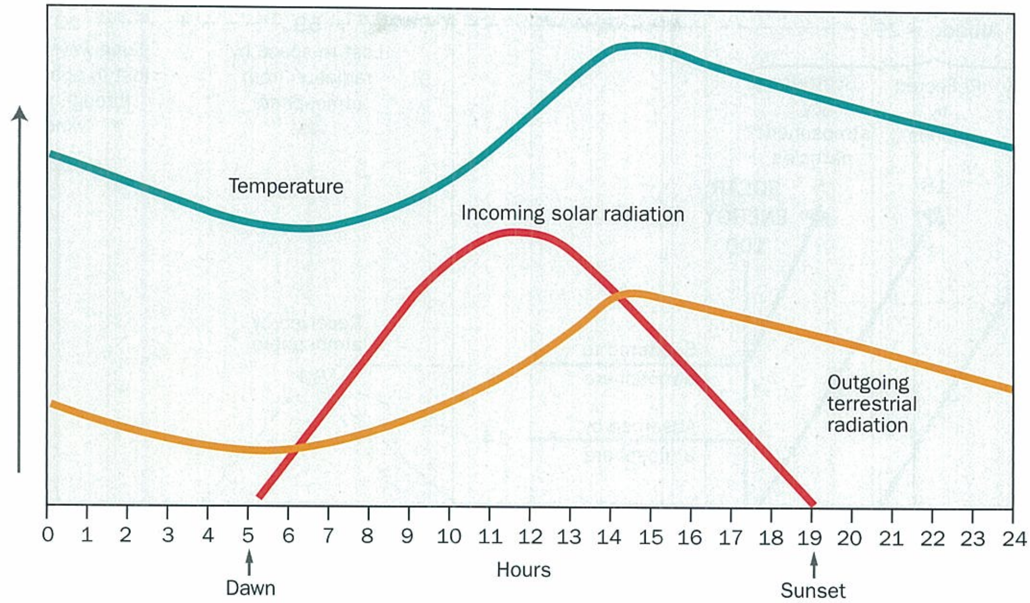


Fig. 2.10 Multiple line graph showing the influence of solar and terrestrial radiation on temperature during a day in mid-latitudes

calm because there is no warmer air coming in to mix with it. When water vapour comes into contact with a cold object whose temperature is below the **dew point** of the air, such as a leaf or spider's web, the water vapour will condense on the object, forming small water droplets known as **dew**. Latent heat, absorbed during evaporation, is released during the condensation process, adding warmth to the air near the ground.

The influence of cloud on night-time energy budgets: absorbed energy returned to Earth

A thick cloud cover at night acts as a 'blanket', keeping the Earth and lower atmosphere warm by absorbing and re-radiating the emissions of long-wave radiation from Earth to atmosphere and back to Earth. This results in little difference in temperature between day and night, especially when the day has also been cloudy.

Some of the Earth's long-wave radiation absorbed by clouds is re-radiated to space. The warmer the cloud, the more long-wave radiation is re-radiated. Little is radiated from high level clouds, such as those in Fig. 2.8, because their upper surfaces are cold.

Eventually a balance is achieved between incoming solar radiation and long-wave radiation to space.

The global energy budget

Variations in the energy budget occur from place to place and time to time. However, globally and in general, incoming solar radiation must have been balanced by outgoing terrestrial radiation because, if that was not so, the Earth's atmosphere would have been getting hotter or colder. As 71 per cent of incoming solar radiation is absorbed (48 per cent by the Earth and 23 per cent by 'greenhouse' gases in its atmosphere), those amounts must be radiated back to space to keep the balance, as shown in Tables 2.3 and 2.4. If global warming (page 62) is now occurring, these processes are no longer in balance.

Incoming short-wave radiation at the Earth's surface (estimates)	Outgoing radiation (estimates)	
absorbed by the Earth = 48%	latent heat transfer (evaporation):	25%
	sensible heat transfer (convection):	5%
	long-wave radiation direct to space:	12%
		total = 42%
	long-wave radiation absorbed by greenhouse gases in the atmosphere:	6%
		total = **48%**

Table 2.3 The surface energy budget of the Earth's surface

Gains		Losses
absorbed solar radiation:	23%	long-wave radiation from the atmosphere to space = **59%**
latent heat transfer (evaporation):	25%	
sensible heat transfer (convection):	5%	
absorbed long-wave radiation:	6%	
	Total = **59%**	

