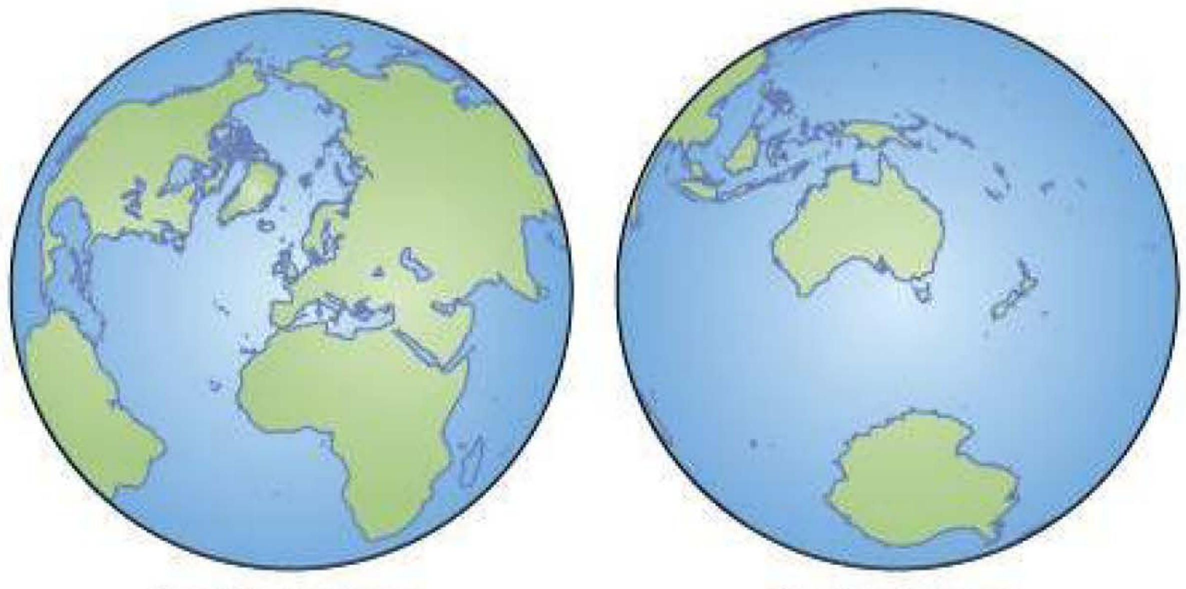


Figure 2.15 Land and sea hemispheres

The distribution of land and sea is important because land and water have different thermal properties. The specific heat capacity is the amount of heat needed to raise the temperature of a body by 1 °C. There are important differences between the heating and cooling of water. Land heats and cools more quickly than water. It takes five times as much heat to raise the temperature of water by 2 °C as it does to raise land temperatures.

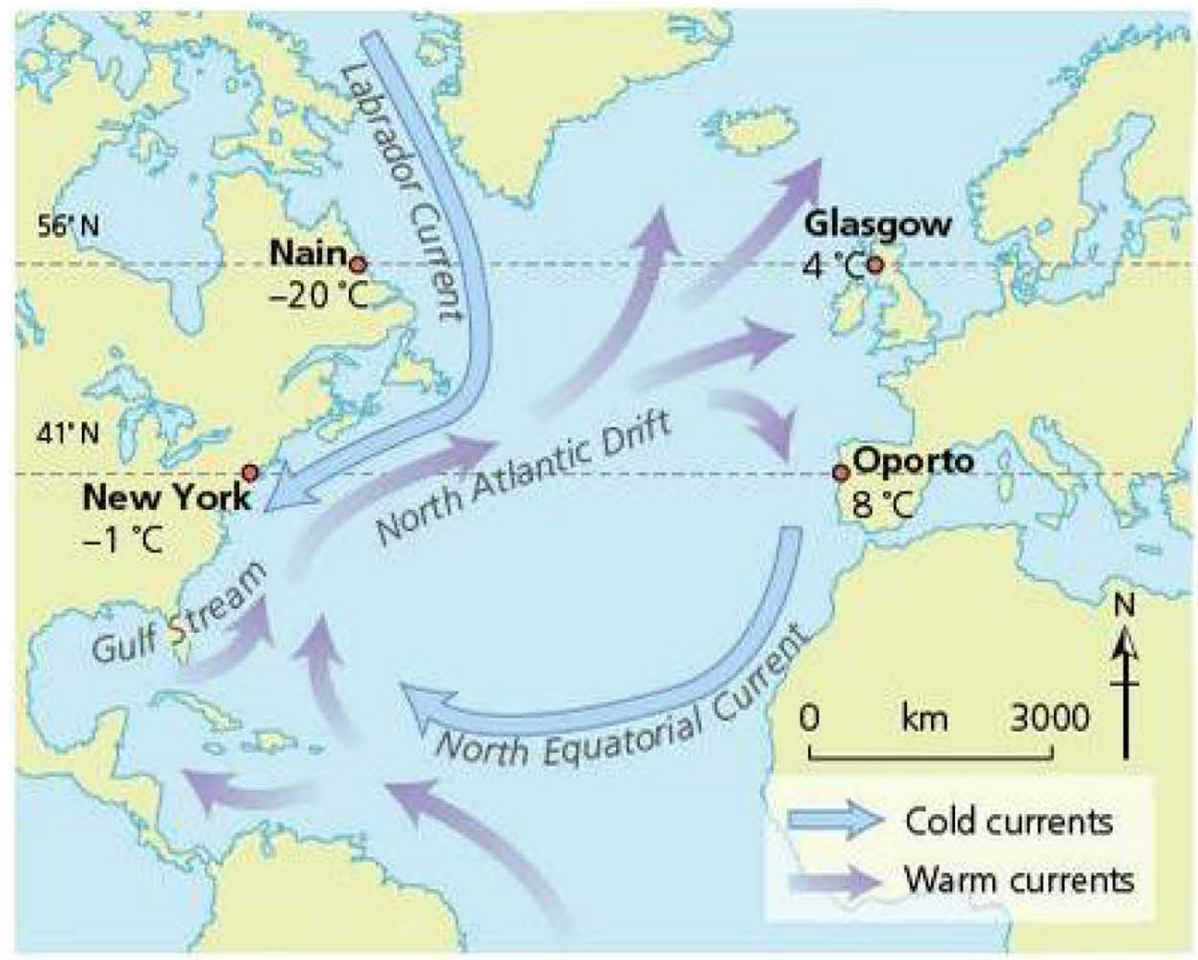
Water heats more slowly because:

- it is clear, so the Sun's rays penetrate to great depth, distributing energy over a wider area
- tides and currents cause the heat to be further distributed.

Therefore a larger volume of water is heated for every unit of energy than the volume of land, so water takes longer to heat up. Distance from the sea has an important influence on temperature. Water takes up heat and gives it back much more slowly than the land. In winter, in mid-latitudes sea air is much warmer than the land air, so onshore winds bring heat to the coastal lands. By contrast, during the summer coastal areas remain much cooler than inland sites. Areas with a coastal influence are termed maritime or oceanic, whereas inland areas are called continental.

Ocean currents

Surface ocean currents are caused by the influence of prevailing winds blowing steadily across the sea. The dominant pattern of surface ocean currents (known as gyres) is roughly a circular flow. The pattern of these currents is clockwise in the northern hemisphere and anti-clockwise in the southern hemisphere. The main exception is the circumpolar current that flows around Antarctica from west to east. There is no equivalent current in the northern hemisphere because of the distribution of land and sea there. Within the circulation of the gyres, water piles up into a dome. The effect of the rotation of the Earth is to cause water in the oceans to push westward; this piles up water on the western edge of ocean basins – rather like water slopping in a bucket. The return flow is often narrow, fast-flowing currents such as the Gulf Stream. The Gulf Stream in particular transports heat northwards and then eastwards across the North Atlantic; the Gulf Stream is the main reason that the British Isles have mild winters and relatively cool summers (Figure 2.16).



The effect of an ocean current depends upon whether it is a warm current or a cold current. Warm currents move away from the equator, whereas cold currents move towards it. The cold Labrador Current reduces the temperatures of the western side of the Atlantic, while the warm North Atlantic Drift raises temperatures on the eastern side.

Source: Nagle, G., *Geography through diagrams*, OUP, 1998

Figure 2.16 The effects of the North Atlantic Drift/Gulf Stream

The effect of ocean currents on temperatures depends upon whether the current is cold or warm. Warm currents from equatorial regions raise the temperature of polar areas (with the aid of prevailing westerly winds). However, the effect is only noticeable in winter. For example, the North Atlantic Drift raises the winter temperatures of north-west Europe. By contrast, other areas are made colder by ocean currents. Cold currents such as the Labrador Current off the north-east coast of North America may reduce summer temperatures, but only if the wind blows from the sea to the land.

In the Pacific Ocean, there are two main atmospheric states. The first is warm surface water in the west with cold surface water in the east; the other is warm surface water in the east with cold in the west. In both cases, the warm surface causes low pressure. As air blows from high pressure to low pressure, there is a movement of water from the colder area to the warmer area. These winds push warm surface water into the warm region, exposing colder deep water behind them and maintaining the pattern.

The ocean conveyor belt

In addition to the transfer of energy by wind and the transfer of energy by ocean currents, there is also a transfer of energy by deep sea currents. Oceanic convection movement is from polar regions where cold salty water sinks into the depths and makes its way towards the equator (Figure 2.17). The densest water is found in the Antarctic, where sea water freezes to form ice at a temperature of around –2 °C. The ice is fresh water, so the sea water that is left behind is much saltier and therefore denser. This cold dense water sweeps around Antarctica at a depth of about 4 kilometres. It then spreads into the deep basins of the Atlantic, the Pacific and the Indian Oceans. In the oceanic conveyor-belt model, surface currents bring warm water to the North Atlantic from the Indian and Pacific Oceans. These waters give up their heat to cold winds that blow from Canada across the North Atlantic. This water then sinks and starts the reverse convection of the deep ocean current. The amount of heat given up is about a third of the energy that is received from the Sun. The pattern is maintained by salt: because the conveyor operates in this way, the North Atlantic is warmer than the North Pacific, so there is proportionally more evaporation there. The water left behind by evaporation is saltier and therefore much denser, which causes it to sink. Eventually, the water is transported into the Pacific where it picks up more water and its density is reduced.

Section 2.2 Activities

Outline the main factors affecting global and regional temperatures.

Factors affecting air movement

Pressure and wind

Vertical air motion is important on a local scale, whereas horizontal motion (wind) is important at many scales, from small-scale eddies to global wind systems. The basic cause of air motion is the unequal heating of the Earth's surface. The major equalising factor is the transfer of heat by air movement. Variable heating of the Earth causes variations in pressure and this in turn sets the air in

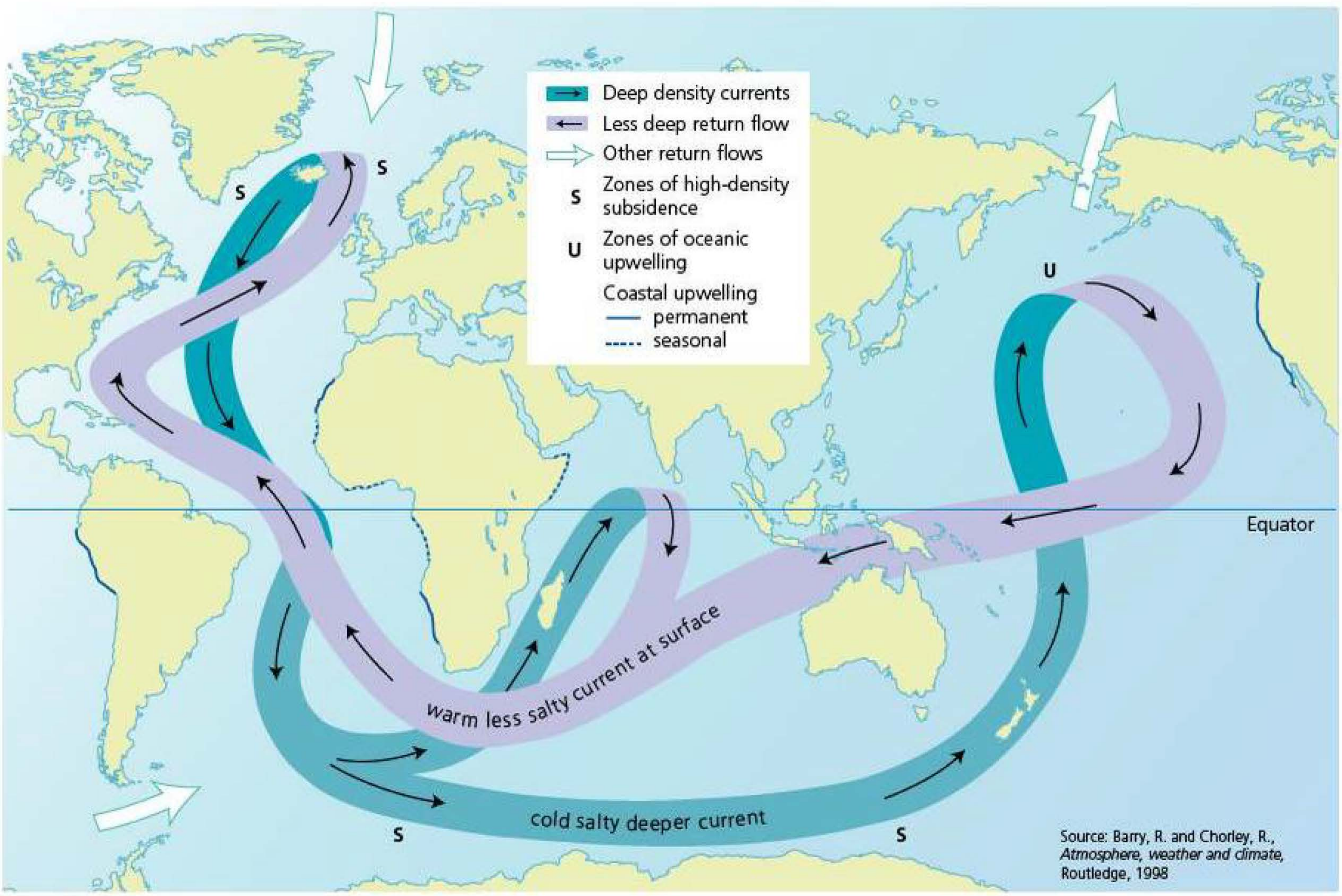
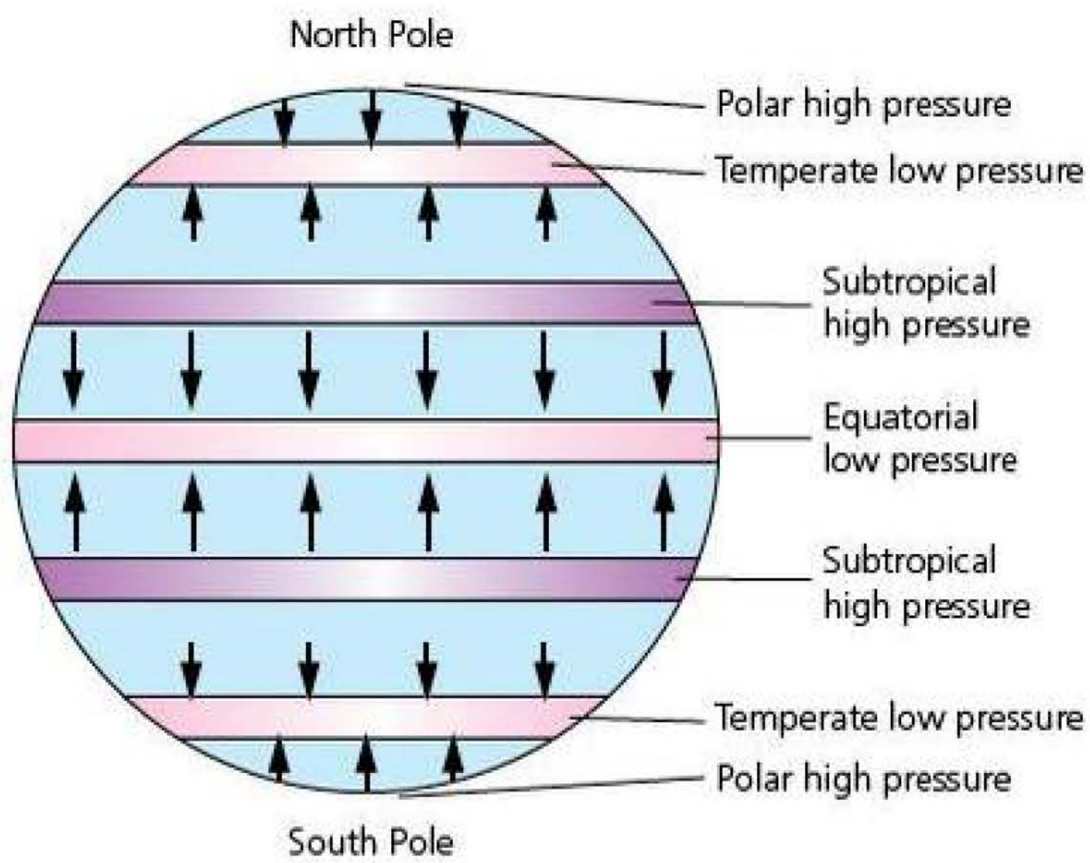


