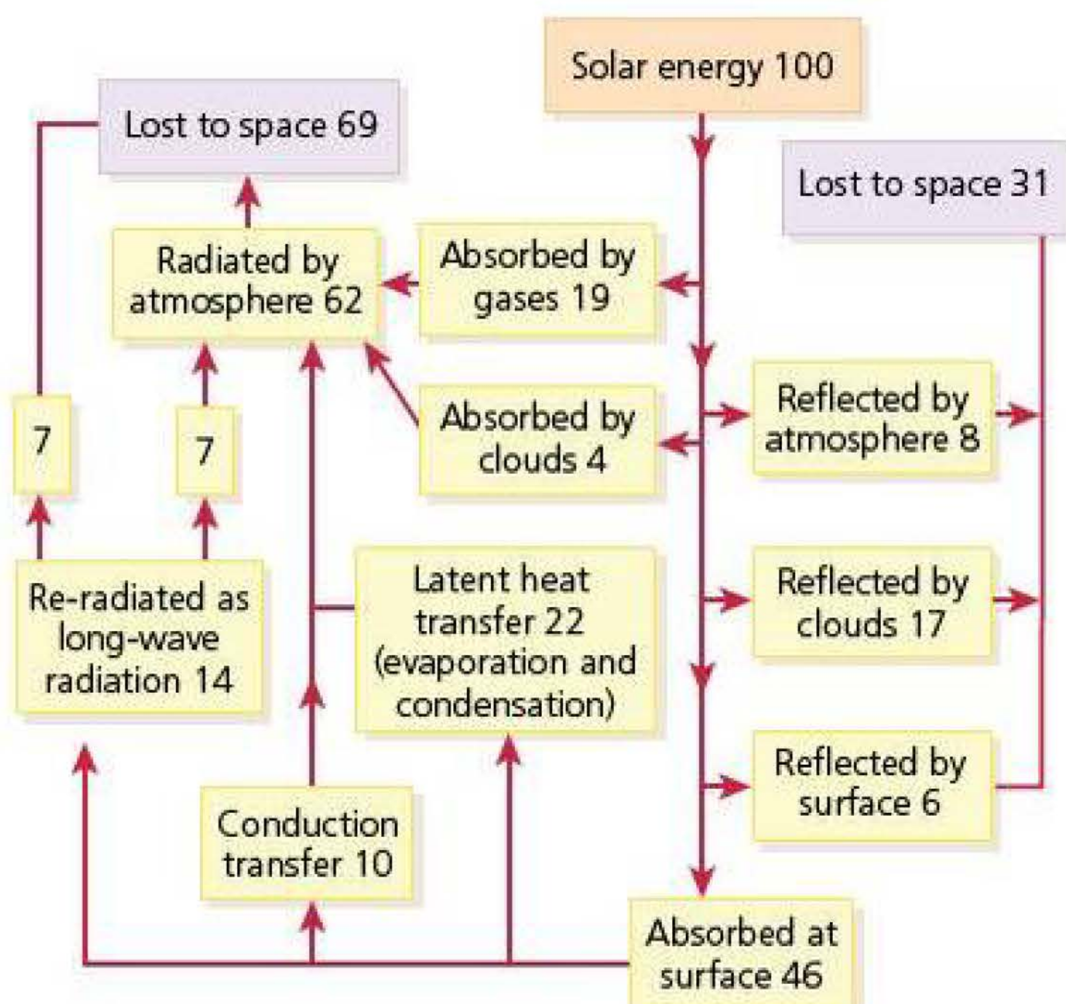
2.2 The global energy budget

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. **Incoming solar radiation** 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.

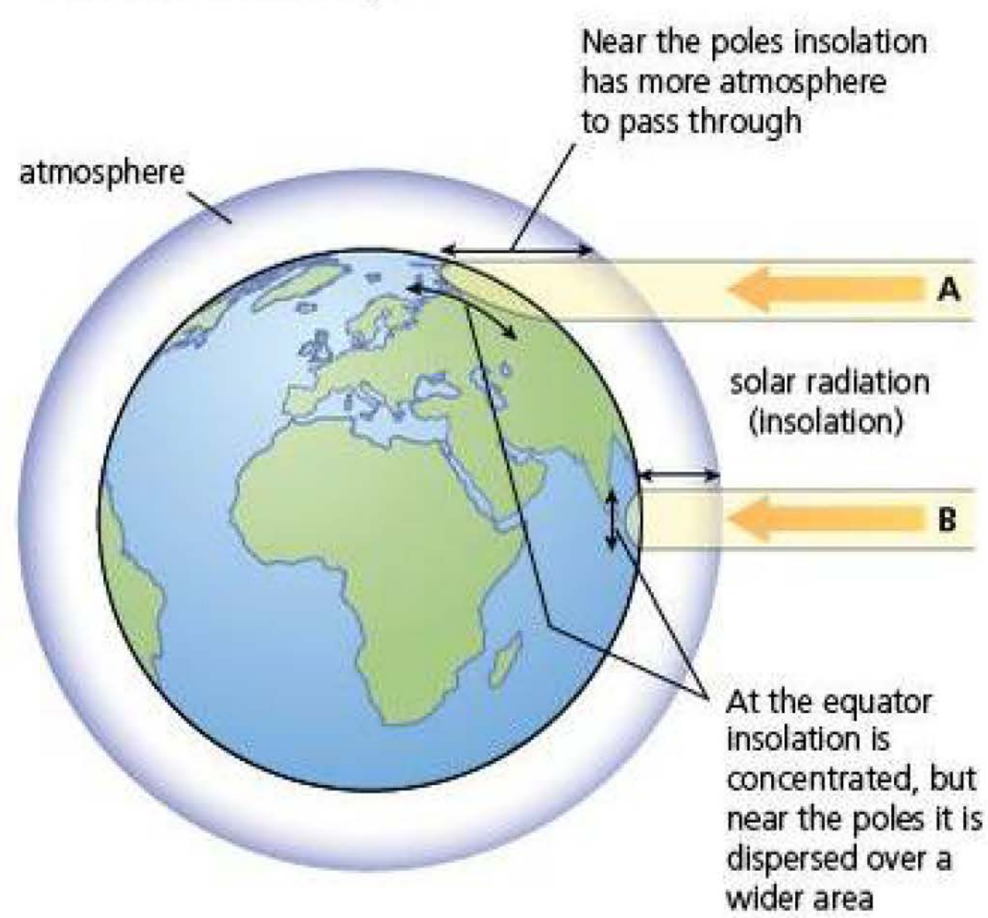
The atmosphere is largely heated from below. Most of the incoming short-wave radiation is let through, but some outgoing long-wave radiation is trapped by greenhouse gases. This is known as the greenhouse principle or greenhouse effect.