Table 2.4 The energy budget of the atmosphere

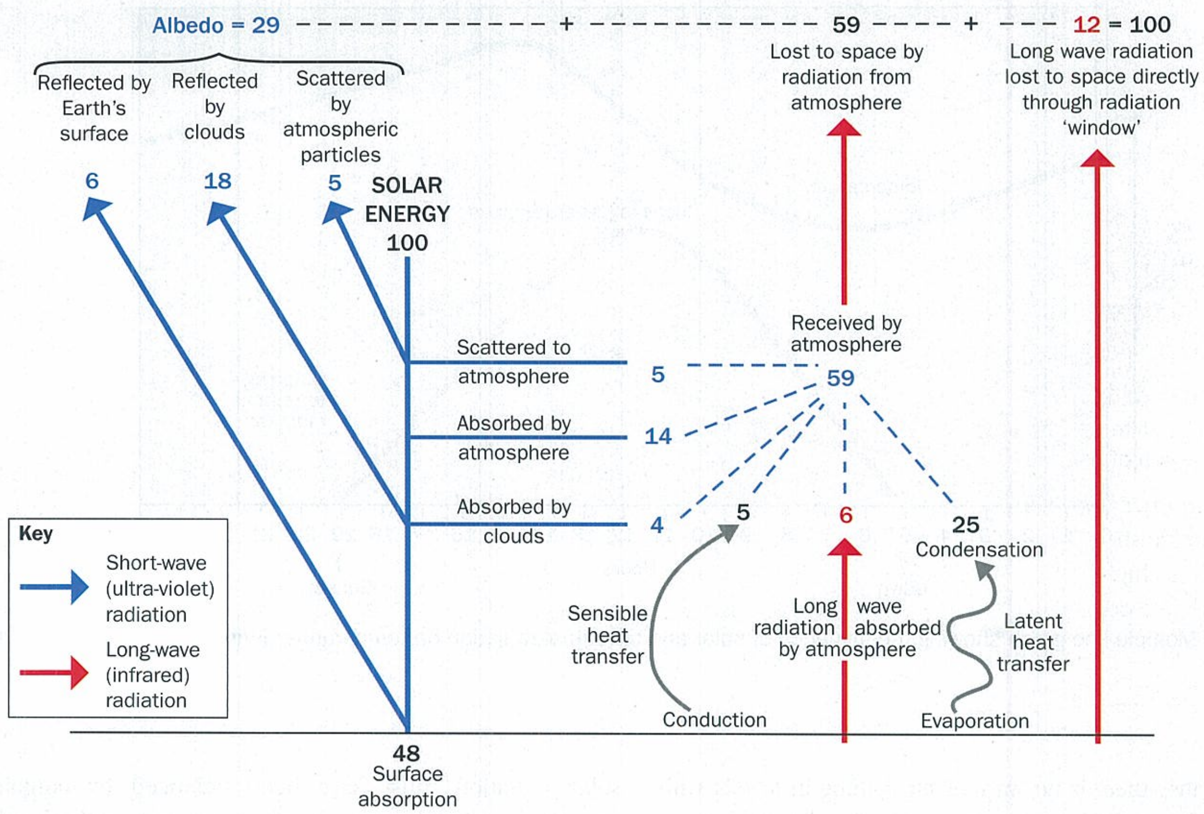


Fig. 2.11 The global energy budget (estimated figures)

As 59 per cent of the received solar radiation is radiated by the atmosphere to space while the surface of the Earth radiates only 12 per cent to space, most cooling by radiation occurs in the atmosphere and most radiative heating occurs at the Earth's surface. Energy is constantly being transferred around the Earth-atmosphere system for this to happen.

If the gases in the atmosphere did not absorb long-wave radiation, the surface temperature of the Earth would be up to 40 °C lower.