Figure 2.17 The ocean conveyor belt

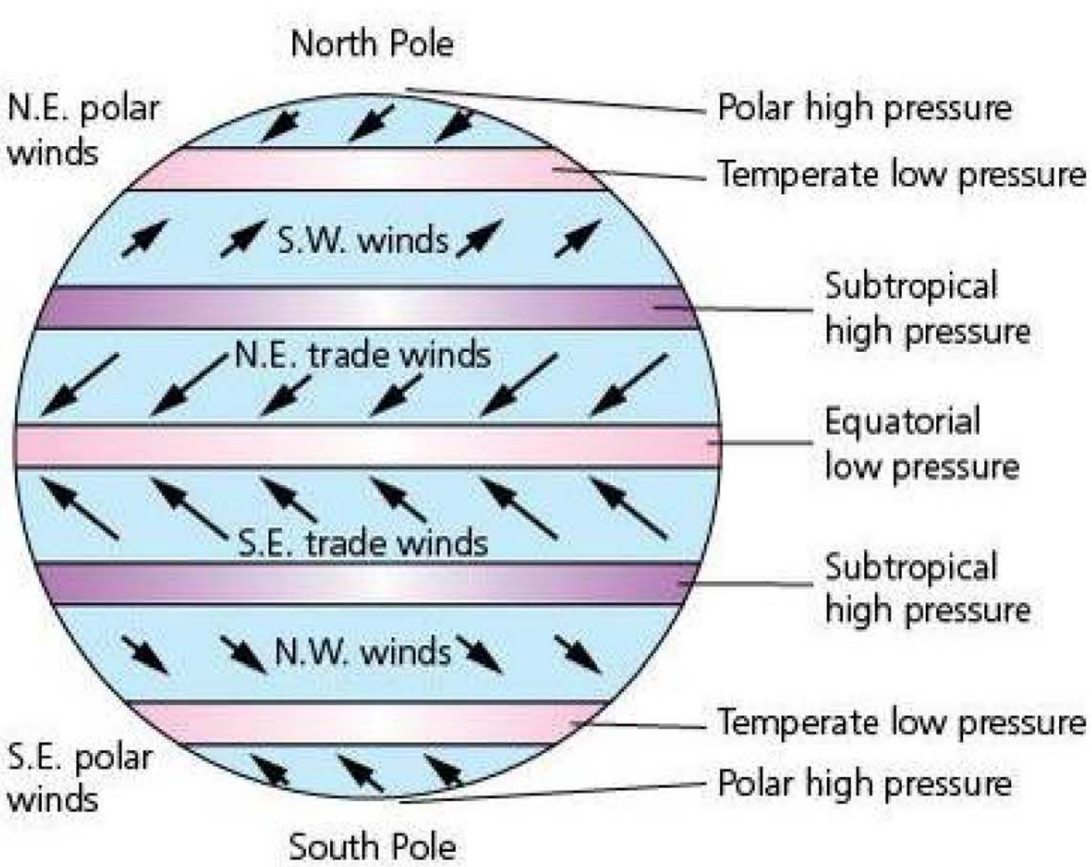
motion. There is thus a basic correlation between winds and pressure.

Pressure gradient

The driving force is the pressure gradient; that is, the difference in pressure between any two points. Air blows from high pressure to low pressure (Figure 2.18). Globally, very high pressure conditions exist over Asia in winter due to the low temperatures. Cold air contracts, leaving room for adjacent air to converge at high altitude, adding to the weight and pressure of the air. By contrast, the mean sea-level pressure is low over continents in summer. High surface temperatures produce atmospheric expansion and therefore a reduction in air pressure. High pressure dominates at around 25–30° latitude. The highs are centred over the oceans in summer and over the continents in winter – whichever is cooler.



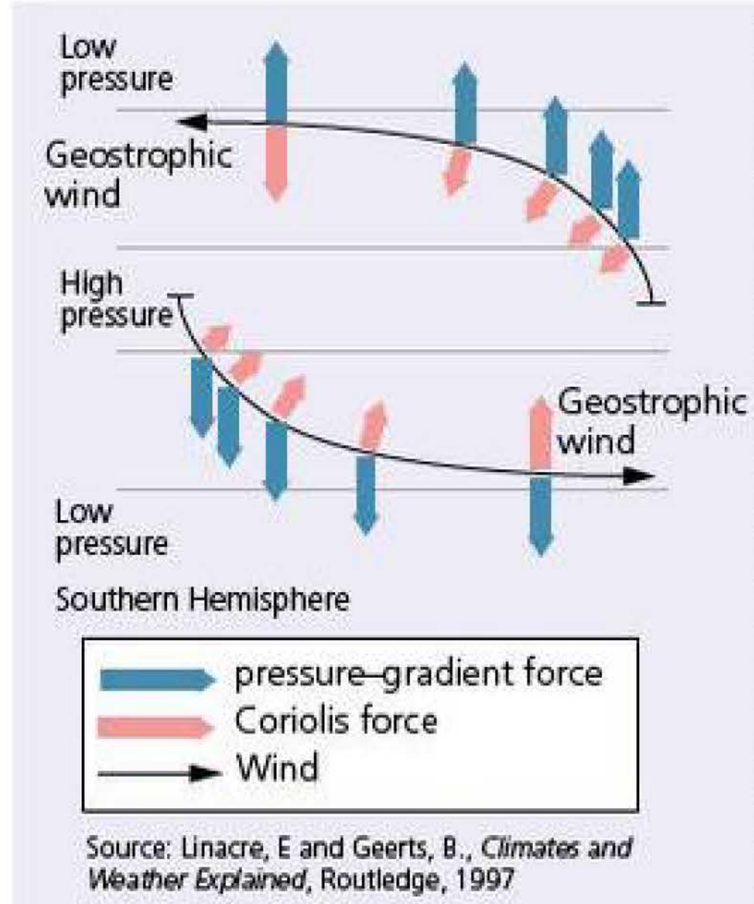
a How the winds would blow on a non-rotating Earth



b How the winds blow on a rotating Earth

Figure 2.18 Pressure gradient winds

The Coriolis force is the deflection of moving objects caused by the easterly rotation of the Earth (Figure 2.19). Air flowing from high pressure to low pressure is deflected to the right of its path in the northern hemisphere and to the left of its path in the southern hemisphere. The Coriolis force is at right angles to wind direction.



The apparent deflection of a parcel of air moving from a belt of high pressure in the southern hemisphere (e.g. from the band of subtropical high pressures). The parcel is assumed stationary initially. As soon as it starts to move, it suffers a sideways Coriolis force, increasing in proportion to its acceleration. The force deflects the parcel until it is travelling along an isobar, with a constant speed such that the Coriolis force balances the pressure–gradient force.

Figure 2.19 The Coriolis force

Every point on the Earth completes one rotation every 24 hours. Air near the equator travels a much greater distance than air near the poles. Air that originates near the equator is carried towards the poles, taking with it a vast momentum. The Coriolis force deflects moving objects to the right of their path in the northern hemisphere and to the left of their path in the southern hemisphere.

The balance of forces between the pressure gradient force and the Coriolis force is known as the geostrophic balance and the resulting wind is known as a geostrophic wind. The geostrophic wind in the northern hemisphere blows anti-clockwise around the centre of low pressure and clockwise around the centre of high pressure.

This centrifugal force is the outward force experienced when you drive a vehicle around a corner. The centrifugal force acts at right angles to the wind, pulling objects outwards, so for a given pressure, airflow is faster around high pressure (because the pressure gradient and centrifugal forces work together rather than in opposite directions).

The drag exerted by the Earth's surface is also important. Friction decreases wind speed, so it decreases the Coriolis force, hence air is more likely to flow towards low pressure.

Section 2.2 Activities

Briefly explain the meaning of the terms **a** *pressure gradient force* and **b** *Coriolis force*.

General circulation model

In general:

- warm air is transferred polewards and is replaced by cold air moving towards the equator
- air that rises is associated with low pressure, whereas air that sinks is associated with high pressure
- low pressure produces rain; high pressure produces dry conditions.

Any circulation model must take into account the meridional (north/south) transfer of heat, and latitudinal variations in rainfall and winds. (Any model is descriptive and static – unlike the atmosphere.) In 1735, George Hadley described the operation of the Hadley cell, produced by the direct heating over the equator. The air here is forced to rise by convection, travels polewards and then sinks at the subtropical anticyclone (high-pressure belt). Hadley suggested that similar cells might exist in mid-latitudes and high latitudes. William Ferrel suggested that Hadley cells interlink with a mid-latitude cell, rotating it in the reverse direction, and these cells in turn rotate the polar cell.

There are very strong differences between surface and upper winds in tropical latitudes. Easterly winds at the surface are replaced by westerly winds above, especially in winter. At the ITCZ, convectional storms lift air into the atmosphere, which increases air pressure near the tropopause, causing winds to diverge at high altitude. They move out of the equatorial regions towards the poles, gradually losing heat by radiation. As they contract, more air moves in and the weight of the air increases the air pressure at the subtropical high-pressure zone (Figure 2.20). The denser air sinks, causing subsidence (stability). The north/south component of the Hadley cell is known as a meridional flow. The Ferrel Cell was originally considered to be a thermally indirect cell (driven by the Hadley cell and polar cell). Now it is known to be more complex, and there is some equator-ward movement of air related to temperate high- and low-pressure systems. These are related to Rossby waves and jet streams (Figure 2.20c).

The zonal flow (east–west) over the Pacific was discovered by Gilbert Walker in the 1920s. The Southern Oscillation Index (SOI) is a measure of how far temperatures vary from the 'average'. A high SOI is associated with strong westward trades (because winds near the equator blow from high pressure to low pressure and are unaffected by the Coriolis force). Tropical cyclones are more common in the South Pacific when there is an El Niño Southern Oscillation warm episode.

The polar cell is found in high latitudes. Winds at the highest latitudes are generally easterly. Air over the North Pole continually cools; and being cold, it is dense and therefore it subsides, creating high pressure. Air above the polar front flows back to the North Pole, creating a polar cell. In between the Hadley cell and the polar cell is an indirect cell, the Ferrel cell, driven by the movement of the other two cells, rather like a cog in a chain.

In the early twentieth century, researchers investigated patterns and mechanisms of upper winds and clouds at an altitude of between 3 and 12 kilometres. They identified large-scale fast-moving belts of westerly winds, which follow a ridge and trough wave-like pattern known as Rossby waves or planetary waves (Figure 2.21). The presence of these winds led to Rossby's 1941 model of the atmosphere. This suggested a three-cell north/south (meridional) circulation, with two thermally direct cells and one thermally indirect cell. The thermally direct cell is driven by the heating at the equator (the Hadley cell) and by the sinking of cold air at the poles (the polar cell). Between them lies the thermally indirect cell whose energy is obtained from the cells to either side by the mixing of the atmosphere at upper levels. The jet streams are therefore key locations in the transfer of energy through the atmosphere. Further modifications of Rossby's models were made by Palmen in 1951.

New models change the relative importance of the three convection cells in each hemisphere. These changes are influenced by jet streams and Rossby waves:

- Jet streams are strong, regular winds that blow in the upper atmosphere about 10 km above the surface; they blow between the poles and tropics (100–300 km/h).
- There are two jet streams in each hemisphere – one between 30° and 50°; the other between 20° and 30°. In the northern hemisphere, the polar jet and the subtropical jet flow eastwards.
- Rossby waves are 'meandering rivers of air' formed by westerly winds. There are three to six waves in each hemisphere. They are formed by major relief barriers such as the Rockies and the Andes, by thermal differences and uneven land–sea interfaces.