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

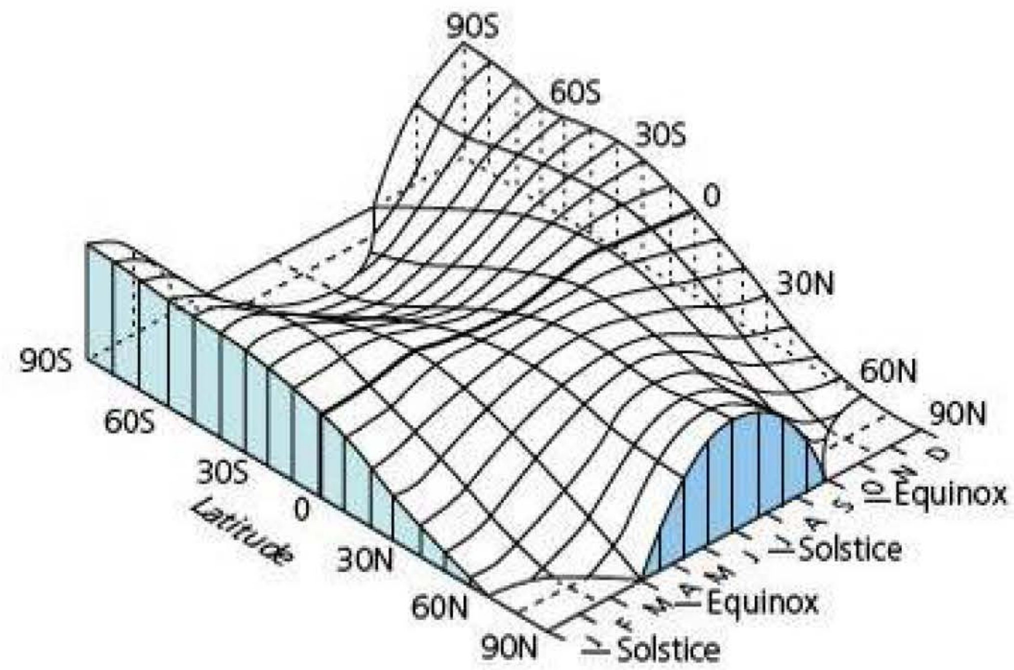
Areas that are close to the equator receive more heat than areas that are close to the poles. This is due to two reasons:

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

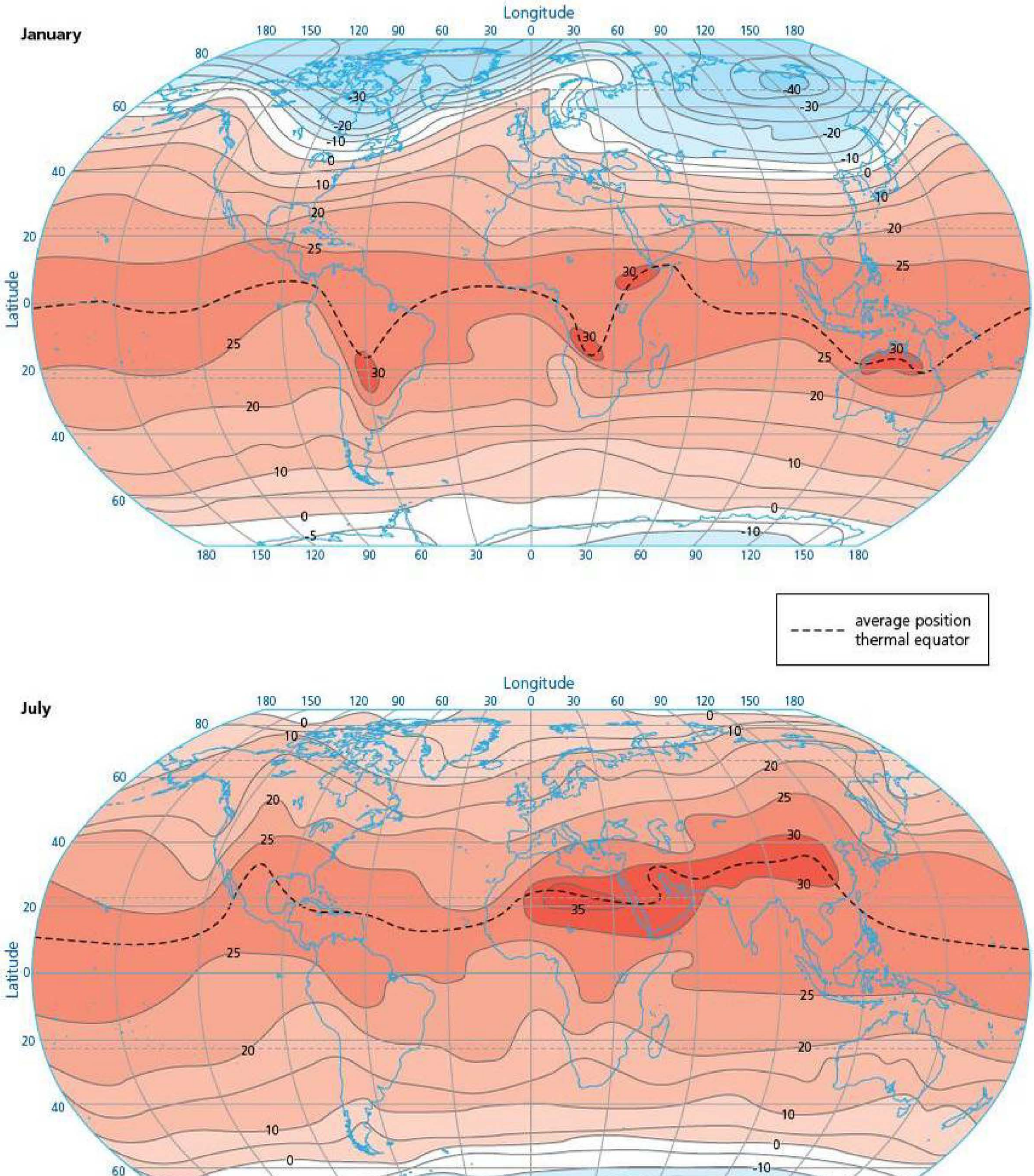
Pressure is measured in millibars (mb) and is represented by isobars, which are lines of equal pressure. On maps, pressure is adjusted to mean sea level (MSL), therefore eliminating elevation as a factor. MSL pressure is 1013 mb, although the mean range is from 1060 mb in the Siberian winter high-pressure 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 3–4 kilometre-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.