6. Produce a detailed key for Fig. 2.12 to explain the higher or lower albedo levels at (a) to (d).

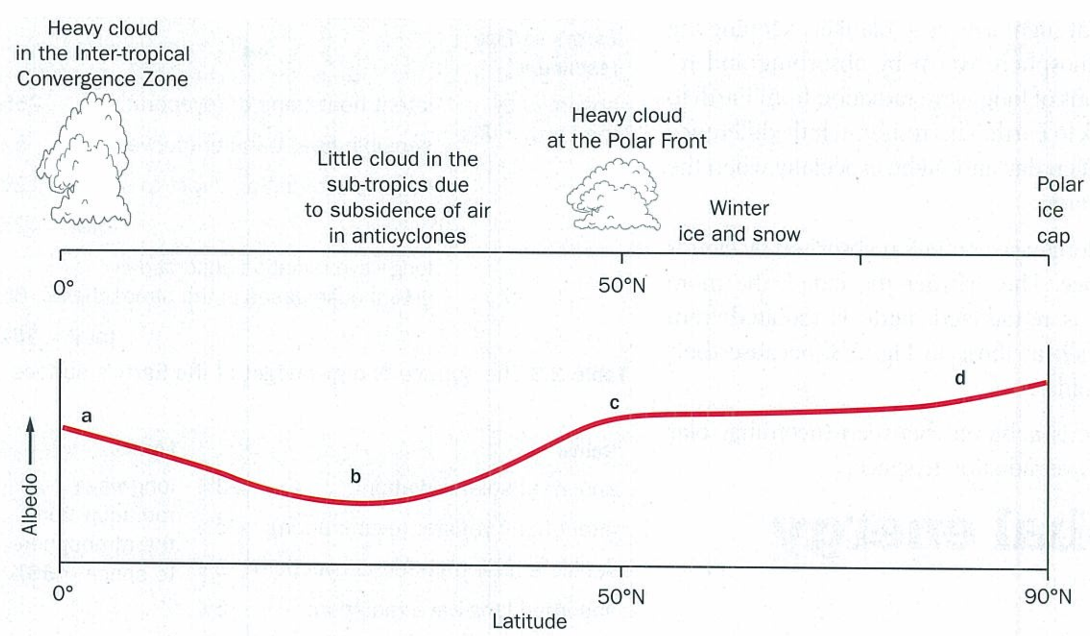


Fig. 2.12 The generalised pattern of albedo in the northern hemisphere

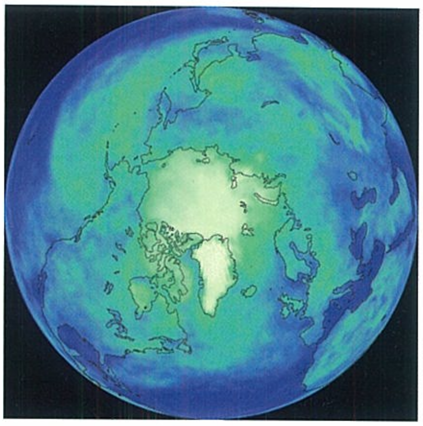


Fig. 2.13 Satellite image of reflected short-wave radiation in the polar region of the northern hemisphere in June 2009

On this summer image (Fig. 2.13), the white areas are the Greenland ice sheet and sea ice in the Arctic Ocean where reflectivity is highest (reaching 425 W/m²). Green areas have moderate reflectivity (about 212 W/m²), from snow on the ground in Eurasia and northern Canada. The darker land and ocean surfaces, shown in blue, have lower albedos so are absorbing more of the summer sun and warming.

The latitudinal pattern of radiation: excesses and deficits

7. Describe the variations in average annual solar radiation measured at the Earth's surface by detailing areas where it is highest at over 225 W/m² and areas where it is lower than 150 W/m². To what extent is the influence of latitude shown?

8. Suggest why total insolation received in the southern hemisphere at any latitude is lower than it is in the northern hemisphere.

Case study: Components of the energy budgets that influence the temperatures in Equatorial regions and hot deserts

	Equatorial regions	Hot deserts (latitude 15–30°)
Incoming solar radiation at the edge of the atmosphere	High (about 440 Watts/m² per year).	Less (about 340 Watts/m² per year).
Radiation at the Earth's surface	150–200 W/m²	250–300 W/m²
Absorption, scattering and radiation by cloud	Very high absorption and scattering by the cover of deep convective cloud in the afternoons and evenings. The high tops of the convective cloud are cold, so outgoing radiation to space is very little.	Low because it is cloudless, so the sun's rays are high intensity and outgoing long-wave radiation from the warm surface is very large.
Surface albedo	Low – tropical rain forest about 10 per cent (but shade from forest cover reduces surface temperatures).	High because soils are dry. Desert 28 per cent (rising to 40 per cent if there is sand cover).
Energy absorbed into the surface	Wet soils conduct energy down.	Little energy is transferred down into the rock or dry sand.
Sensible heat transfer	Strong uplift, especially in the daytime and early evening.	Strong uplift by day, strong conduction cooling at night and sinking of cold air from above in the high pressure zone.
Latent heat transfer	Very high because the air has a high moisture content supplied by evaporation from the many water bodies and transpiration from the forest cover.	Very low because the air is very dry.
Radiation balance	Positive, with a large difference between gains and losses.	Positive but with a smaller surplus.

Table 2.5 Energy budget differences between equatorial regions and hot deserts

The net effect of the energy budget in equatorial regions is hot temperatures all year, with daytime highs of about 30 °C dropping to 23 °C at night, giving a daily range of about 7 °C. The forest increases latent heat transfers but decreases sensible heat transfers.

By contrast, deserts have extreme diurnal (daily) temperatures all year. Daytime temperatures average about 38 °C but can reach 50 °C in summer, dropping down to about 15 °C at night and to 5 °C on winter nights.