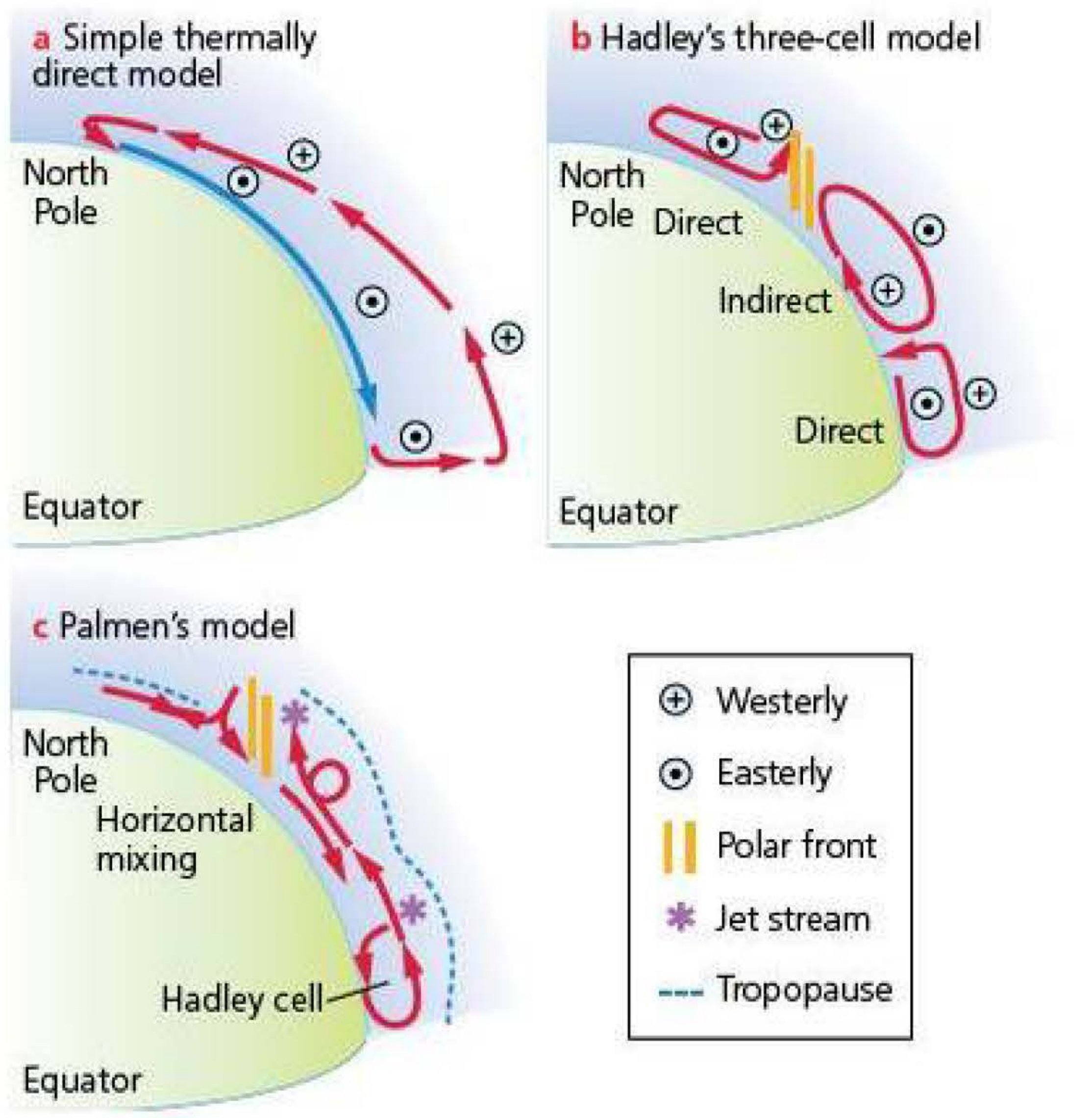


Figure 2.20 General circulation model

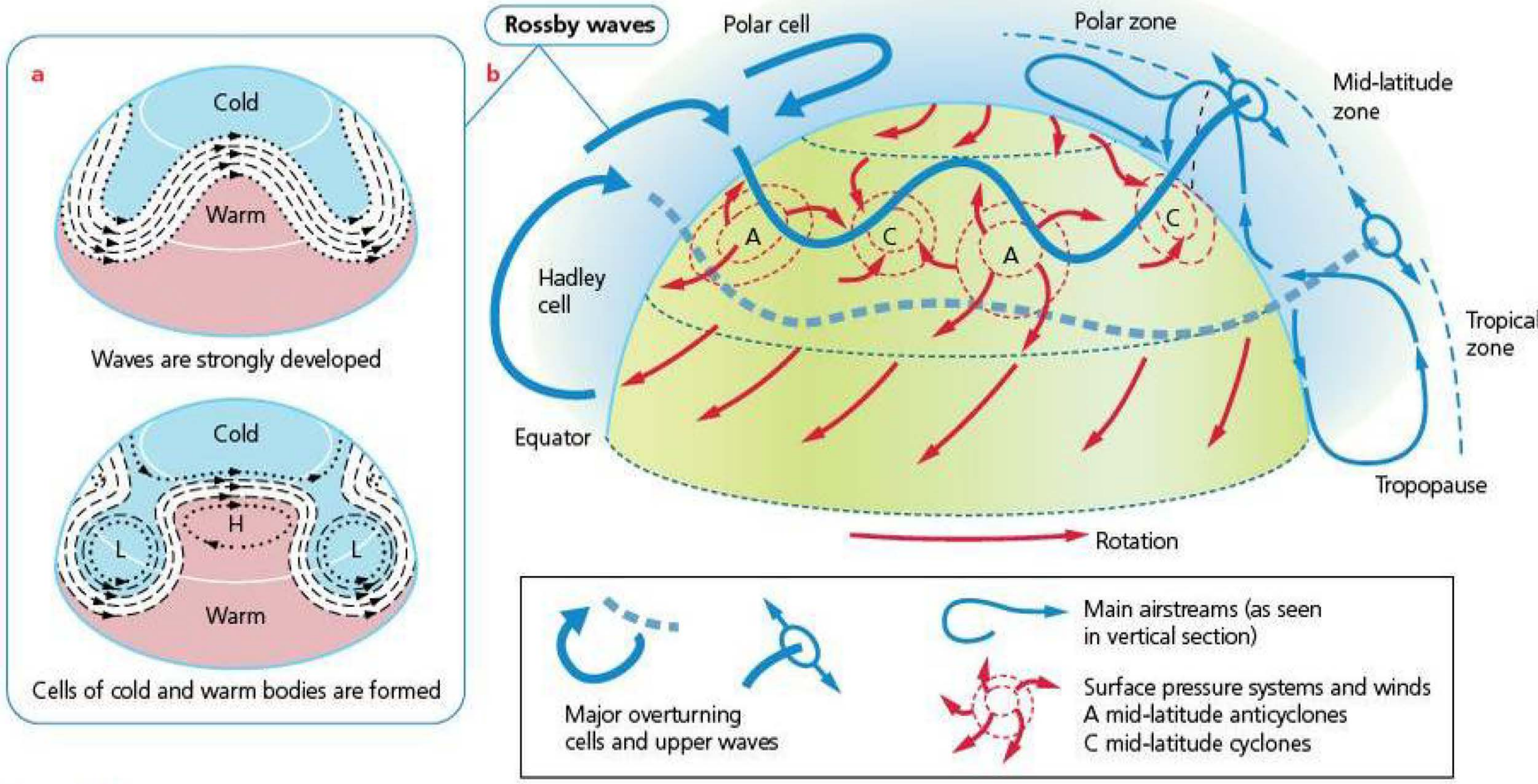


Figure 2.21 Rossby waves

- The jet streams result from differences in equatorial and sub-tropical air, and between polar and sub-tropical air. The greater the temperature difference, the stronger the jet stream.

Rossby waves are affected by major topographic barriers such as the Rockies and the Andes. Mountains create a wave-like pattern, which typically lasts six weeks. As the pattern becomes more exaggerated (Figure 2.21b), it leads to blocking anticyclones (blocking highs) – prolonged periods of unusually warm weather.

Jet streams and Rossby waves are an important means of mixing warm and cold air.

Section 2.2 Activities

1 Describe and explain how the Hadley cell operates.
2 Define the term *Rossby wave*. Suggest how an understanding of Rossby waves may help in our understanding of the general circulation.

2.3 Weather processes and phenomena

Atmospheric moisture processes

Atmospheric moisture exists in all three states – vapour, liquid and solid (Figures 2.22–2.24). Energy is used in the change from one phase to another, for example between a liquid and a gas. In evaporation, heat is absorbed. It takes 600 calories of heat to change 1 gram of water from a liquid to a vapour. Heat loss during evaporation passes into the water as latent heat (of vaporisation). This would cool 1 kilogram of air by 2.5 °C. By contrast, when condensation occurs, latent heat locked in the water vapour is released, causing a rise in temperature. In the changes between vapour and ice, heat is released when vapour is converted to ice (solid), for example rime at high altitudes and high latitudes. In contrast, heat is absorbed in the process of sublimation, for example when snow patches disappear without melting. When liquid water turns to ice, heat is released and temperatures drop. In contrast, in melting ice heat is absorbed and temperatures rise.

Figure 2.22 Atmospheric moisture – condensation

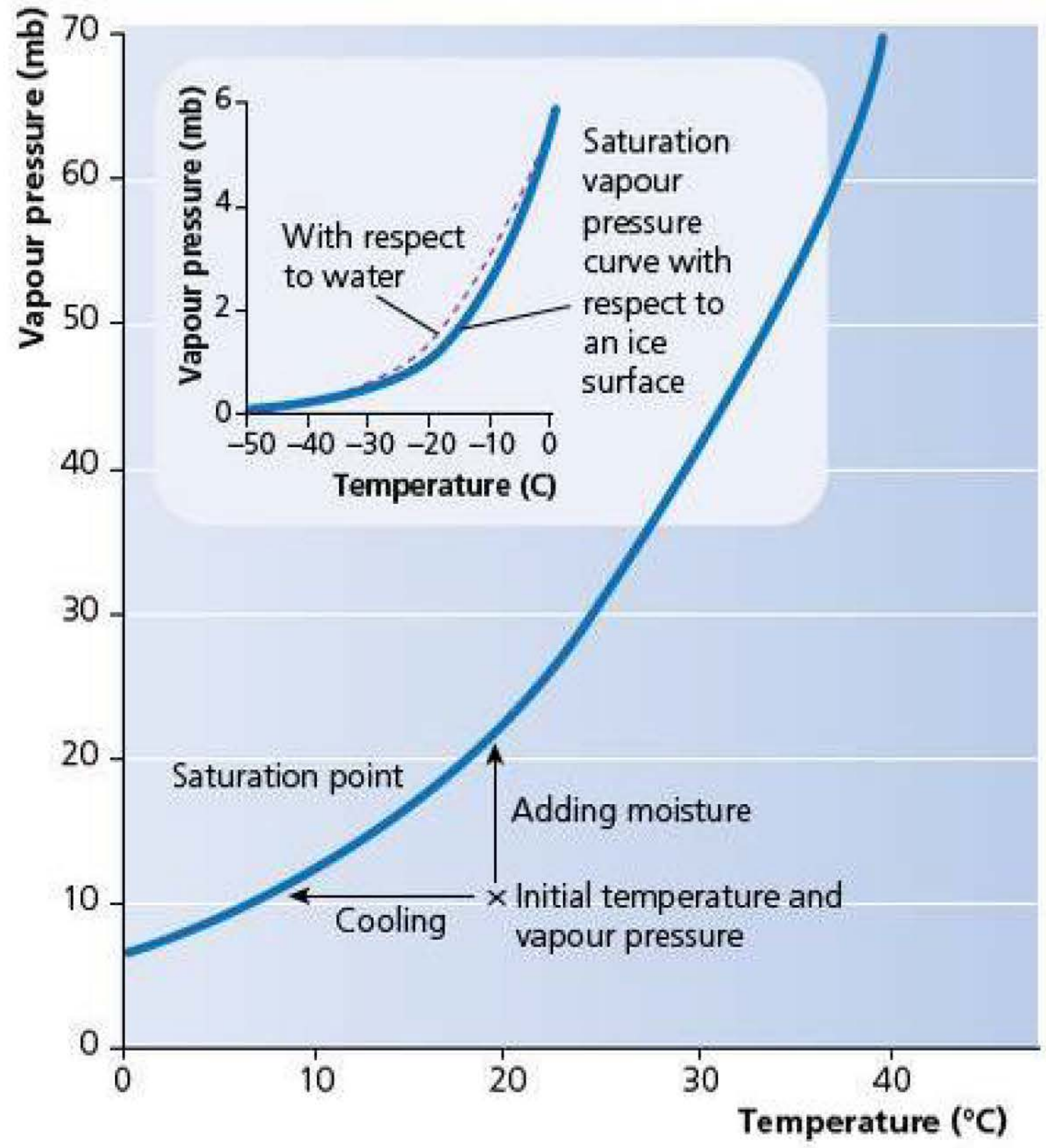
Figure 2.23 Radiation fog in the lower part of alpine valleys

Figure 2.24 Moisture in its liquid state – Augher Lake, Gap of Dunloe, Killarney, Ireland

Factors affecting evaporation

Evaporation occurs when vapour pressure of a water surface exceeds that in the atmosphere. Vapour pressure is the pressure exerted by the water vapour in the atmosphere. The maximum vapour pressure at any temperature occurs when the air is saturated (Figure 2.25). Evaporation aims to equalise the pressures. It depends on three main factors:

- **initial humidity of the air** – if air is very dry then strong evaporation occurs; if it is saturated then very little occurs
- **supply of heat** – the hotter the air, the more evaporation that takes place
- **wind strength** – under calm conditions the air becomes saturated rapidly.



The curves demonstrate how much moisture the air can hold for any temperature. Below 0 °C the curve is slightly different for an ice surface than for a supercooled water droplet.

Source: Briggs et al., *Fundamentals of the physical environment*, Routledge, 1997

Figure 2.25 Maximum vapour pressure

Factors affecting condensation

Condensation occurs when either **a** enough water vapour is evaporated into an air mass for it to become saturated or **b** when the temperature drops so that dew point (the temperature at which air is saturated) is reached. The first is relatively rare; the second common. Such cooling occurs in three main ways:

- radiation cooling of the air
- contact cooling of the air when it rests over a cold surface
- adiabatic (expansive) cooling of air when it rises.

Condensation is very difficult to achieve in pure air. It requires some tiny particle or nucleus onto which the vapour can condense. In the lower atmosphere these are quite common, such as sea salt, dust and pollution particles. Some of these particles are hygroscopic – that is, water-seeking – and condensation may occur when the relative humidity is as low as 80 per cent.

Other processes

Freezing refers to the change of liquid water into a solid, namely ice, once the temperature falls below 0 °C. Melting is the change from a solid to a liquid when the air temperature rises above 0 °C. Sublimation is the conversion of a solid into a vapour with no intermediate liquid state. Under conditions of low humidity, snow can be evaporated directly into water vapour without entering the liquid state. Sublimation is also used to describe the direct deposition of water vapour onto ice. In some cases, water droplets may be deposited directly onto natural features (such as plants and animals) as well as built structures (for example buildings and vehicles).

Precipitation

The term 'precipitation' refers to all forms of deposition of moisture from the atmosphere in either solid or liquid states. It includes rain, hail, snow and dew. Because rain is the most common form of precipitation in many areas, the term is sometimes applied to rainfall alone. For any type of precipitation, except dew, to form, clouds must first be produced.

When minute droplets of water are condensed from water vapour, they float in the atmosphere as clouds. If droplets coalesce, they form large droplets that, when heavy enough to overcome by gravity an ascending current, they fall as rain. Therefore cloud droplets must get much larger to form rain. There are a number of theories to suggest how raindrops are formed.

The Bergeron theory suggests that for rain to form, water and ice must exist in clouds at temperatures below 0 °C. Indeed, the temperature in clouds may be as low as –40 °C. At such temperatures, water droplets and ice droplets form. Ice crystals grow by condensation and become big enough to overcome turbulence and cloud updrafts, so they fall. As they fall, crystals coalesce to form larger snowflakes. These generally melt and become rain as they pass into the warm air layers near the ground. Thus, according to Bergeron, rain comes from clouds that are well below freezing at high altitudes, where the coexistence of water and ice is possible. The snow/ice melts as it passes into clouds at low altitude where the temperatures are above freezing level.

Other mechanisms must also exist as rain also comes from clouds that are not so cold. Mechanisms include:

- condensation on extra-large hygroscopic nuclei
- coalescence by sweeping, whereby a falling droplet sweeps up others in its path
- the growth of droplets by electrical attraction.

Causes of precipitation

The Bergeron theory relates mostly to snow-making. Snow is a single flake of frozen water. Rain and drizzle are found when the temperature is above 0°C (drizzle has a diameter of < 0.5 mm). Sleet is partially melted snow.