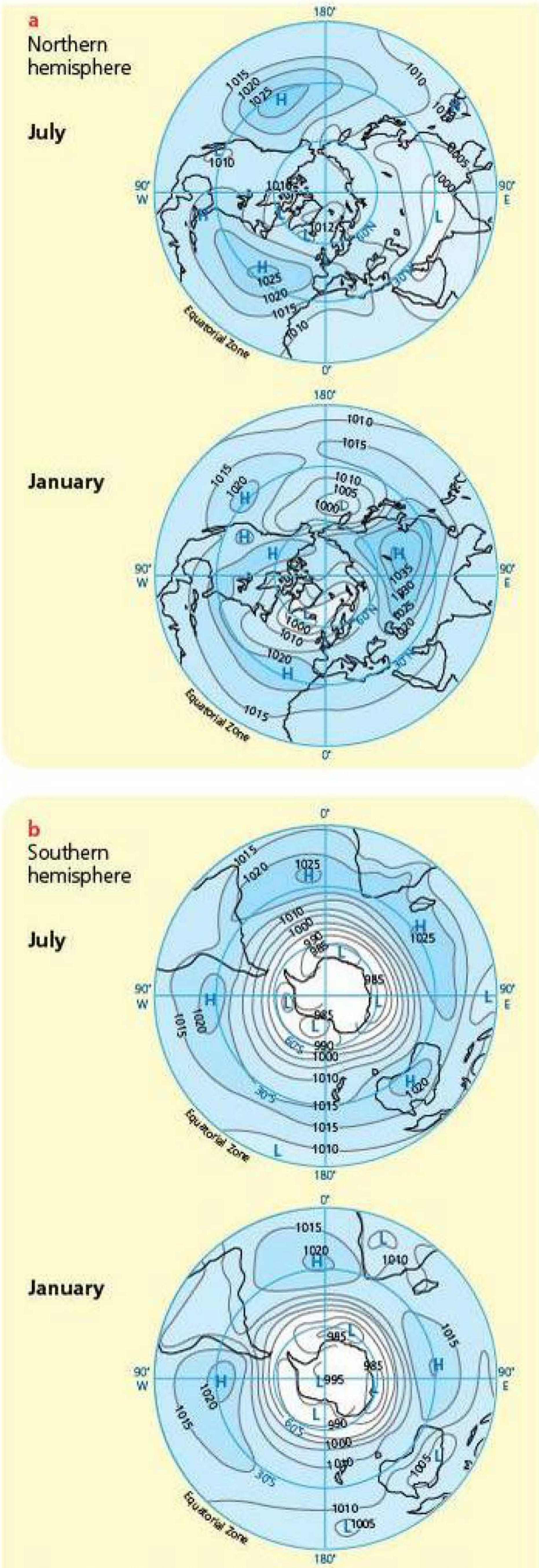
One of the more permanent features is the subtropical high-pressure (STHP) belts, especially over ocean areas. In the southern hemisphere these are almost continuous at about 30° latitude, although in summer over South Africa and Australia they tend to be broken. Generally pressure is about 1026 mb. 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: 1008–1010 mb. 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