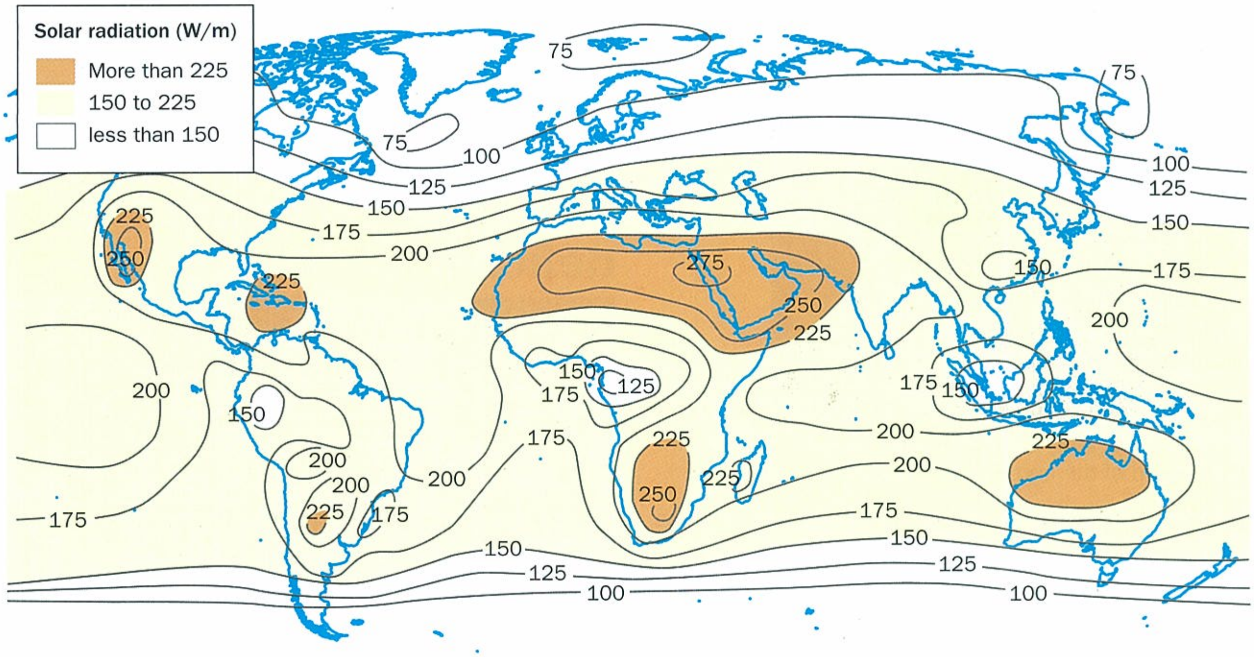


Fig. 2.14 Isoline map of the average annual distribution of solar radiation (W/m²) received at the Earth's surface

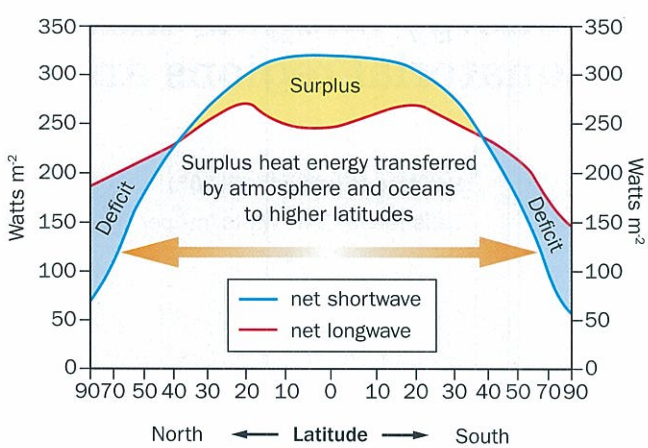


Fig. 2.15 The net radiation balance

The surplus and deficits shown in Fig. 2.15 would lead to the area between 40° and the Equator becoming increasingly warmer and the areas between 40° and the poles becoming increasingly colder if surplus heat energy from the Equator was not transferred to higher latitudes to inject warmth there. This transfer of heat is achieved by winds and ocean currents. They also move moisture.

Atmospheric transfers by wind belts

Winds are moving **air masses** - large bodies of air which are almost uniform horizontally in temperature and moisture characteristics. They are separated from adjacent different air masses by **frontal zones** along which there are usually large temperature and humidity gradients.

Air masses gain their characteristics in their source regions by prolonged contact with the ground or sea surface. Subtropical high pressure belts are the source regions for warm tropical air masses which undergo much heating. Heat energy is moved from these areas of surplus towards the poles by the south-westerly and north-westerly wind belts.