There are three main types of rainfall: convectional, frontal (depressional) and orographic (relief) (Figure 2.26).

Convectional rainfall

When the land becomes very hot, it heats the air above it. This air expands and rises. As it rises, cooling and condensation take place. If it continues to rise, rain will fall. It is very common in tropical areas (Figure 2.27) and is associated with the permanence of the ITCZ. In temperate areas, convectional rain is more common in summer.

Frontal or cyclonic rainfall

Frontal rain occurs when warm air meets cold air. The warm air, being lighter and less dense, is forced to rise over the cold, denser air. As it rises, it cools, condenses and forms rain. It is most common in middle and high latitudes where warm tropical air and cold polar air converge.

Orographic (or relief) rainfall

Air may be forced to rise over a barrier such as a mountain. As it rises, it cools, condenses and forms rain. There is often a rainshadow effect, whereby the leeward slope receives a relatively small amount of rain. Altitude is important, especially on a local scale. In general, there are increases of precipitation up to about 2 kilometres. Above this level, rainfall decreases because the air temperature is so low.

Thunderstorms (intense convectional rainfall)

Thunderstorms are special cases of rapid cloud formation and heavy precipitation in unstable air conditions. Absolute or conditional instability exists to great heights, causing strong updraughts to develop within cumulonimbus clouds. Air continues to rise as long as it is saturated (relative humidity is 100 per cent; that is, it has reached its dew point). Thunderstorms are especially common in tropical and warm areas where air can hold large amounts of water. They are rare in polar areas.

Several stages can be identified (Figure 2.28):

1. **Developing stage:** updraught caused by uplift; energy (latent heat) is released as condensation occurs; air becomes very unstable; rainfall occurs as cloud temperature is greater than 0 °C; the great strength of uplift prevents snow and ice from falling.
2. **Mature stage:** sudden onset of heavy rain and maybe thunder and lightning; rainfall drags cold air down with it; upper parts of the cloud may reach the tropopause; the cloud spreads, giving the characteristic anvil shape.
3. **Dissipating stage:** downdraughts prevent any further convective instability; the new cells may be initiated by the meeting of cold downdraughts from cells some distance apart, triggering the rise of warm air in between.

Lightning occurs to relieve the tension between different charged areas, for example between cloud and ground or within the cloud itself. The upper parts of the cloud are positive, whereas the lower parts are negative. The very base of the cloud is positively charged. The origin of the charges is not very clear, although they are thought to be

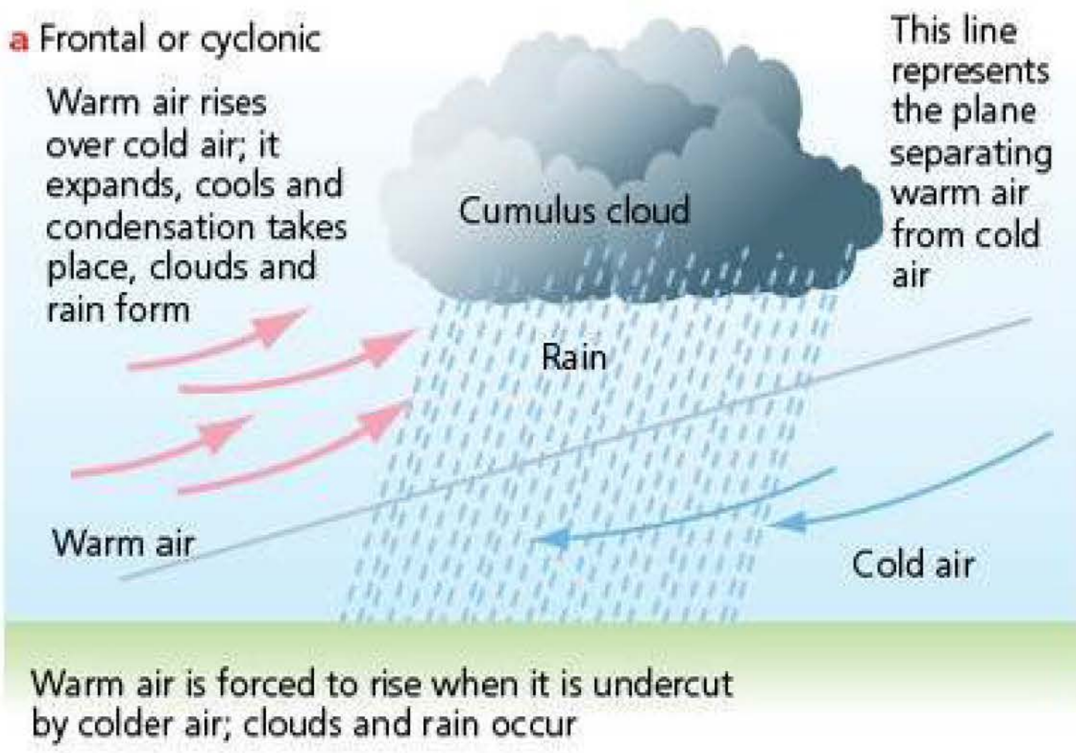
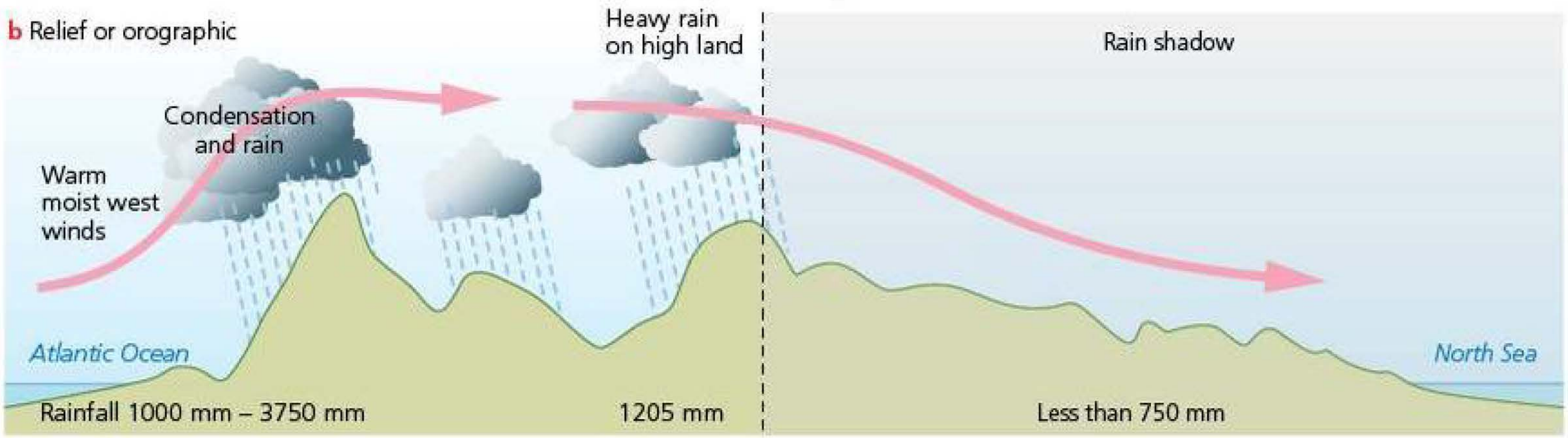
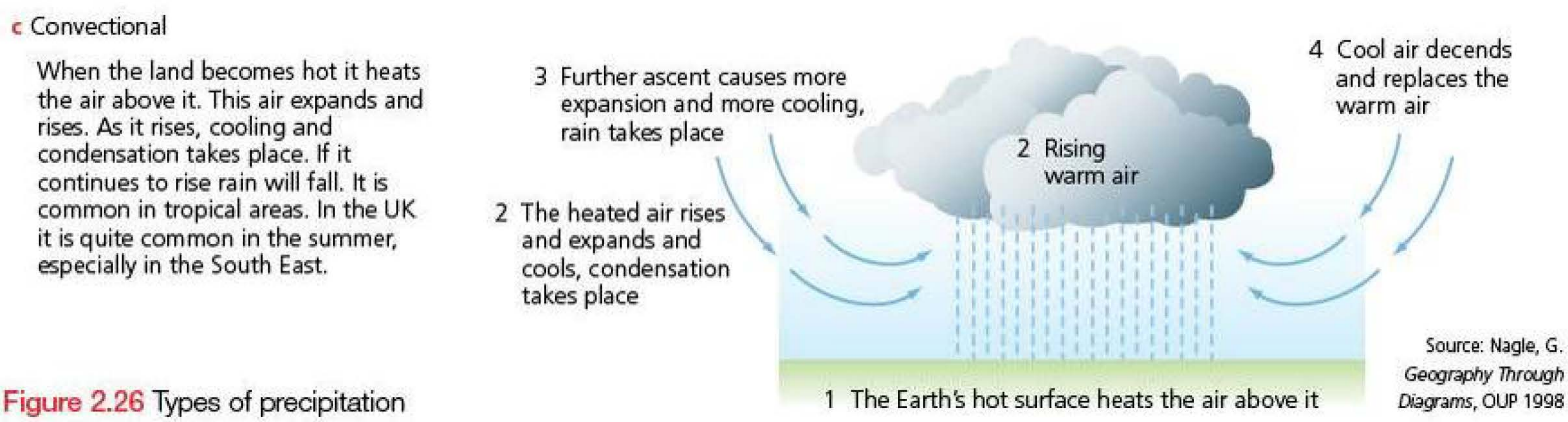


Figure 2.27 Convectional rain in Brunei





Source: Nagle, G. *Geography Through Diagrams*, OUP 1998

Figure 2.26 Types of precipitation

due to condensation and evaporation. Lightning heats the air to very high temperatures. Rapid expansion and vibration of the column of air produces thunder.

Section 2.3 Activities

Using diagrams, explain the meaning of the terms **a** *convectional rainfall*, **b** *orographic rainfall*, **c** *frontal rainfall*.

Clouds

Clouds are formed of millions of tiny water droplets held in suspension. They are classified in a number of ways, the most important being:

- **form or shape**, such as stratiform (layers) and cumuliform (heaped type)
- **height**, such as low (<2000m), medium or alto (2000–7000m) and high (7000–13000m).

There are a number of different types of clouds (Figure 2.29). High clouds consist mostly of ice crystals. Cirrus are wispy clouds, and include cirrocumulus (mackerel sky) and cirrostratus (halo effect around the Sun or Moon). Alto or middle-height clouds generally consist of water drops. They exist at temperatures lower than 0°C. Low clouds indicate poor weather. Stratus clouds are dense, grey and low lying (Figure 2.30). Nimbostratus are those that produce rain ('nimbus' means 'storm'). Stratocumulus are long cloud rolls, and a mixture of stratus and cumulus (see Figure 2.6 in Section 2.1).

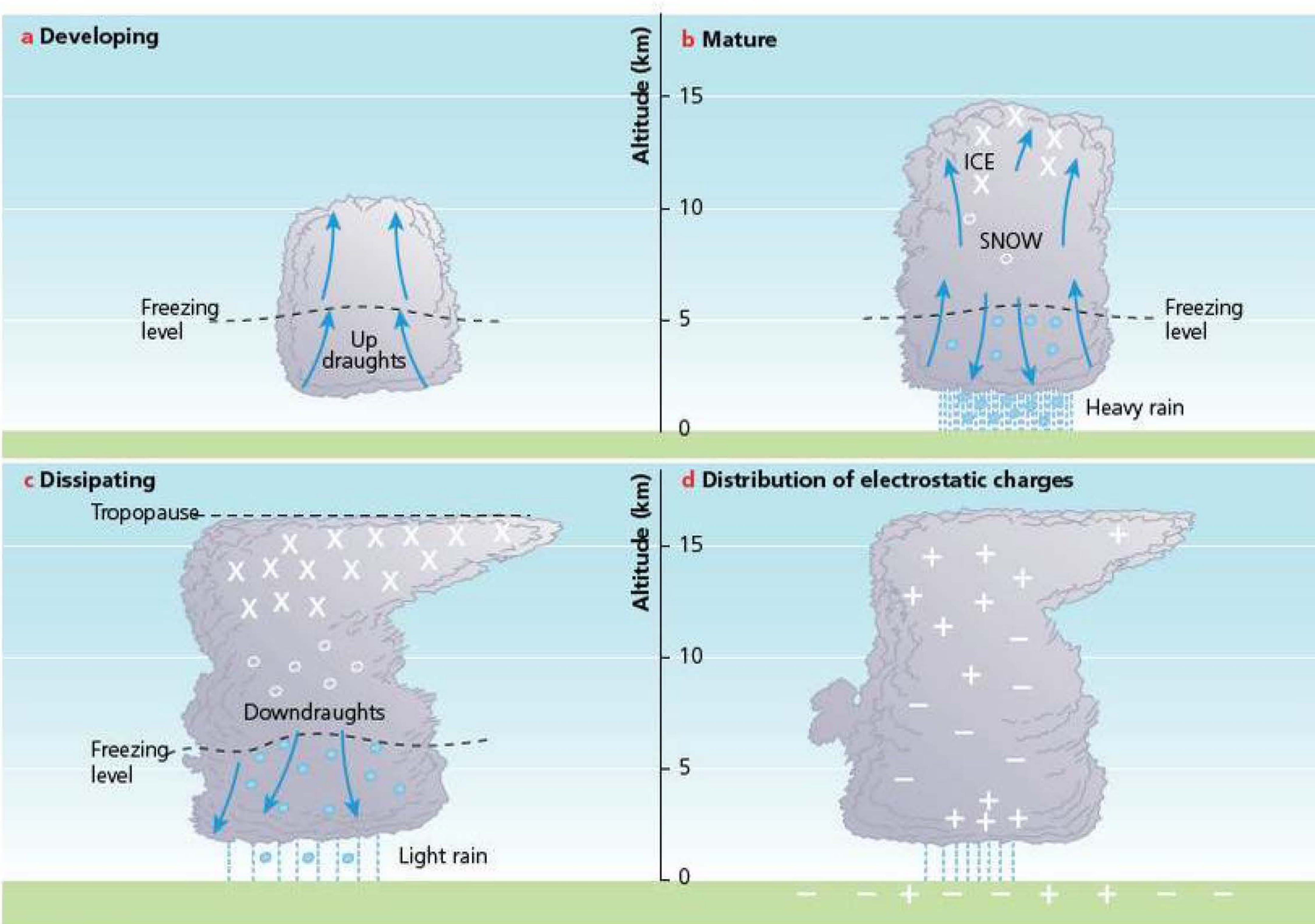


Figure 2.28 Stages in a thunderstorm

Vertical development suggests upward movement. Cumulus clouds are flat-bottomed and heaped. They indicate bright brisk weather. Cumulonimbus clouds produce heavy rainfall and often thunderstorms.

The important facts to keep in mind:

- In unstable conditions, the dominant form of uplift is convection and this may cause cumulus clouds.
- With stable conditions, stratiform clouds generally occur.
- Where fronts are involved, a variety of clouds exist.
- Relief or topography causes stratiform or cumuliform clouds, depending on the stability of the air.

Banner clouds

These are formed by orographic uplift (that is, air forced to rise, over a mountain for example) under stable air conditions. Uplifted moist air streams reach condensation only at the very summit, and form a small cloud. Further downwind the air sinks, and the cloud disappears. Wave clouds reflect the influence of the topography on the flow of air.