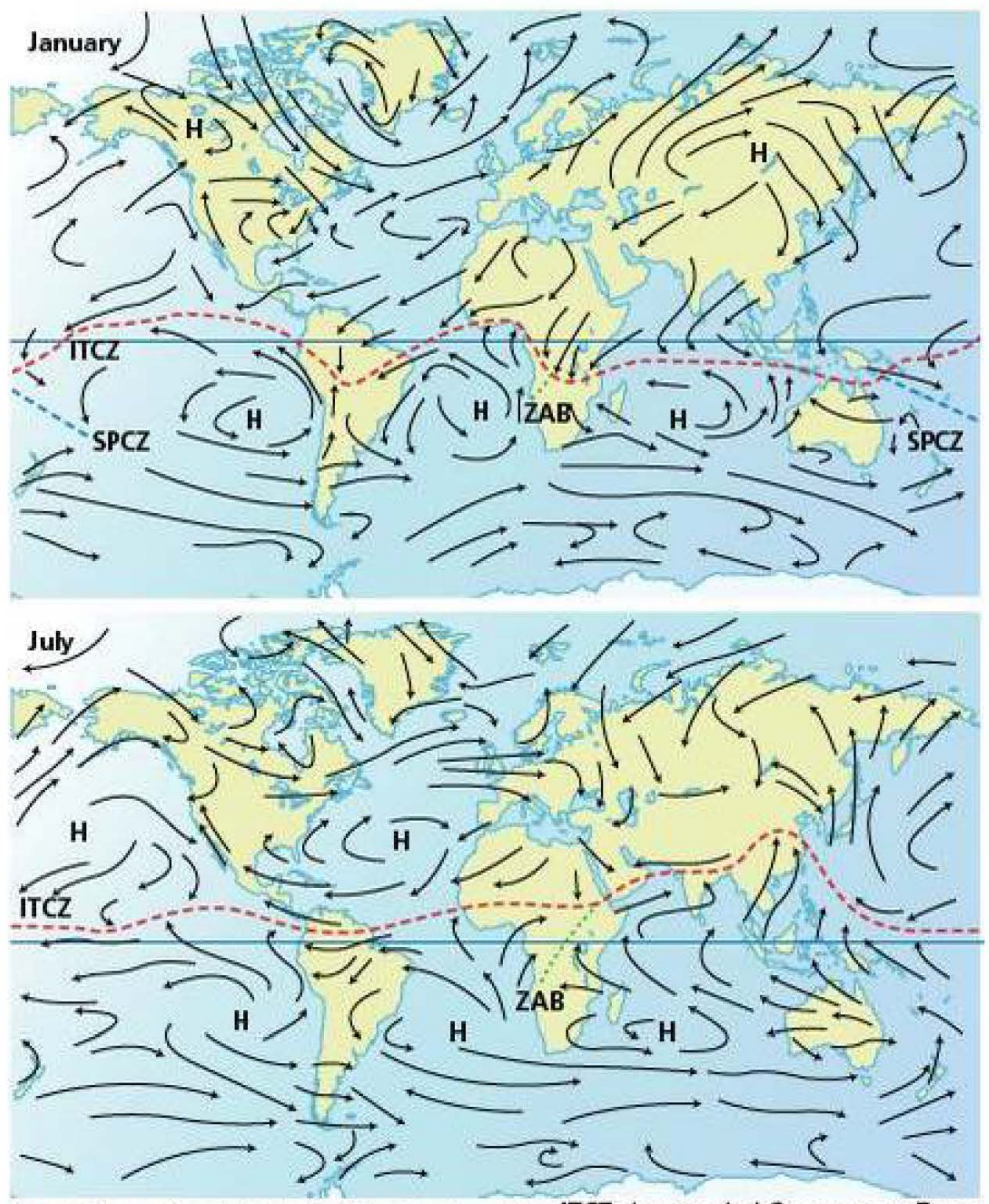
Section 2.2 Activities

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

--- ITCZ Intertropical Convergence Zone
..... ZAB Zaïre Air Boundary
--- SPCZ South Pacific Convergence Zone
H Centre of high pressure
— Equator

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.

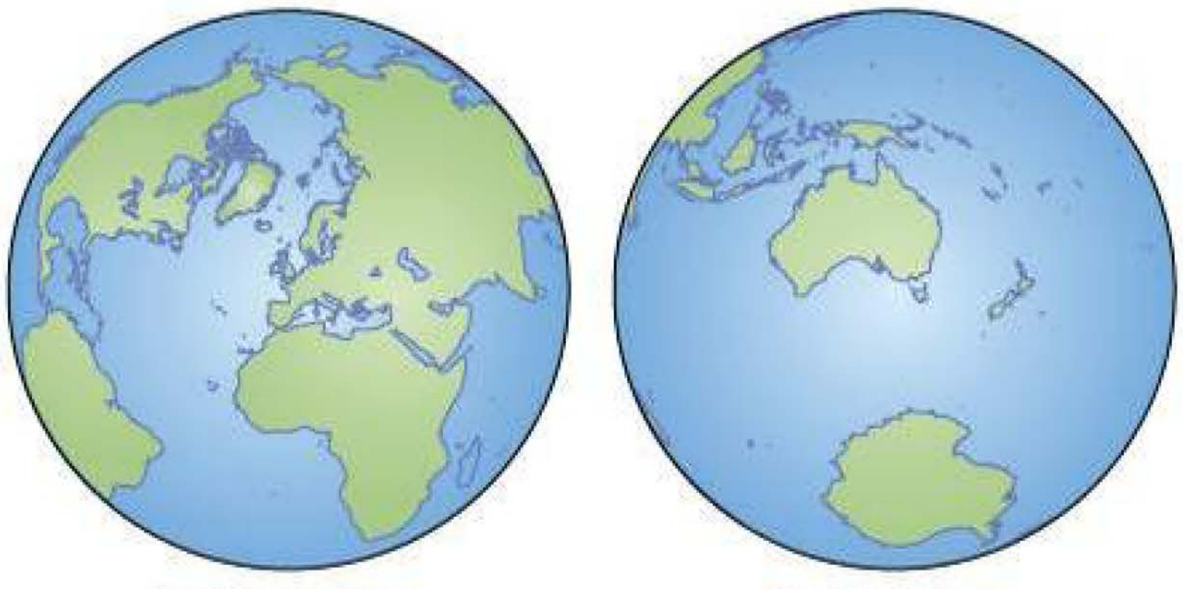


Figure 2.15 Land and sea hemispheres

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.