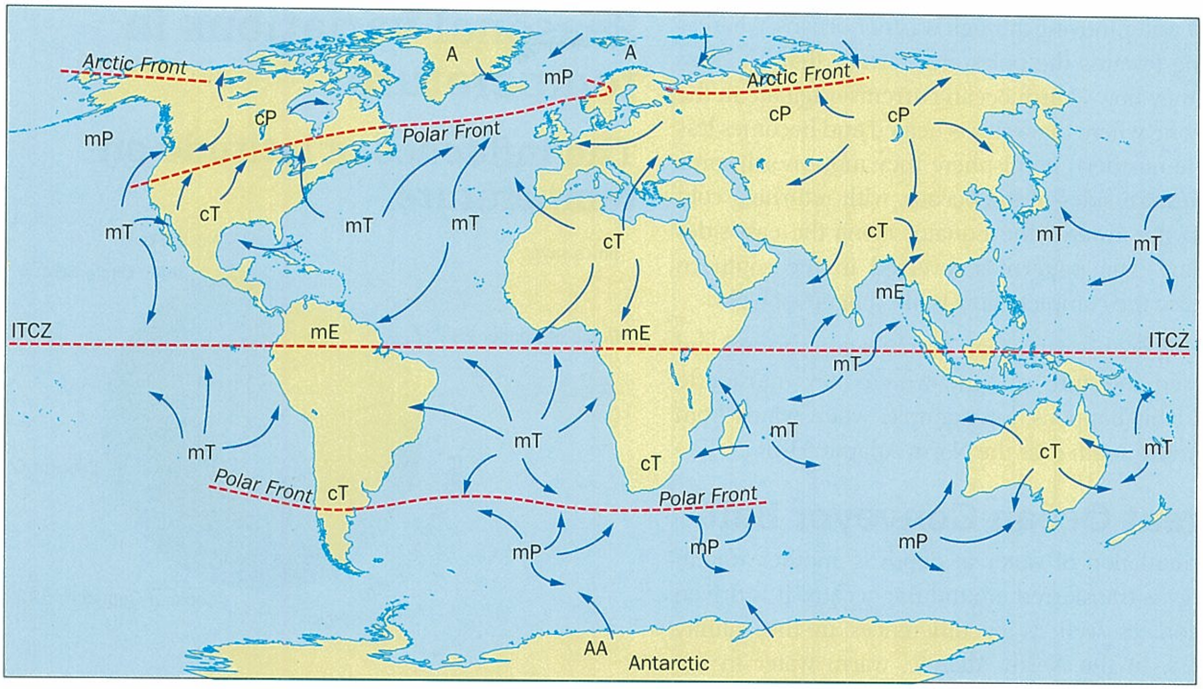
Source regions for polar air masses are high pressure systems over the continents. In winter the air becomes very cold by conduction cooling during contact with the cold land surface and in summer they are relatively cool. North-easterly winds in the northern hemisphere and south-easterly winds in the southern hemisphere move air towards the Equator to gain heat.

Air masses are further classified into continental or maritime according to where they formed, or were modified as they moved. A maritime track allows the lower layers of the air mass to become saturated with moisture, whereas a continental track leaves the air mass with low **humidity**. Thus, winds from the sea also transfer moisture from one place to another.

Further information about surface and upper wind transfers of heat energy is given later in the chapter.

Air mass	Temperature	Humidity
Equatorial maritime (mE)	warm	very moist
Tropical maritime (mT)	mild in winter warm in summer	moist
Tropical continental (cT)	very warm	dry
Polar maritime (mP)	cool	moist
Polar continental (cP)	cold	dry
Continental Arctic and Antarctic (cA and cAA)	very cold	very dry

Table 2.6 Characteristics of air masses



Key							
Air mass source region							
A	Arctic	cP	Polar continental	mE	Equatorial	mT	Tropical maritime
AA	Antarctic	cT	Tropical continental	mP	Polar maritime	→	Wind

Fig. 2.16 Winds moving from air mass source regions

Atmospheric transfers by ocean current

Ocean currents are mainly driven by prevailing surface winds and are another mechanism by which surplus heat energy in the tropics is distributed to higher latitudes. Warm currents transfer 20 per cent of the energy compared with the 80 per cent transferred by winds.

The pattern of surface ocean currents in January and July are similar. Ocean currents moving towards the Equator are