Types of precipitation

Rain

Rain refers to liquid drops of water with a diameter of between 0.5 millimetres and 5 millimetres. It is heavy enough to fall to the ground. Drizzle refers to rainfall with a diameter of less than 0.5 millimetres. Rainfall varies in terms of total amount, seasonality, intensity, duration and effectiveness; that is, whether there is more rainfall than potential evapotranspiration. (Refer back to page 45 for more information on the three main types of rainfall.)

Hail

Hail is alternate concentric rings of clear and opaque ice, formed by raindrops being carried up and down in vertical air currents in large cumulonimbus clouds. Freezing and partial melting may occur several times before the pellet is large enough to escape from the cloud. As the raindrops are carried high up in the cumulonimbus cloud they freeze. The hailstones may collide with droplets of supercooled water, which freeze on impact with and form a layer of opaque ice around the hailstone. As the hailstone falls, the outer layer may be melted but may freeze again with further uplift. The process can occur many times before the hail finally falls to ground, when its weight is great enough to overcome the strong updraughts of air.

Snow

Snow is frozen precipitation (Figure 2.31). Snow crystals form when the temperature is below freezing and water vapour is converted into a solid. However, very cold air contains a limited amount of moisture, so the heaviest snowfalls tend to occur when warm moist air is forced over very high mountains or when warm moist air comes into contact with very cold air at a front.

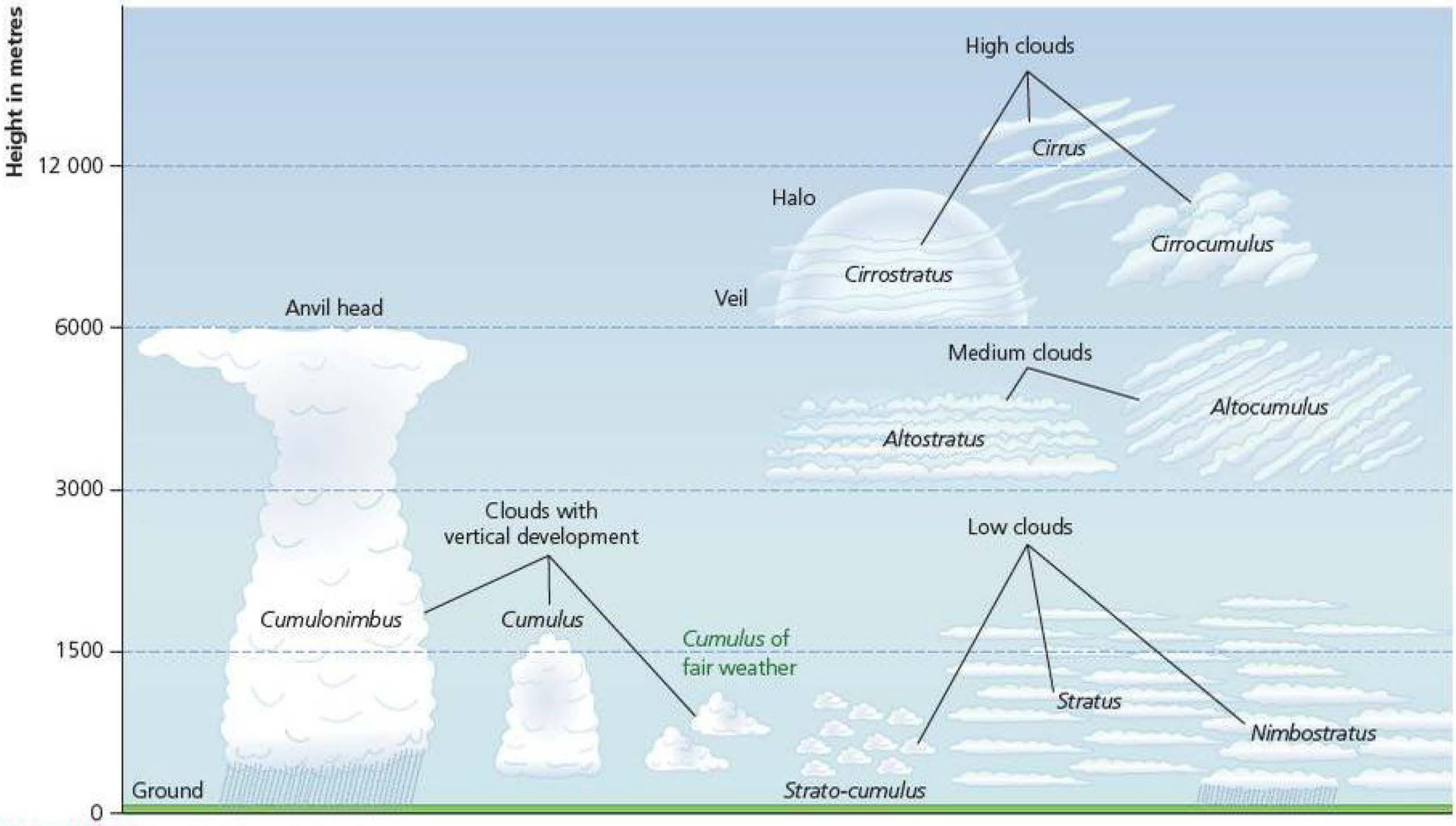


Figure 2.29 Classification of clouds

Figure 2.30 Stratus clouds

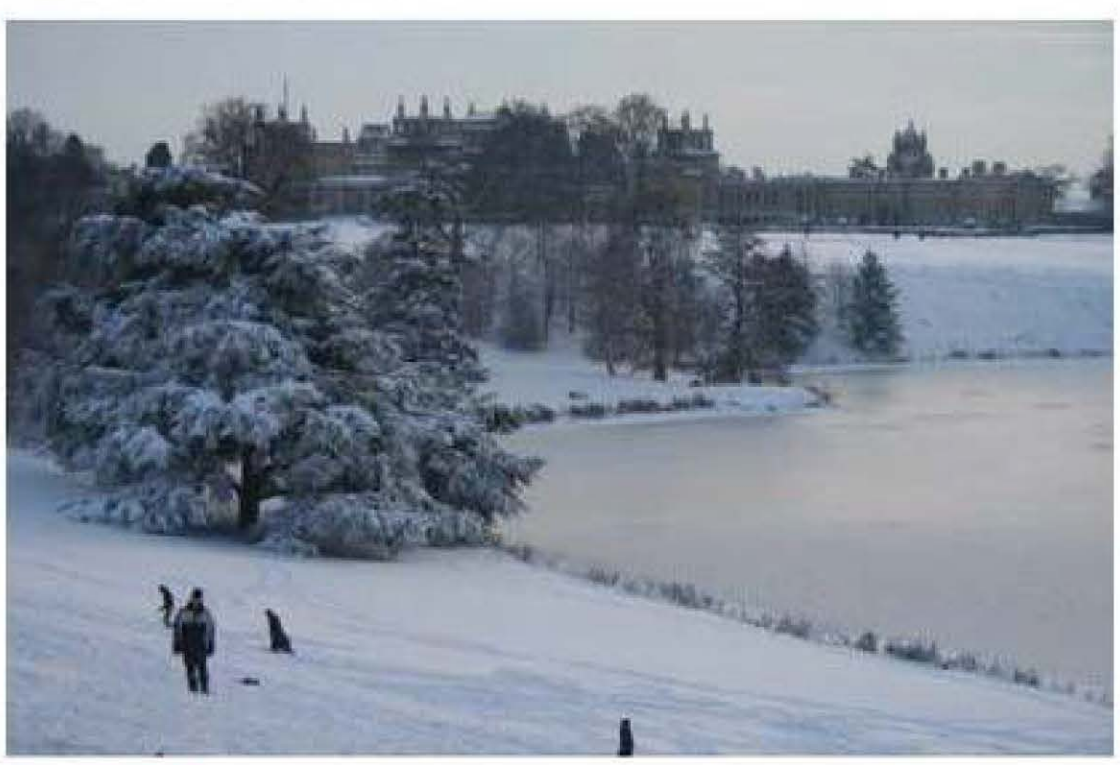


Figure 2.31 Snow at Blenheim Palace, Oxfordshire, UK

Dew

Dew is the direct deposition of water droplets onto a surface. It occurs in clear, calm anticyclonic conditions (high pressure) where there is rapid radiation cooling by night. The temperature reaches dew point, and further cooling causes condensation and direct precipitation onto the ground and vegetation (Figure 2.32).

Figure 2.32 Dew – direct condensation onto vegetation

Fog

Fog is cloud at ground level. Radiation fog (Figure 2.33) is formed in low-lying areas during calm weather, especially during spring and autumn. The surface of the ground, cooled rapidly at night by radiation, cools the air immediately above it. This air then flows into hollows by gravity and is cooled to dew point (the temperature at which condensation occurs). Ideal conditions include a surface layer of moist air and clear skies, which allow rapid radiation cooling.

Figure 2.33 Fog in the Wicklow Mountains, Ireland

The decrease in temperature of the lower layers of the air causes air to go below the dew point. With fairly light winds, the fog forms close to the water surface, but with stronger turbulence the condensed layer may be uplifted to form a low stratus sheet.

As the Sun rises, radiation fog disperses. Under cold anticyclonic conditions in late autumn and winter, fog may be thicker and more persistent, and around large towns smog may develop under an inversion layer. An inversion means that cold air is found at ground level, whereas warm air is above it – unlike the normal conditions in which air temperature declines with height. In industrial areas, emissions of sulphur dioxide act as condensation nuclei and allow fog to form. Along motorways, the heavy concentration of vehicle emissions does the same. By contrast, in coastal areas the higher minimum temperatures means that condensation during high-pressure conditions is less likely.

Fog commonly occurs over the sea in autumn and spring because the contrast in temperature between land and sea is significant. Warm air from over the sea is cooled when it moves on land during anticyclonic conditions. In summer, the sea is cooler than the land so air is not cooled when it blows onto the land. By contrast, in winter there are more low-pressure systems, causing stronger winds and mixing the air.

Fog is more common in anticyclonic conditions. Anticyclones are stable high-pressure systems characterised by clear skies and low wind speeds. Clear skies allow maximum cooling by night. Air is rapidly cooled to dew point, condensation occurs and fog is formed.

Advection fog is formed when warm moist air flows horizontally over a cooler land or sea surface. Steam fog is very localised. Cold air blows over much warmer water. Evaporation from the water quickly saturates the air and the resulting condensation leads to steaming. It occurs when very cold polar air meets the surrounding relatively warm water.

Section 2.3 Activities

1. Distinguish between *radiation fog* and *advection fog*.
2. Under which atmospheric conditions (stability or instability) do mist and fog form? Briefly explain how fog is formed.
3. Under which atmospheric conditions do thunder and lightning form? Briefly explain how thunder is created.

2.4 The human impact

Global warming

The role of greenhouse gases

Greenhouse gases are essential for life on Earth. The Moon is an airless planet that is almost the same distance from the Sun as is the Earth. Average temperatures on the Moon are about −18 °C, compared with about 15 °C on Earth. The Earth's atmosphere therefore raises temperatures by about 33 °C. This is due to the greenhouse gases, such as water vapour, carbon dioxide, methane, ozone, nitrous oxides and chlorofluorocarbons (CFCs). They are called greenhouse gases because, as in a greenhouse, they allow short-wave radiation from the Sun to pass through them, but they trap outgoing long-wave radiation, thereby raising the temperature of the lower atmosphere (Figure 2.34). The greenhouse effect is both natural and good – without it there would be no human life on Earth. On the other hand, there are concerns about the enhanced greenhouse effect.

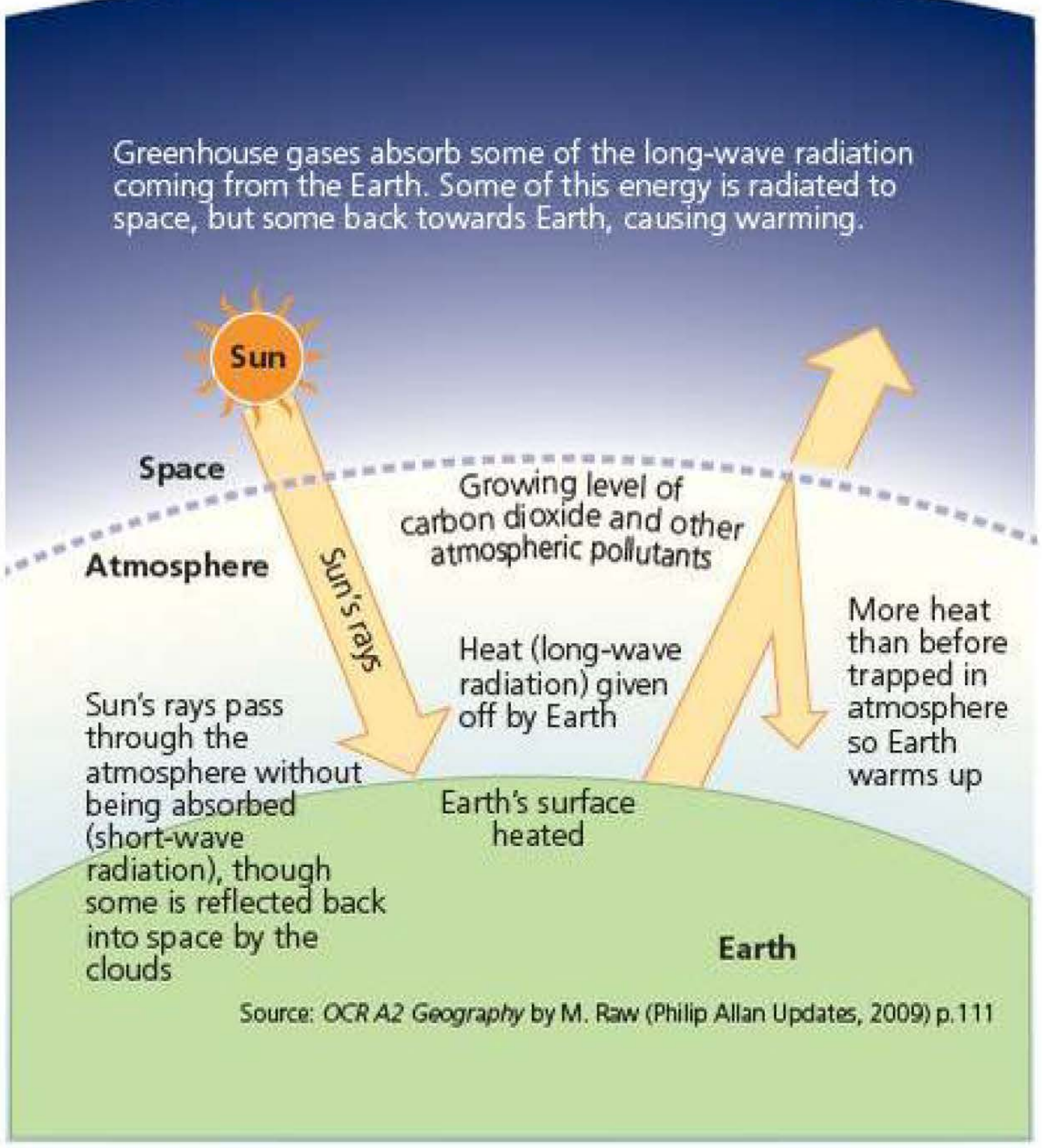


Figure 2.34 The greenhouse effect

The enhanced greenhouse effect is built up of certain greenhouse gases as a result of human activity (Table 2.2). **Carbon dioxide** (CO_2) levels have risen from about 315 ppm (parts per million) in 1950 to over 400 ppm and are expected to reach 600 ppm by 2050. The increase is due to human activities: burning fossil fuels (coal, oil and natural gas) and deforestation. Deforestation of the tropical rainforest is a double blow – not only does it increase atmospheric CO_2 levels, it also removes the trees that convert CO_2 into oxygen.