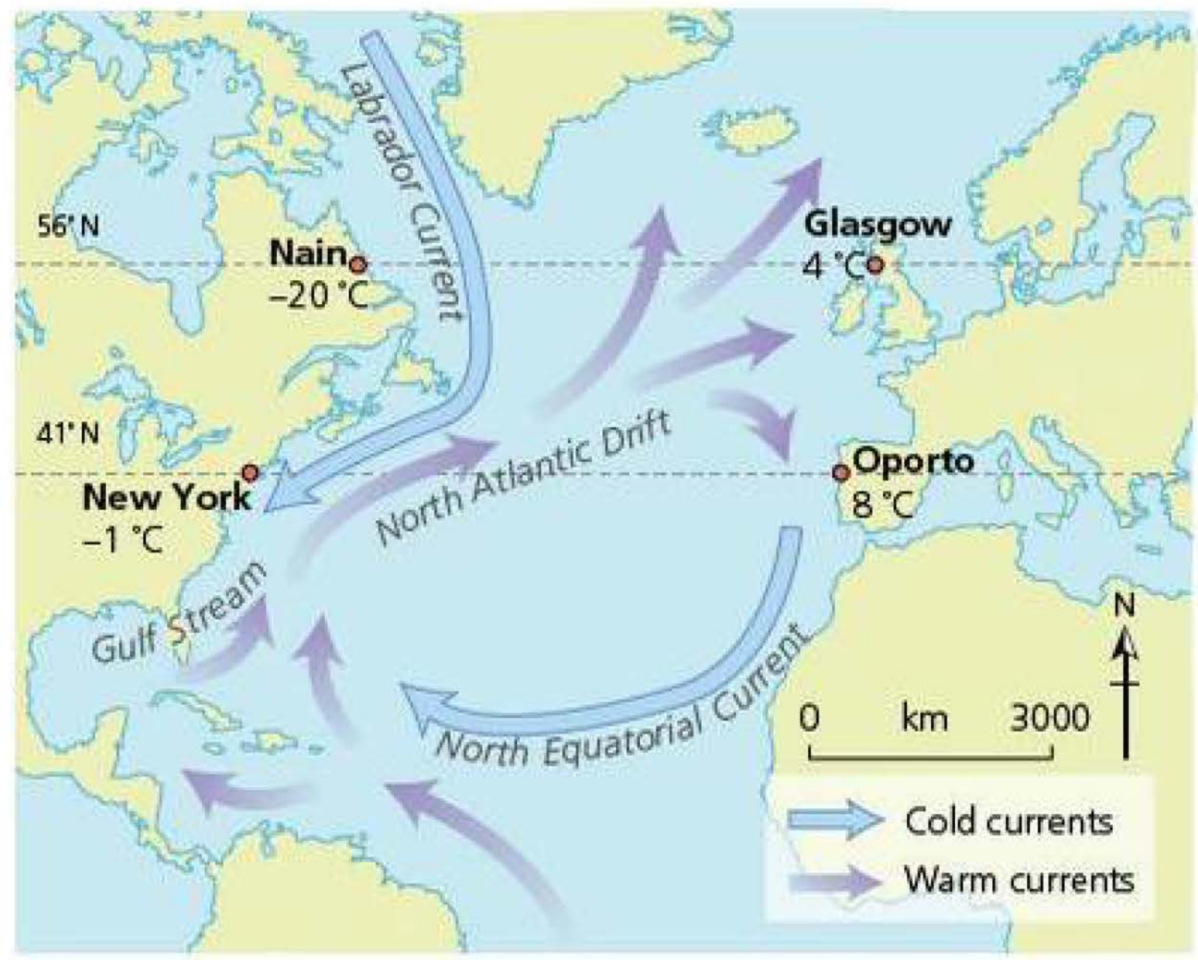
Water heats more slowly because:

- 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

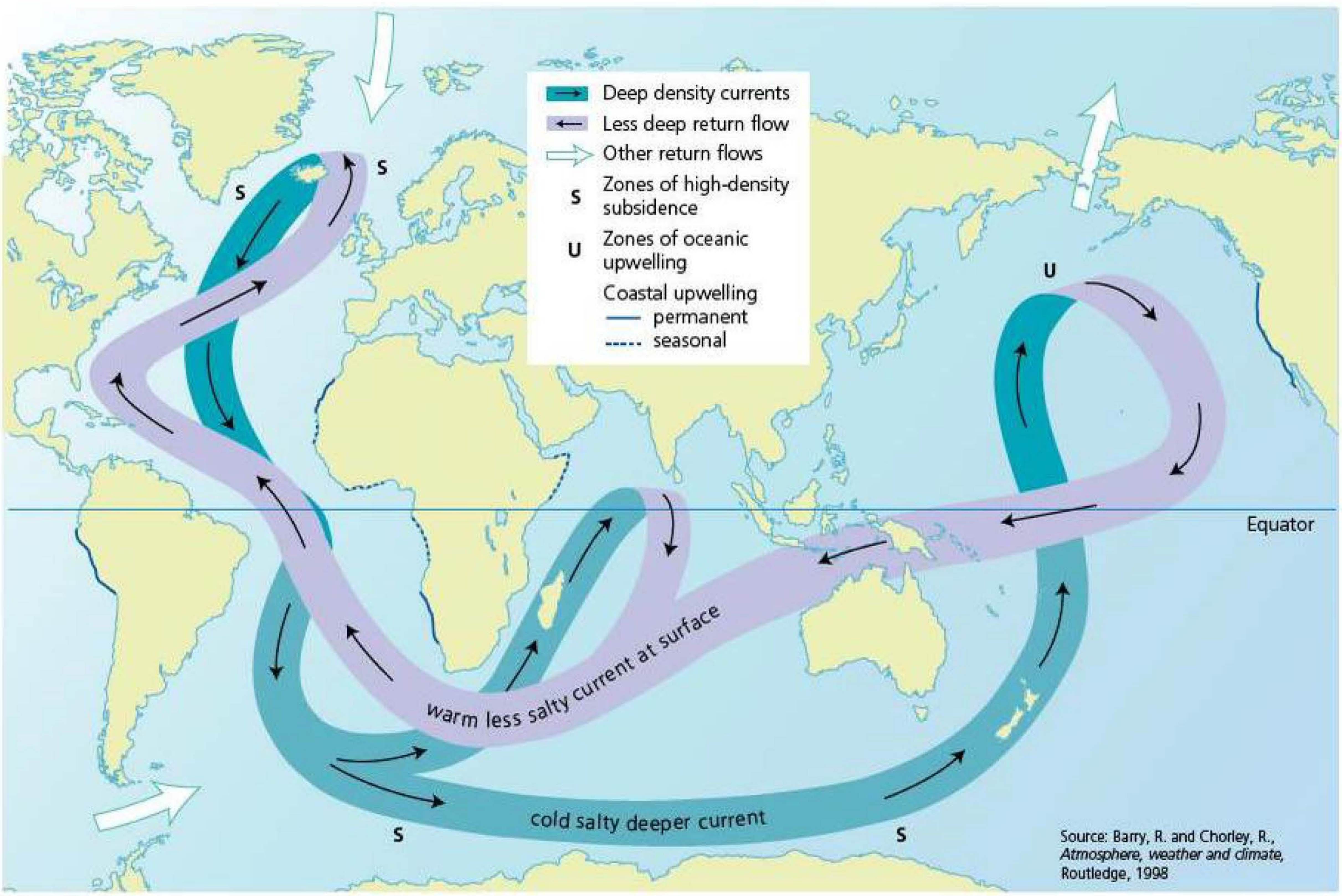
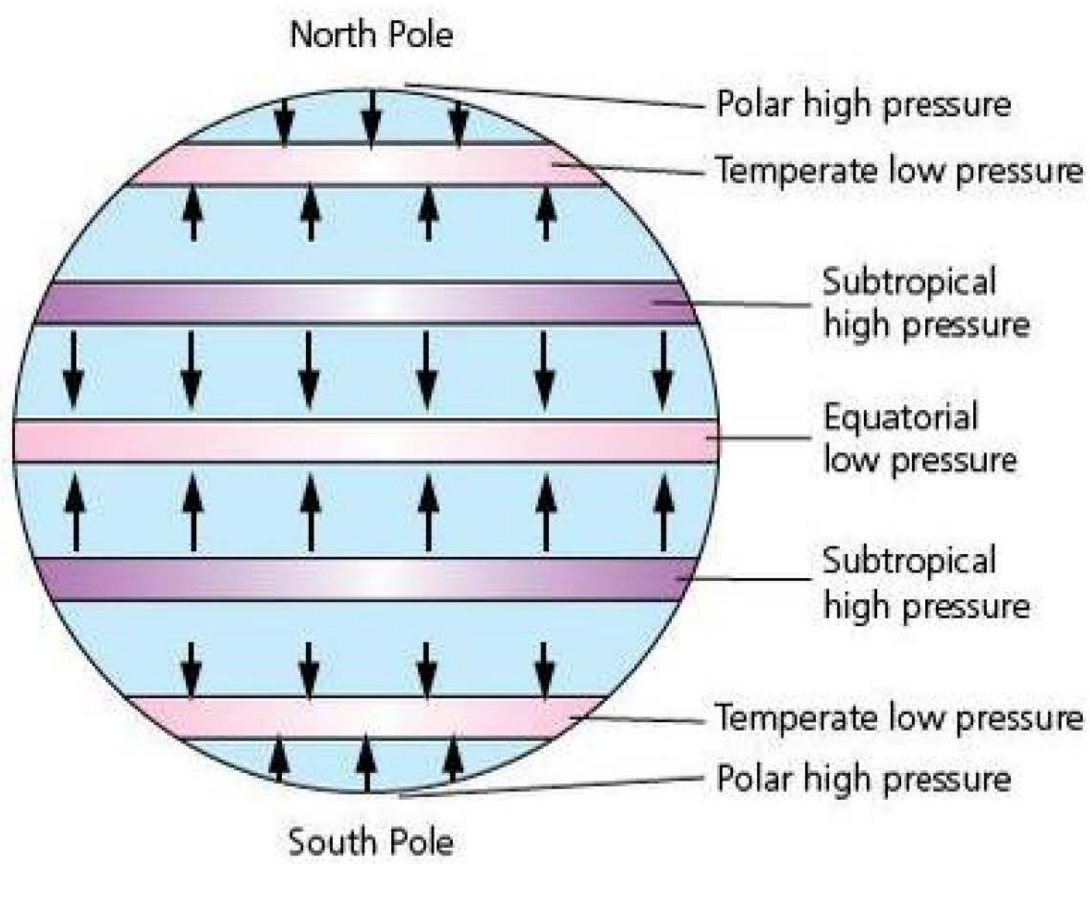