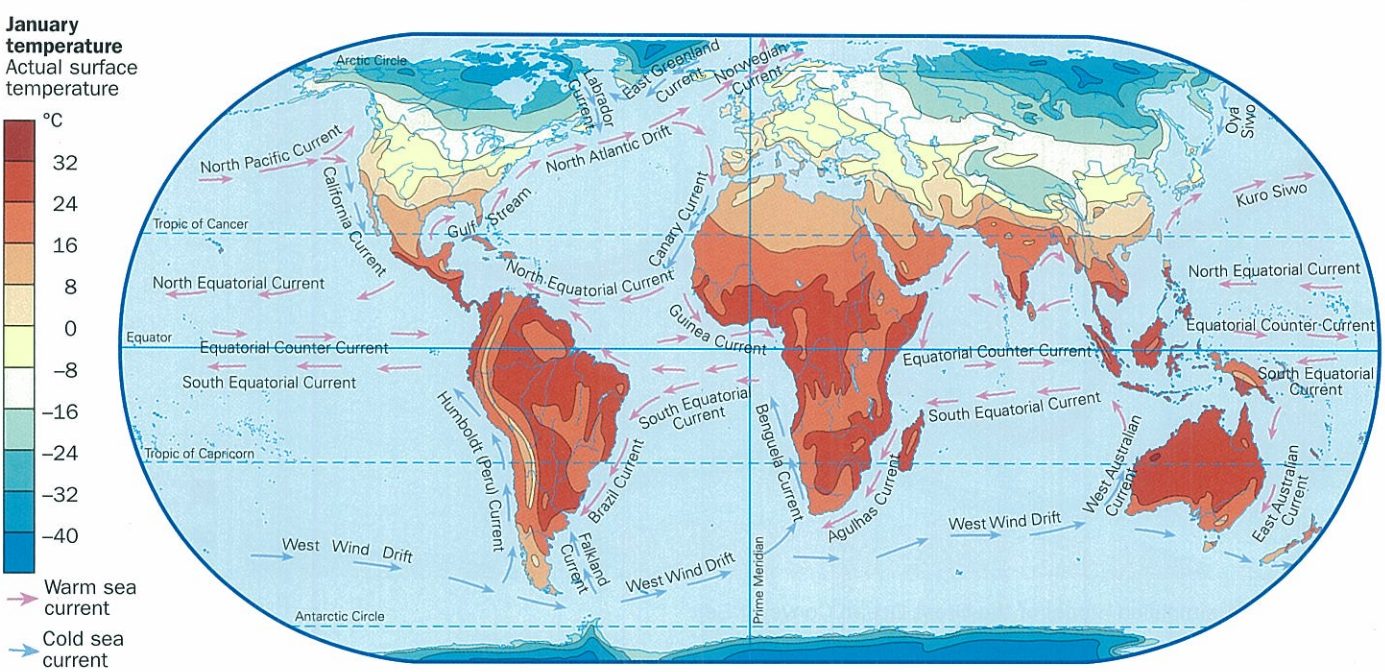


Fig. 2.17 Temperature and ocean currents in January

cold flows of water moving through warmer oceans, whereas those flowing towards the poles are warmer than the seas into which they flow. Warm ocean currents originate in the equatorial zone where the water is heated and becomes less dense. In the northern hemisphere they move northwards along the western sides of the oceans, with returning cold currents moving towards the Equator along the east sides of the oceans. The pattern is reversed in the southern hemisphere, so the complete flow is like a figure of eight.

Winds moving over warm ocean currents are warmed and gain increased moisture content, which they transfer to other areas. This happens, for example, when winds from Greenland move south over the North Atlantic Drift.

The Great Ocean Conveyor Belt

This slow circulation of water at depth is another way in which energy is transferred around the oceans. It is driven by convection, as well as by differences in the salinity of the waters. In the North Atlantic, warm water in the North Atlantic Drift heats the air. This loss of heat to the atmosphere makes the water colder and denser. Its density also increases because evaporation makes the water saltier. This cold, salty water sinks and moves towards the Equator. It flows at a depth of about 4 km to the Antarctic and then into the Pacific Ocean. By the time it reaches the north Pacific it has warmed enough to rise back to the surface. The warm water then moves into the Indian and Atlantic Oceans as a surface current, until it completes the conveyor belt and sinks again in the North Atlantic.

Seasonal variations in temperature

The influence of latitude on temperature

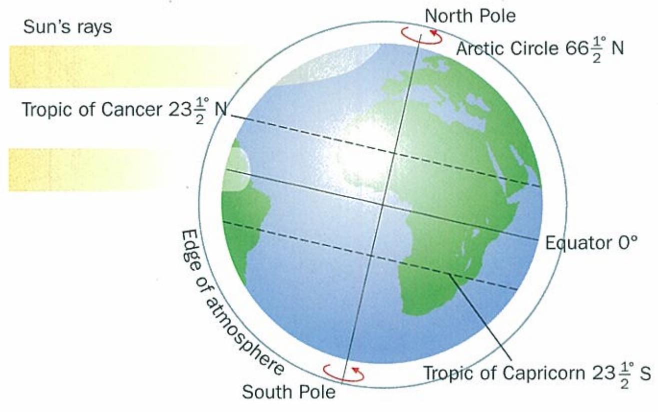


Fig. 2.19 The influence of latitude on temperature

Angle of the sun's rays

At the Equator the sun's rays are vertical, or nearly so, all year at noon. Insolation is intense because a given amount of solar radiation heats a relatively small part of the Earth's surface. Towards the poles the sun's rays strike the surface at increasingly lower angles, increasing the area heated by the same amount of solar radiation and reducing the intensity of the insolation and temperatures.