Methane is the second largest contributor to global warming, and is increasing at a rate of between 0.5 and 2 per cent per annum. Cattle alone give off between 65 and 85 million tonnes of methane per year. Natural wetland and paddy fields are another important source – paddy fields emit up to 150 million tonnes of methane annually. As global warming increases, bogs trapped in permafrost will melt and release vast quantities of methane. **Chlorofluorocarbons** (CFCs) are synthetic chemicals that destroy ozone, as well as absorb long-wave radiation. CFCs are increasing at a rate of 6 per cent per annum, and are up to 10 000 times more efficient at trapping heat than CO_2.

As long as the amount of water vapour and carbon dioxide stay the same and the amount of solar energy remains the same, the temperature of the Earth should remain in equilibrium. However, human activities are upsetting the natural balance by increasing the amount of CO_2 in the atmosphere, as well as the other greenhouse gases.

How human activities add to greenhouse gases

Much of the evidence for the greenhouse effect has been taken from ice cores dating back 160 000 years. These show that the Earth's temperature closely paralleled the levels of CO_2 and methane in the atmosphere. Calculations indicate that changes in these greenhouse gases were part, but not all, of the reason for the large (5°–7°) global temperature swings between ice ages and interglacial periods.

Accurate measurements of the levels of CO_2 in the atmosphere began in 1957 in Hawaii. The site chosen was far away from major sources of industrial pollution and shows a good representation of unpolluted atmosphere. The trend in CO_2 levels shows a clear annual pattern, associated with seasonal changes in vegetation, especially those over the northern hemisphere. By the 1970s there was a second trend, one of a long-term increase in CO_2 levels, superimposed upon the annual trends.

Studies of cores taken from ice packs in Antarctica and Greenland show that the level of CO_2 between 10 000 years ago and the mid-nineteenth century was stable, at about 270 ppm. By 1957, the concentration of CO_2 in the atmosphere was 315 ppm, and it has since risen to about 360 ppm. Most of the extra CO_2 has come

Table 2.2 Properties of key greenhouse gases

	Average atmospheric concentration (ppmv)	Rate of change (% per annum)	Direct global warming potential (GWP)	Lifetime (years)	Type of indirect effect
Carbon dioxide	400	0.5	1	120	None
Methane	1.72	0.6–0.75	11	10.5	Positive
Nitrous oxide	0.31	0.2–0.3	270	132	Uncertain
CFC-11	0.000255	4	3400	55	Negative
CFC-12	0.000453	4	7100	116	Negative
CO				Months	Positive
NOx					Uncertain

from the burning of fossil fuels, especially coal, although some of the increase may be due to the disruption of the rainforests. For every tonne of carbon burned, 4 tonnes of CO_2 are released.

By the early 1980s, 5 gigatonnes (5000 million tonnes, or 5 Gt) of fuel were burned every year. Roughly half the CO_2 produced is absorbed by natural sinks, such as vegetation and plankton.

Other factors have the potential to affect climate too. For example, a change in the albedo (reflectivity of the land brought about by desertification or deforestation) affects the amount of solar energy absorbed at the Earth's surface. Aerosols made from sulphur, emitted largely in fossil-fuel combustion, can modify clouds and may act to lower temperatures. Changes in ozone in the stratosphere due to CFCs may also influence climate.

Since the Industrial Revolution, the combustion of fossil fuels and deforestation have led to an increase of 26 per cent of CO_2 concentration in the atmosphere (Figure 2.35). Emissions of CFCs used as aerosol propellants, solvents, refrigerants and foam-blowing agents are also well known. They were not present in the atmosphere before their invention in the 1930s. The sources of methane and nitrous oxides are less well known. Methane concentrations have more than doubled because of rice production, cattle rearing, biomass burning, coal mining and ventilation of natural gas. Fossil fuel combustion may have also contributed through chemical reactions in the atmosphere, which reduce the rate of removal of methane. Nitrous oxide has increased by about 8 per cent since pre-industrial times, presumably due to human activities. The effect of ozone on climate is strongest in the upper troposphere and lower stratosphere.

- The increasing carbon dioxide in the atmosphere since the pre-industrial era, from about 280 to 382 ppmv (parts per million by volume), makes the largest individual contribution to greenhouse gas radiative forcing: 1.56 W/m² (watts per square metre).
- The increase of methane (CH_4) since pre-industrial times (from 0.7 to 1.7 ppmv) contributes about 0.5 W/m².
- The increase in nitrous oxide (NOx) since pre-industrial times, from about 275 to 310 ppbv3, contributes about 0.1 W/m².
- The observed concentrations of halocarbons, including CFCs, have resulted in direct radiative forcing of about 0.3 W/m².

Figure 2.35 Changes in greenhouse gases since pre-industrial times

Arguments surrounding global warming

There are many causes of global warming and climate change. Natural causes include:

- variations in the Earth's orbit around the Sun
- variations in the tilt of the Earth's axis
- changes in the aspect of the poles from towards the Sun to away from it
- variations in solar output (sunspot activity)
- changes in the amount of dust in the atmosphere (partly due to volcanic activity)
- changes in the Earth's ocean currents as a result of continental drift.

All of these have helped cause climate change, and may still be doing so, despite anthropogenic forces.

Complexity of the problem

Climate change is a very complex issue for a number of reasons:

- Scale – it includes the atmosphere, oceans and land masses across the world.
- Interactions between these three areas are complex.
- It includes natural as well as anthropogenic forces.
- There are feedback mechanisms involved, not all of which are fully understood.
- Many of the processes are long term and so the impact of changes may not yet have occurred.

The effects of increased global temperature change

The effects of global warming are varied (see Table 2.3). Much depends on the scale of the changes. For example, some impacts could include:

- a rise in sea levels, causing flooding in low-lying areas such as the Netherlands, Egypt and Bangladesh – up to 200 million people could be displaced
- 200 million people at risk of being driven from their homes by flood or drought by 2050
- 4 million km² of land, home to one-twentieth of the world's population, threatened by floods from melting glaciers
- an increase in storm activity, such as more frequent and intense hurricanes (owing to more atmospheric energy)
- changes in agricultural patterns, for example a decline in the USA's grain belt, but an increase in Canada's growing season
- reduced rainfall over the USA, southern Europe and the Commonwealth of Independent States (CIS), leading to widespread drought (Figure 2.36)
- 4 billion people could suffer from water shortages if temperatures rise by 2 °C
- a 35 per cent drop in crop yields across Africa and the Middle East expected if temperatures rise by 3 °C
- 200 million more people could be exposed to hunger if world temperatures rise by 2 °C; 550 million if temperatures rise by 3 °C
- 60 million more Africans could be exposed to malaria if world temperatures rise by 2 °C
- extinction of up to 40 per cent of species of wildlife if temperatures rise by 2 °C.

Table 2.3 Some potential effects of a changing climate in the UK

Positive effects	Negative effects
■ An increase in timber yields (up to 25% by 2050), especially in the north (with perhaps some decrease in the south). ■ A northward shift of farming zones by about 200–300 km per 1°C of warming, or 50–80 km per decade, will improve some forms of agriculture, especially pastoral farming in the north-west. ■ Enhanced potential for tourism and recreation as a result of increased temperatures and reduced precipitation in the summer, especially in the south.	■ Increased damage effects of increased storminess, flooding and erosion on natural and human resources and human resource assets in coastal areas. ■ An increase in animal species, especially insects, as a result of northward migration from the continent and a small decrease in the number of plant species due to the loss of northern and montane (mountain types). ■ An increase in soil drought, soil erosion and the shrinkage of clay soils.

The Stern Review

The Stern Review (2006) was a report by Sir Nicholas Stern that analysed the financial implications of climate change. The report has a simple message:

- Climate change is fundamentally altering the planet.
- The risks of inaction are high.
- Time is running out.

The effects of climate change vary with the degree of temperature change (Figure 2.37). The report states that climate change poses a threat to the world economy and it will be cheaper to address the problem than to deal with the consequences. The global-warming argument seemed a straight fight between the scientific case to act, and the economic case not to. Now, economists are urging action.

The Stern Review says that doing nothing about climate change – the business-as-usual (BAU) approach – would lead to a reduction in global per person consumption of at least 5 per cent now and for ever. According to the report, global warming could deliver an economic blow of between 5 and 20 per cent of GDP to world economies because of natural disasters and the creation of hundreds of millions of climate refugees displaced by sea-level rise. Dealing with the problem, by comparison, will cost just 1 per cent of GDP, equivalent to £184 billion.

Main points

- Carbon emissions have already increased global temperatures by more than 0.5°C.
- With no action to cut greenhouse gases, we will warm the planet by another 2–3°C within 50 years.
- Temperature rise will transform the physical geography of the planet and the way we live.
- Floods, disease, storms and water shortages will become more frequent.
- The poorest countries will suffer the earliest and the most.
- The effects of climate change could cost the world between 5 and 20 per cent of GDP.
- Action to reduce greenhouse-gas emissions and the worst of global warming would cost 1 per cent of GDP.
- With no action, each tonne of carbon dioxide we emit will cause at least $85 (£45) of damage.
- Levels of carbon dioxide in the atmosphere should be limited to the equivalent of 450–550 ppm.
- Action should include carbon pricing, new technology and robust international agreements.

Figure 2.36 The effects of global warming

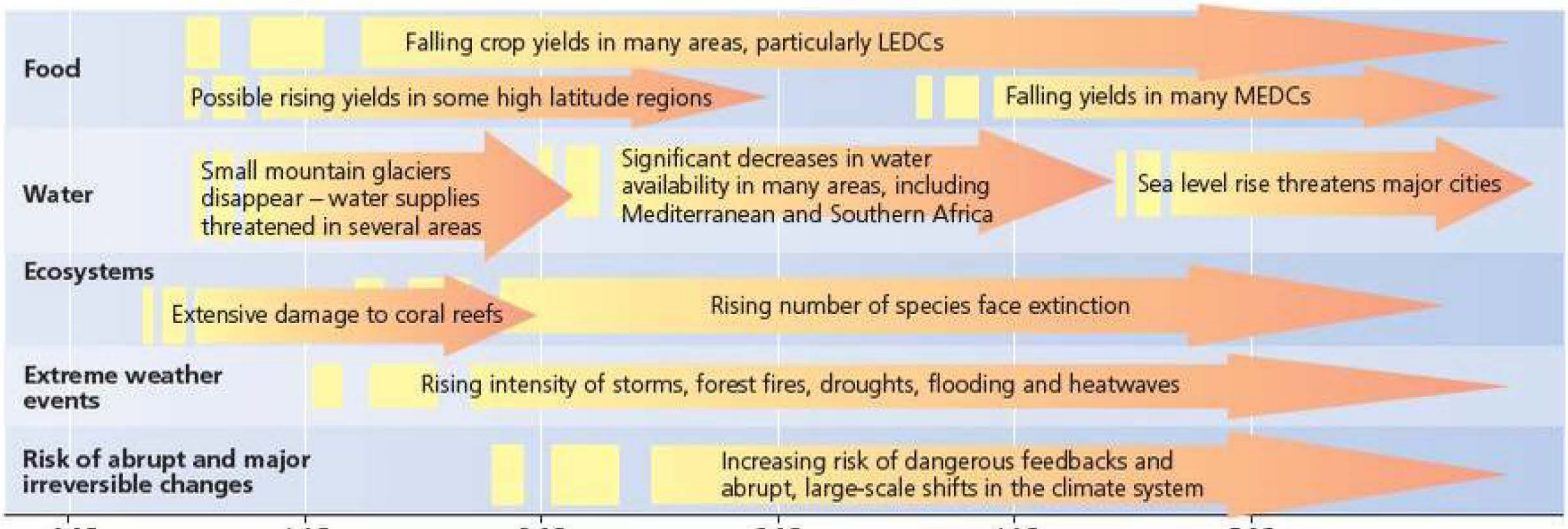


Figure 2.37 Projected impacts of climate change, according to the Stern Review

International policy to protect climate

The first world conference on climate change was held in Geneva in 1979. The Toronto Conference of 1988 called for the reduction of carbon dioxide emissions by 20 per cent of the 1988 levels by 2005. Also in 1988, UNEP and the World Meteorological Organization established the Intergovernmental Panel on Climate Change (IPCC).

> 'The ultimate objective is to achieve ... stabilisation of greenhouse gas concentrations in the atmosphere at a level that would prevent dangerous anthropogenic interference with the climate system.'

The Kyoto Protocol (1997) gave all high-income countries (HICs) legally binding targets for cuts in emissions from the 1990 level by 2008–12. The EU agreed to cut emissions by 8 per cent, Japan 7 per cent and the USA by 6 per cent. The Paris round of global climate-change talks (2015) attempted to bring all countries in line with plans to reduce climate change. However, many of the plans discussed had a very long time frame and there appeared to be little hope for a quick solution.

Section 2.4 Activities

1 Figure 2.37 shows some of the projected impacts related to global warming.
   a Describe the potential changes as a result of a 3 °C rise in temperature.
   b Explain why there is an increased risk of hazards in coastal cities.
   c Outline the ways in which it is possible to manage the impacts of global warming.
   d Evaluate the potential impacts of global warming.
2 Figure 2.38 shows variations in mean air temperature between 1880 and 2000.
   a i Identify the reason that the temperature in the early 1960s fell below 15 °C.
     ii Describe the impact of Pinatubo on global climate in the 1990s.
   b Outline the natural sources of greenhouse gases.
   c Using an annotated diagram, explain what is meant by the term *the greenhouse effect*.
   d Outline the benefits of the greenhouse effect.

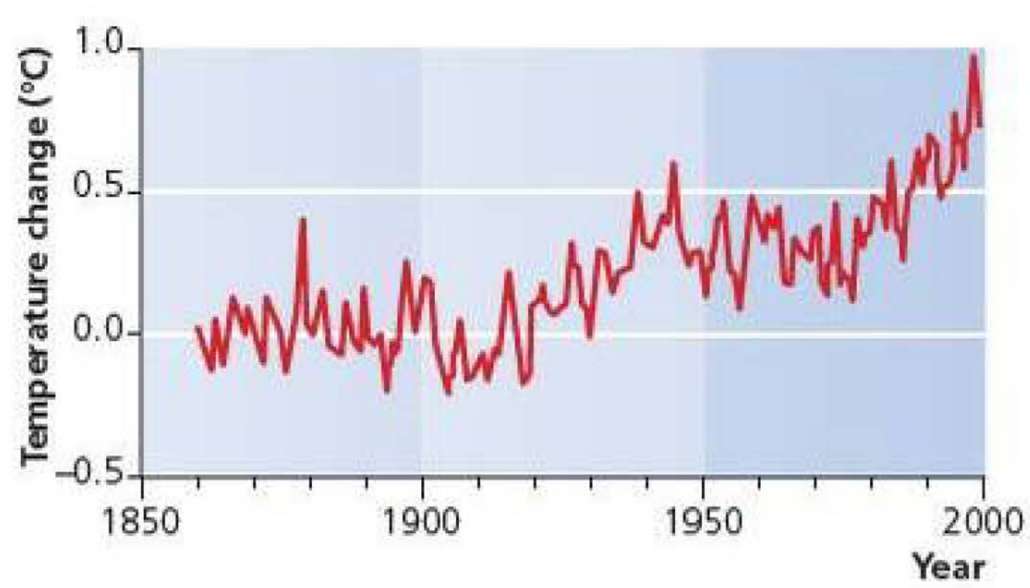


Figure 2.38 Variations in mean air temperature, 1880–2000

Urban climates

Urban climates occur as a result of extra sources of heat released from industry; commercial and residential buildings; as well as from vehicles, concrete, glass, bricks, tarmac – all of these act very differently from soil and vegetation. For example, the albedo (reflectivity) of tarmac is about 5–10 per cent, while that of concrete is 17–27 per cent. In contrast, that of grass is 20–30 per cent.