Figure 2.17 The ocean conveyor belt

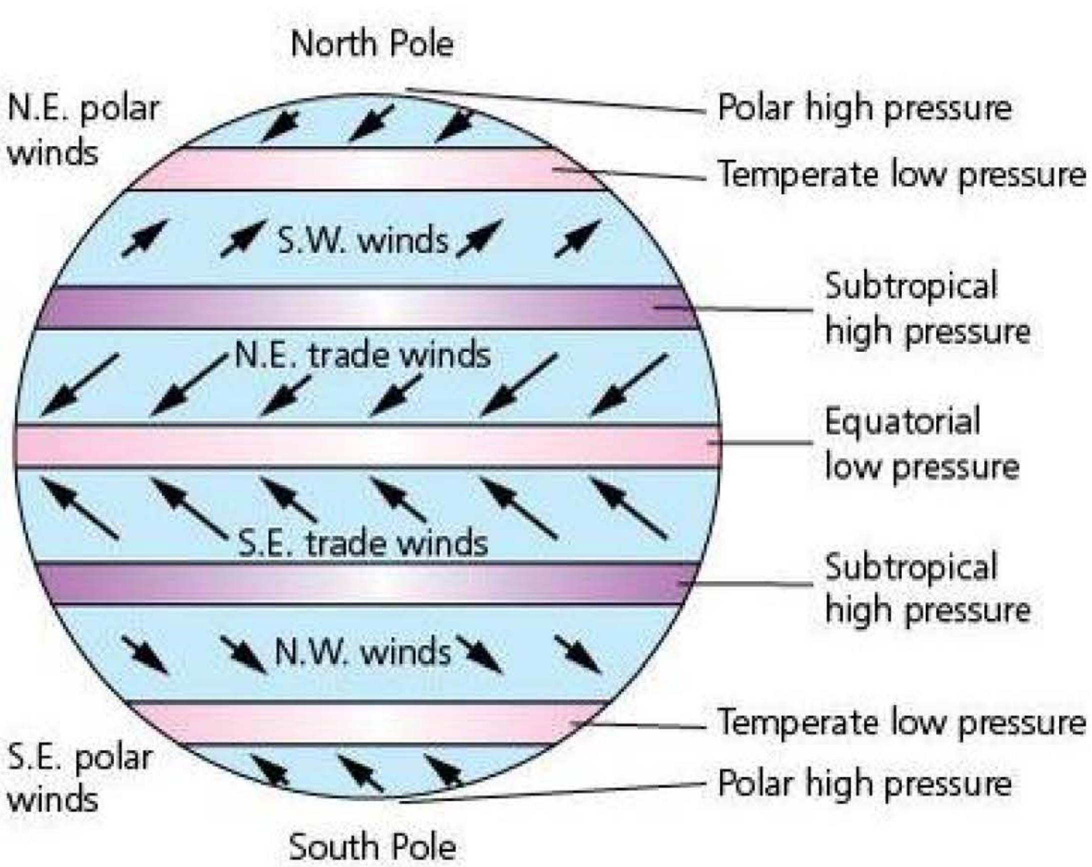
motion. There is thus a basic correlation between winds and pressure.

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.



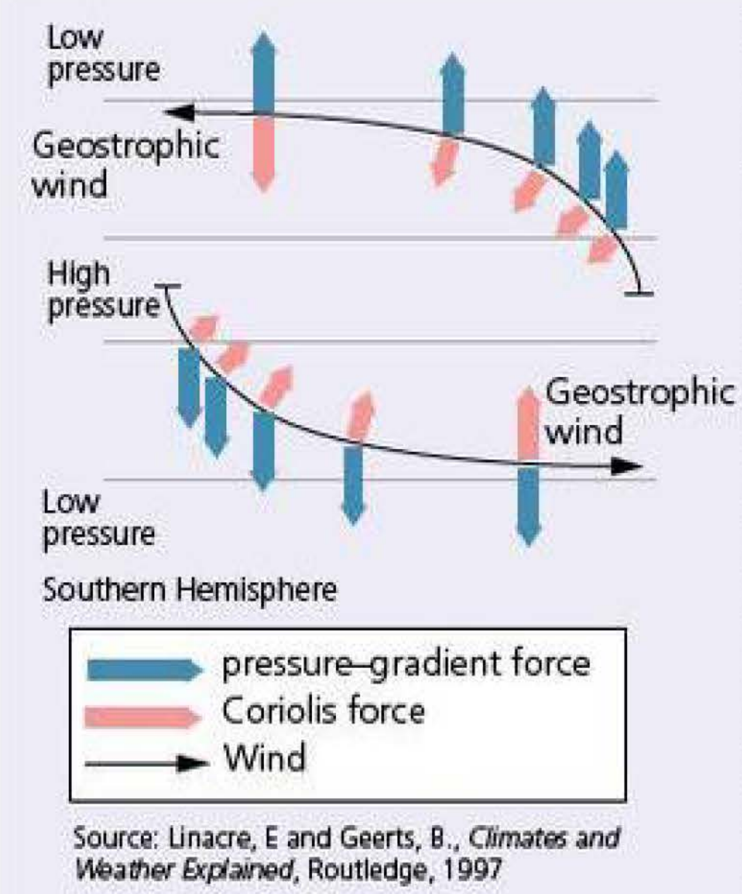
a How the winds would blow on a non-rotating Earth



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).