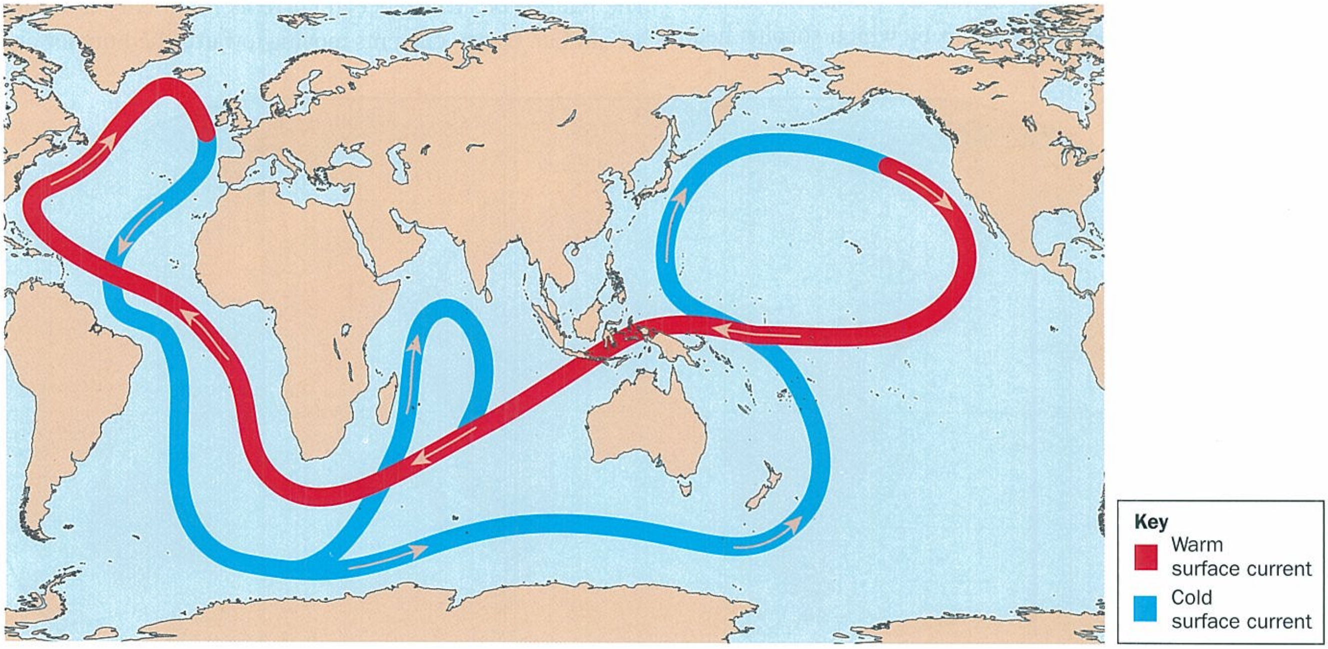


Fig. 2.18 A simplified model of the Great Ocean Conveyor Belt

Thickness of atmosphere

The higher the latitude, the greater the thickness of atmosphere though which the sun's rays have to pass. Consequently, absorption, scattering and reflection of solar radiation increase with increased latitude (except in the polar region where the air is clean and contains very little water vapour).

If latitude was the only influence on temperature, the world would be hottest at the Equator and become progressively colder towards the poles. This is not the case.