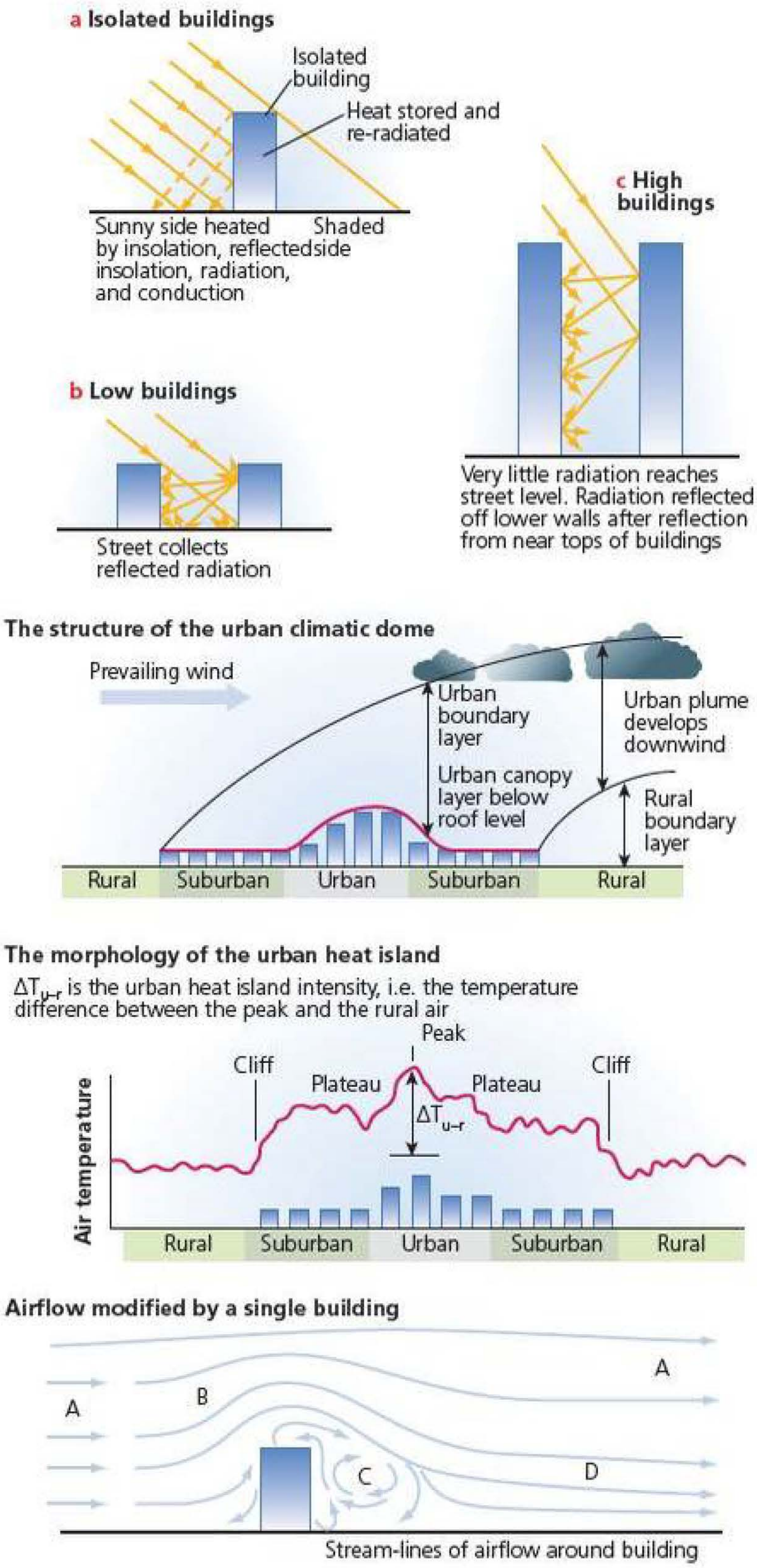


Figure 2.39 Processes in the urban heat island

Some of these – notably dark bricks – absorb large quantities of heat and release them slowly by night (Figure 2.39). In addition, the release of **pollutants** helps trap radiation in urban areas. Consequently, urban microclimates can be very different from rural ones. Greater amounts of dust mean an increasing concentration of hygroscopic particles. There is less water vapour, but more carbon dioxide and higher proportions of noxious fumes owing to combustion of imported fuels. Discharge of waste gases by industry is also increased.

Urban heat budgets differ from rural ones. By day, the major source of heat is solar energy; and in urban areas brick, concrete and stone have high heat capacities. A kilometre of an urban area contains a greater surface area than a kilometre of countryside, and the greater number of surfaces in urban areas allow a greater area to be heated. There are more heat-retaining materials with lower albedo and better radiation-absorbing properties in urban areas than in rural ones.

Moisture and humidity

In urban areas, there is relative lack of moisture. This is due to:

- a lack of vegetation
- a high drainage density (sewers and drains), which removes water.

Thus there are decreases in relative humidity in inner cities due to the lack of available moisture and higher temperatures there. However, this is partly countered in very cold, stable conditions by early onset of condensation in low-lying districts and industrial zones.

Nevertheless, there are more intense storms, particularly during hot summer evenings and nights, owing to greater **instability** and stronger convection above built-up areas. There is a higher incidence of thunder (due to more heating and instability) but less snowfall (due to higher temperatures), and any snow that does fall tends to melt rapidly.

Hence little energy is used for evapotranspiration, so more is available to heat the atmosphere. This is in addition to the sources of heating produced by people, such as in industry and by cars.

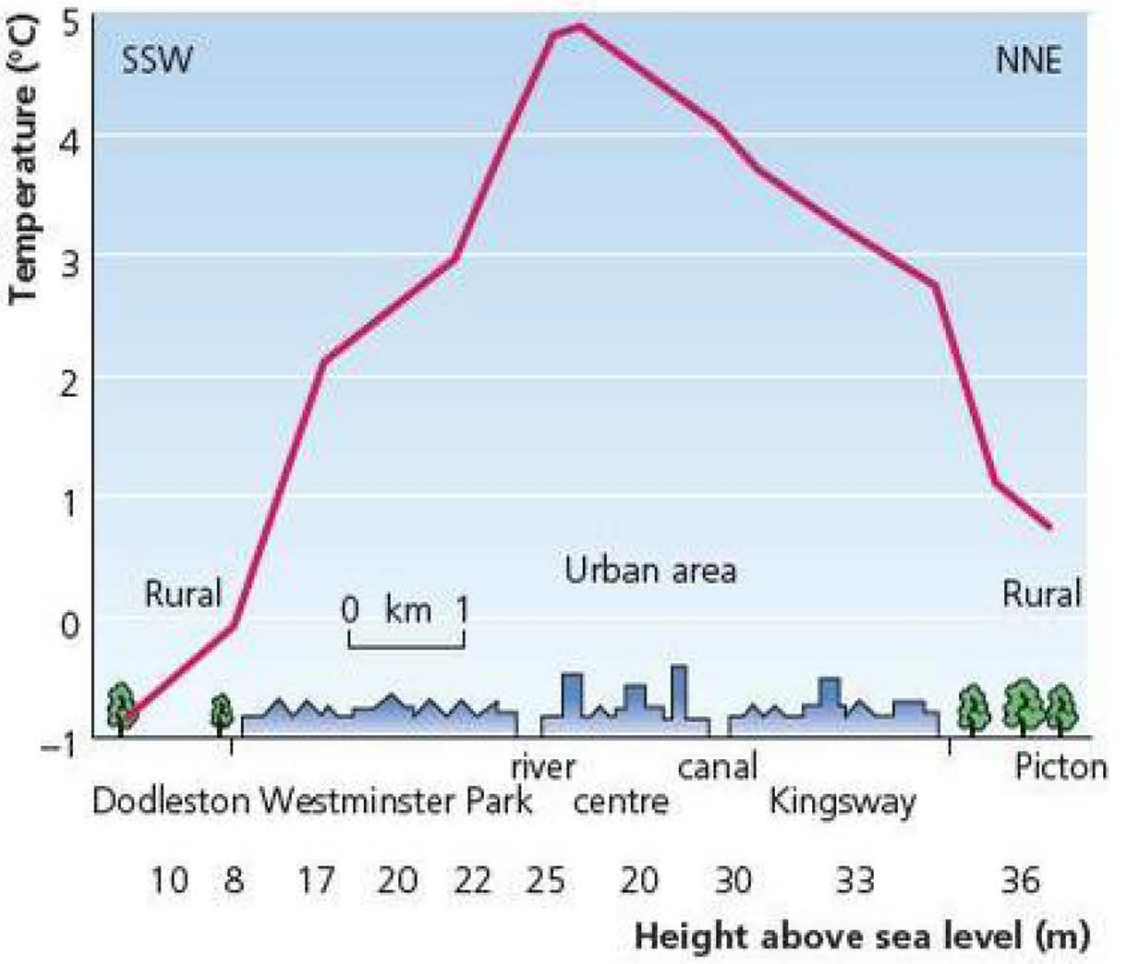
At night, the ground radiates heat and cools. In urban areas, the release of heat by buildings offsets the cooling process, and some industries, commercial activities and transport networks continue to release heat throughout the night.

There is greater scattering of shorter-wave radiation by dust, but much higher absorption of longer waves owing to the surfaces and to carbon dioxide. Hence there is more diffuse radiation, with considerable local contrasts owing to variable screening by tall buildings in shaded narrow streets. There is reduced visibility arising from industrial haze.

There is a higher incidence of thicker cloud cover in summer because of increased convection, and radiation fogs or smogs in winter because of air pollution. The concentration of hygroscopic particles accelerates the onset of condensation. Daytime temperatures are, on average, 0.6°C higher.

Urban heat island effect

The contrast between urban and rural areas is greatest under calm high-pressure conditions. The typical heat profile of an urban **heat island** shows a maximum at the city centre, a plateau across the suburbs and a temperature cliff between the suburban and rural areas (Figure 2.40). Small-scale variations within the urban heat island occur with the distribution of industries, open spaces, rivers, canals, and so on.

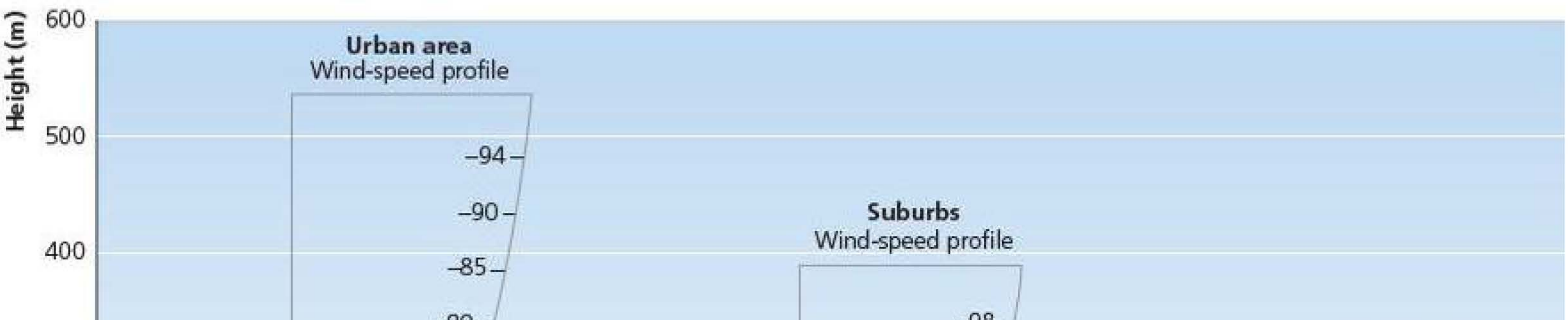


Source: Briggs, D. et al., *Fundamentals of the Physical Environment*, Routledge, 1997

Figure 2.40 The urban heat island (Chester, UK)

The heat island is a feature that is delimited by isotherms (lines of equal temperature), normally in an urban area. This shows that the urban area is warmer than the surrounding rural area, especially by dawn during anticyclonic conditions (Figure 2.41). The heat-island effect is caused by a number of factors:

- heat produced by human activity – a low level of radiant heat can be up to 50per cent of incoming energy in winter
- changes of energy balance – buildings have a high thermal capacity in comparison to rural areas; up to six times greater than agricultural land
- the effect on airflow – turbulence of air may be reduced overall, although buildings may cause funnelling effects
- there are fewer bodies of open water, so less evaporation and fewer plants, therefore less transpiration
- the composition of the atmosphere – the blanketing effect of smog, smoke or haze
- reduction in thermal energy required for evaporation and evapotranspiration due to the surface character, rapid drainage and generally lower wind speeds
- reduction of heat diffusion due to changes in airflow patterns as a result of urban surface roughness.



Source: Briggs, D. et al., *Fundamentals of the Physical Environment*, Routledge, 1997

Figure 2.41 The effect of terrain roughness on wind speed – with decreasing roughness, the depth of the affected layer becomes shallower and the profile steeper (numbers refer to wind strength as a percentage of maximum air speed)

Air flow

Urban areas may also develop a pollution dome. Highest temperatures are generally found over the city centre – or downwind of the city centre if there is a breeze present. Pollutants may be trapped under the dome. Cooler air above the dome prevents the pollutants from dispersing. These pollutants may prevent some incoming radiation from passing through, thereby reducing the impact of the heat island. By night, the pollutants may trap some long-wave radiation from escaping, thereby keeping urban areas warmer than surrounding rural areas.

Airflow over an urban area is disrupted; winds are slow and deflected over buildings (Figure 2.41). Large buildings can produce eddying. Severe gusting and turbulence around tall buildings causes strong local pressure gradients from windward to leeward walls. Deep narrow streets are much calmer unless they are aligned with prevailing winds to funnel flows along them – the 'canyon effect'.

The nature of urban climates is changing (Table 2.4). With the decline in coal as a source of energy, there is less sulphur dioxide pollution and so fewer hygroscopic nuclei; there is therefore less fog. However, the increase in cloud cover has occurred for a number of reasons:

- greater heating of the air (rising air, hence condensation)
- increase in pollutants
- frictional and turbulent effects on airflow
- changes in moisture.