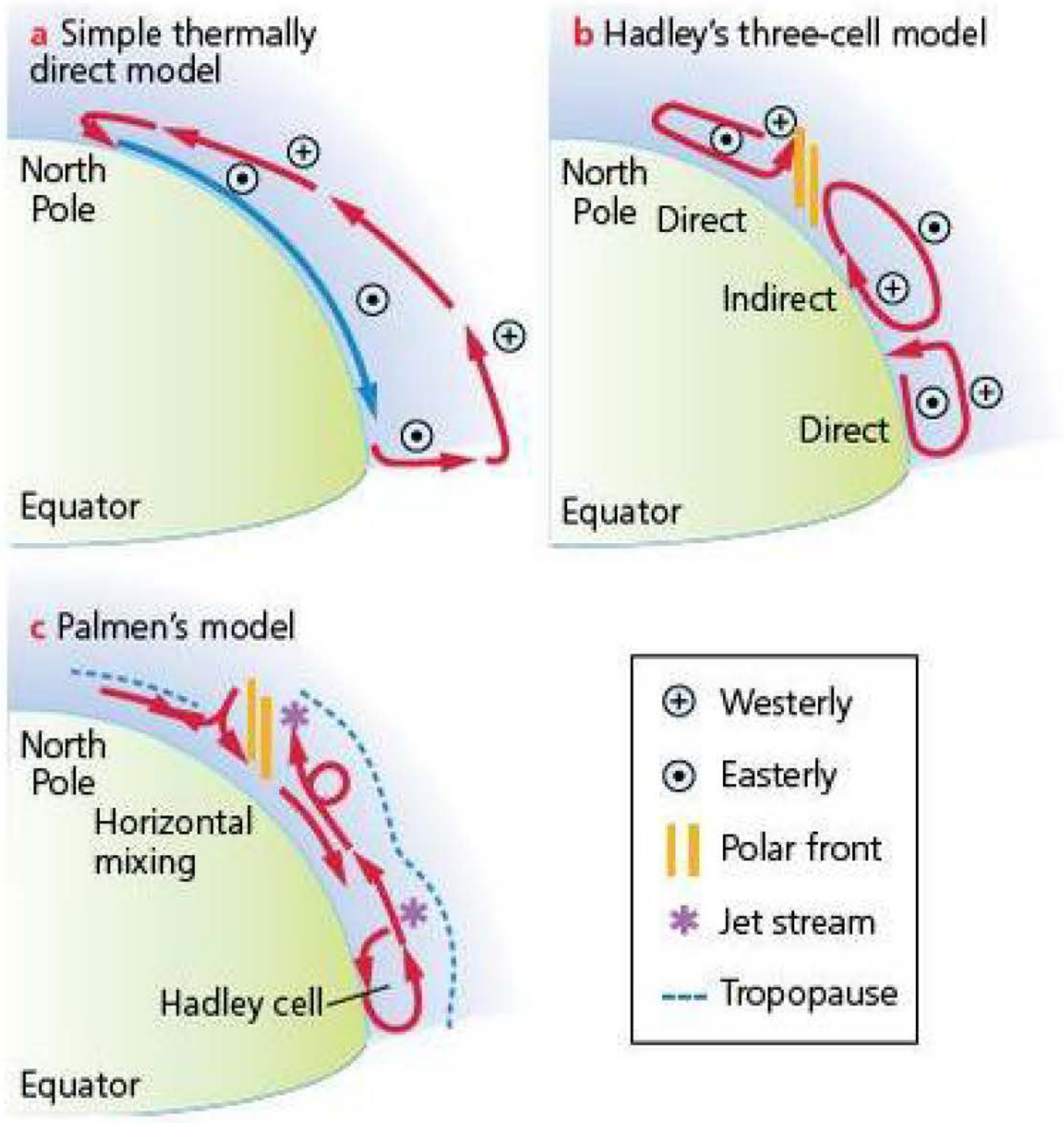


Figure 2.20 General circulation model

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.

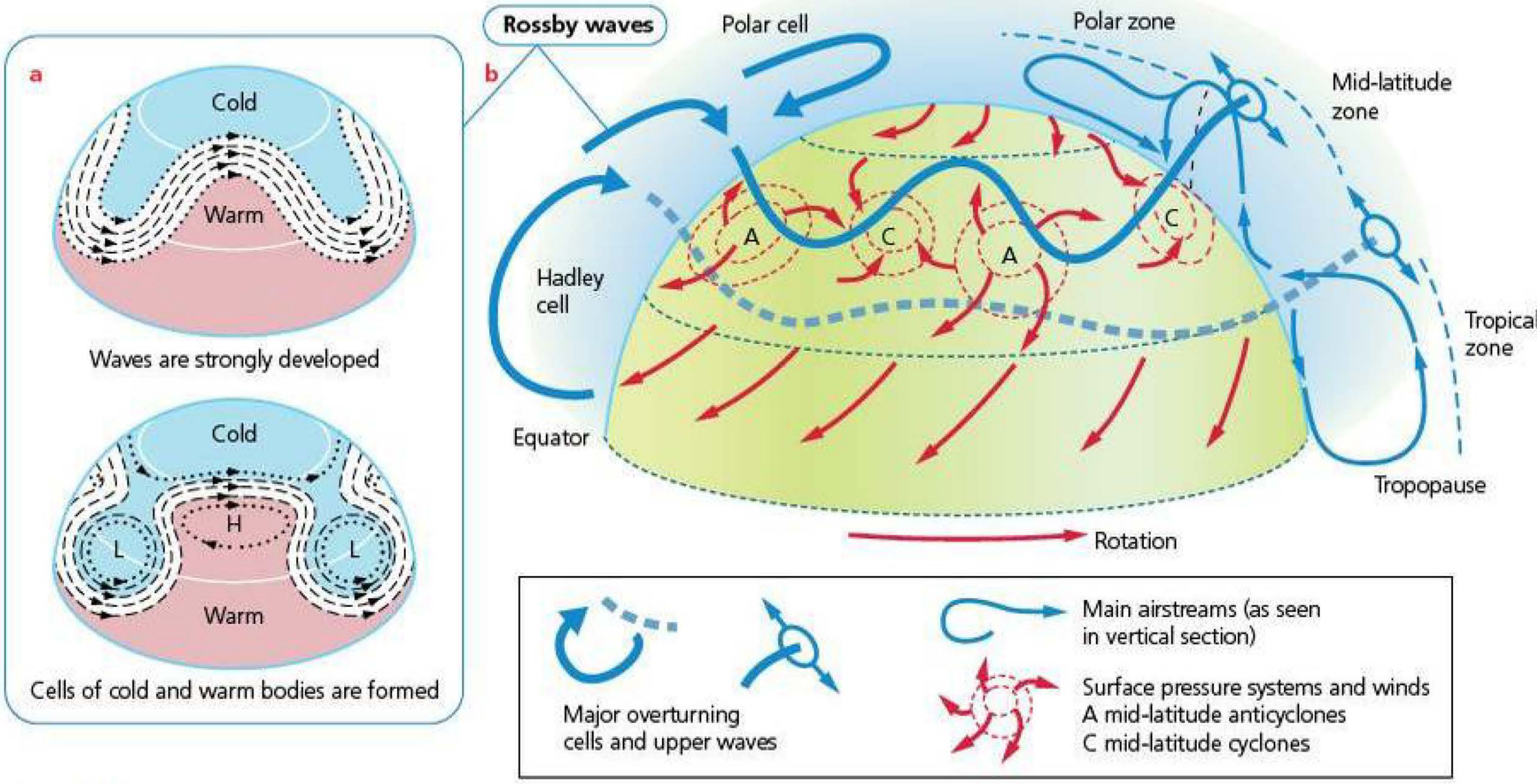


Figure 2.21 Rossby waves

- 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