Table 2.4 Average changes in climate caused by urbanisation

Factor		Comparison with rural environments
Radiation	Global	2–10 % less
	Ultraviolet, winter	30 % less
	Ultraviolet, summer	5 % less
	Sunshine duration	5–15 % less
Temperature	Annual mean	1 °C more
	Sunshine days	2–6 °C more
	Greatest difference at night	11 °C more
	Winter maximum	1.5 °C more
	Frost-free season	2–3 weeks more
Wind speed	Annual mean	10–20 % less
	Gusts	10–20 % less
	Calms	5–20 % more
Relative humidity	Winter	2 % less
	Summer	8–10 % less
Precipitation	Total	5–30 % more
	Number of rain days	10 % more
	Snow days	14 % less
Cloudiness	Cover	5–10 % more
	Fog, winter	100 % more
	Fog, summer	30 % more
	Condensation nuclei	10 times more
	Gases	5–25 times more

Source: J. Tivy, Agricultural Ecology, *Longman 1990 p.372*

Section 2.4 Activities

1 Describe and account for the main differences in the climates of urban areas and their surrounding rural areas.
2 What is meant by the *urban heat island*?
3 Describe **one** effect that atmospheric pollution may have on urban climates.
4 Explain how buildings, tarmac and concrete can affect the climate in urban areas.
5 Why are microclimates, such as urban heat islands, best observed during high-pressure (anticyclonic) weather conditions?

Case Study: Urban microclimate – London

The heat island effect

Urban microclimates are perhaps the most complex of all microclimates. The general pattern in Figure 2.42 shows the highest temperatures in the city centre, reaching 10–11 °C, compared with the rural fringe temperature of 5 °C. Temperature falls more rapidly along the River Thames to the east of the City. The temperature gradient is more gentle in the west of the city, due to the density of urban infrastructure. Over steep temperature gradients there is a low density of urban infrastructure, for example the river and its vegetated banks. Where there is a gentle temperature gradient, there is a high density of urban infrastructure. Effectively from the map we can see that the east of London is less built up than the west. Temperature remains relatively constant for approximately 15 kilometres west of the city centre before rapidly falling within a 5–6 kilometre distance.

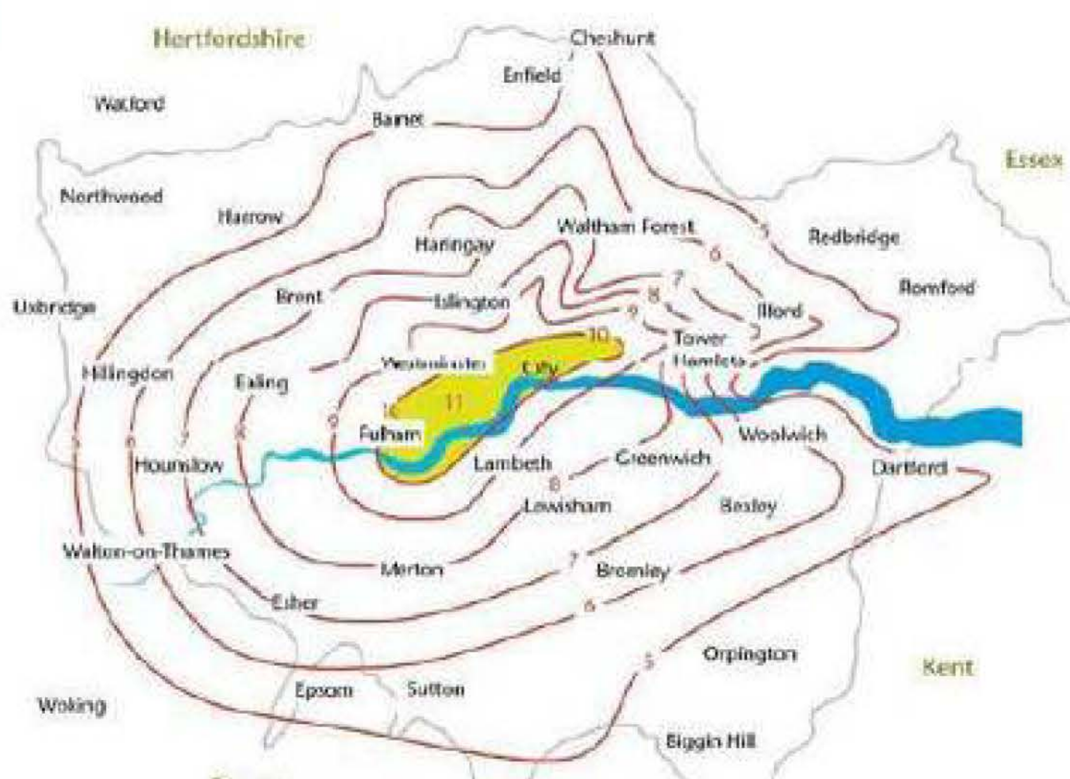
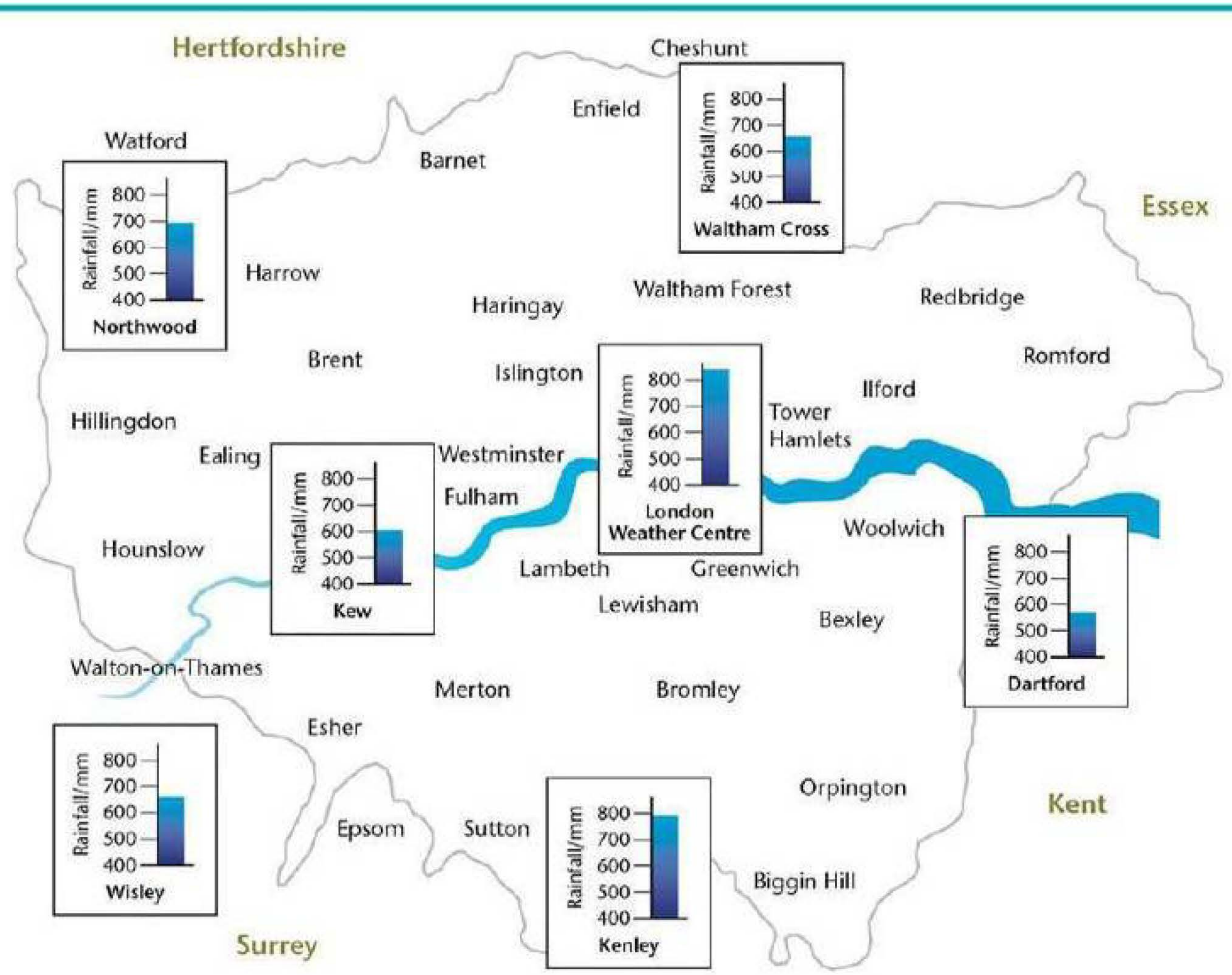


Figure 2.42 London's heat-island effect, showing minimum temperatures (°C) in mid-May



Source: National Meteorological Library and Archive Fact sheet 14 — Microclimates; Figure 16. Mean annual rainfall totals for a number of stations around London. www.metoffice.gov.uk/media/pdf/n/9/Fact_sheet_No._14.pdf

Figure 2.43 Variations in mean annual rainfall around London

Recent research on London's heat island has shown that the pollution domes can also filter incoming solar radiation, thereby reducing the build-up of heat during the day. At night, the dome may trap some of the heat accrued during the day, so these domes might be reducing the sharp differences between urban and rural areas.

There is an absence of strong winds both to disperse the heat and to bring in cooler air from rural and suburban areas. Indeed, urban heat islands are often most clearly defined on calm summer evenings, often under blocking anticyclones.

The distribution of rainfall is very much influenced by topography, with the largest values occurring over the more hilly regions, and lowest values in more low-lying areas. Figure 2.43 illustrates this point quite clearly. Kenley on the North Downs, at an altitude of 170 metres above mean sea-level, has an average annual rainfall of nearly 800 millimetres, whereas London Weather Centre, at 43 metres above mean sea-level, has an average annual rainfall of less than 550 millimetres. Overall, humidity is lower in London than surrounding areas, partly due to higher temperatures (warm air can hold more moisture, hence relative humidity may be lower), but water is removed from large urban areas due to the combination of drains and sewers, the large amount of impermeable surfaces and the reduced vegetation cover.

The urban heat island creates the urban boundary layer, which is a dome of rising warm air and low pressure. As ground surfaces are heated, rapid evapotranspiration takes place. This evapotranspiration, although lower compared to rural areas, occurs more rapidly and can result in cumulus cloud and convectional weather patterns. Due to the low pressure caused by rising air, surface winds are drawn in from the surrounding rural fringe. This air then converges as it is forced to rise over the high urban canopy. The urban boundary essentially creates an orographic process similar to a mountain barrier. The movement of winds contributes to increased rainfall patterns over the city that are most pronounced to the leeward side of the city core. However, as air passes over the urban boundary layer it begins to sink, leading to lower precipitation at the leeward rural area. These differences are also more pronounced in the summer compared to the winter.

Some studies have demonstrated a pattern of increased rain through the week and have shown Saturday rain to be a result of a build-up of pollutants due to five consecutive commutes. By Monday, pollutants have fallen and rainfall is less likely to form.

Case Study: Urban microclimate – Cheong Gye Cheon, Seoul, South Korea

The impact of river restoration on urban microclimates

In Seoul, capital of South Korea, there has been a very marked change in the urban microclimate following the removal of a large, downtown elevated motorway, and the restoration of a river and floodplain that had been built over. Since the restoration of the stream, air temperature has decreased by up to 10–13 per cent; that is, by 3-4 °C during the hottest days. Before the restoration, the area was showing a temperature about 5 °C higher than the average temperature of the city. The decrease in the number of vehicles passing by also contributed to the drop in the temperature. The heat island phenomenon used to be observed in the Cheong Gye Cheon Stream area under the impact of the heavy traffic, concentration of commercial facilities and the impermeable surface.

Following the completion of the restoration, the wind speed has become faster (by 2.2–7.1 per cent). The average wind speed measured at Cheong Gye Cheon is up to 7.8 per cent faster than before, apparently under the influence of the cool air forming along the stream.

Figure 2.44 Cheong Gye Cheon – **a** when the area was developed with an elevated highway and **b** after restoration

Case Study: Urban microclimate – Melbourne, Australia

With increasing distance from the city centre, the amount of tree cover in a suburb decreases, while the amount of green space, such as lawns and parks, increases. In Melbourne, for every 10 kilometres from the city centre, the tree cover drops by more than 2 per cent. That means Melbourne's inner suburbs might have more than 15 per cent cover, but an outer suburb could have less than 10 per cent. A 5 per cent fall in urban tree cover can lead to a 1–2 °C rise in air temperature. This matters for community health and well-being, especially for the vulnerable – the elderly, young children and those with existing health issues.

Trees are missing from back gardens – partly because modern houses in the outer suburbs take up more space, leaving less room for trees – and they are missing from the streets. The property boom led to a gradual thinning out of tree cover in established suburbs, as residential plots were sub-divided.

Melbourne aims to increase tree cover by 75 per cent before 2040, Sydney by 50 per cent before 2030 and Brisbane is targeting tree cover for cycleways and footpaths.

Microclimate mitigation

Increasingly, there are attempts to reverse urban microclimates. Heat-island mitigation strategies include urban forestry, living/green roofs and light surfaces.

In general, substantial reductions in surface and near-surface air temperature can be achieved by implementing heat-island mitigation strategies. Vegetation cools surfaces more effectively than increases in albedo, and curbside planting is the most effective mitigation strategy per unit area redeveloped. However, the greatest absolute temperature reductions are possible with light surfaces.

Table 2.5 Characteristics of the London Plane tree

Characteristic	The London Plane tree
Aesthetic value	A tall elegant tree providing pleasant shade in summer and a pleasing winter silhouette. Flaking bark creates attractive colours on trunk.
Does it make a mess?	Leaves, fruit and bark need clearing from streets and pavements.
Pollution tolerance	Very tolerant of air pollution. Hairs on young shoots and leaves help to trap particulate pollution.
Pests and diseases	Rarely affected by disease and pests (although some shoots are killed each year by fungal infection).
Soil conditions	Very tolerant of poor soil conditions, including compacted soil (although some stunting of growth is caused by road salt).
Space	Grows vigorously and is very tolerant of pruning.
Safety hazards	Trees rarely blow over or shed branches. Fine hairs on young shoots, leaves and fruit may cause irritation and even allergies in some people.
Microclimate	Open canopy produces light shade. Will intercept some rain, especially when in leaf.
Biodiversity	Provides valuable nesting sites for birds. Sufficient light below canopy to allow significant plant growth.

Source: Adapted from the Field Studies Council's Urban Ecosystems website www.field-studies-council.org/urbaneco

The London Plane tree – urban saviour

With an extensive and healthy urban forest, air quality can be drastically improved. Trees help to lower air temperatures through increasing evapotranspiration. This reduction of temperature not only lowers energy use, it also improves air quality, as the formation of ozone is dependent on temperature. Large shade trees can reduce local ambient temperatures by 3 to 5 °C. Maximum midday temperature reductions due to trees range from 0.04 °C to 0.2 °C per 1 per cent canopy cover increase.

Living roofs offer greater cooling per unit area than light surfaces, but less cooling per unit area than curbside planting.

Although street trees provide the greatest cooling potential per unit area, light surfaces provide the greatest overall cooling potential when available area is taken into account because there is more available area in which to implement this strategy compared to the other strategies.

Section 2.4 Activities

1 State the conditions in which London's heat island is most pronounced.
2 Briefly suggest reasons for variations in temperature as shown in Figure 2.42.
3 Describe and explain how the urban microclimate in Seoul changed after river restoration.
4 Explain the advantages of the London Plane tree for urban areas.