Mean Sea Level Temperatures (°C) – January

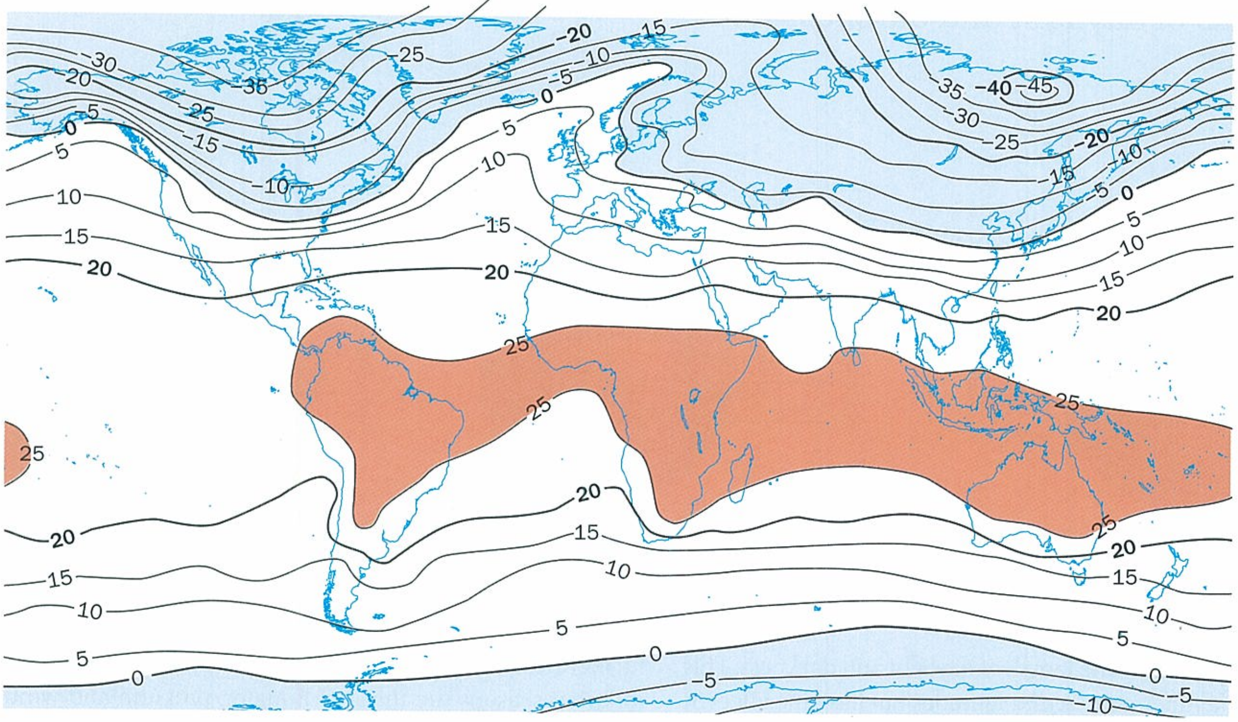


Fig. 2.20 Isotherm map showing the global pattern of temperatures in January

Key to Figs 2.20 and 2.21

- Above 25 °C
- 0–25 °C
- Below 0 °C

Mean Sea Level Temperatures (°C) – July

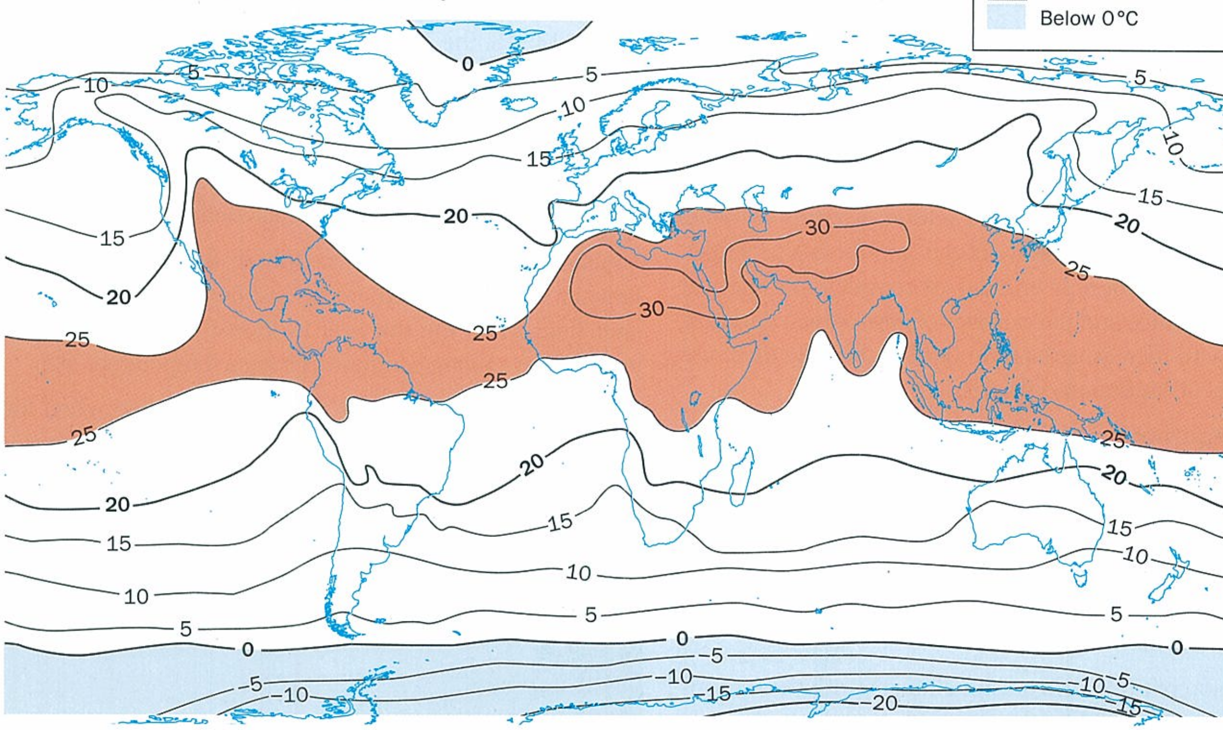
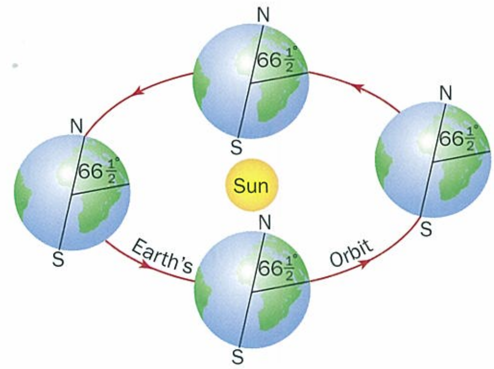


Fig. 2.21 Isotherm map showing the global pattern of temperatures in July

9. Describe how the temperature pattern in January (Fig. 2.20) shows the influence of latitude and comment on ways in which the pattern is distorted by the distribution of land and sea.

10 Describe, with reasons, how the July temperature pattern (Fig. 2.21) differs from that of January.



Lengths of daylight and darkness and seasons

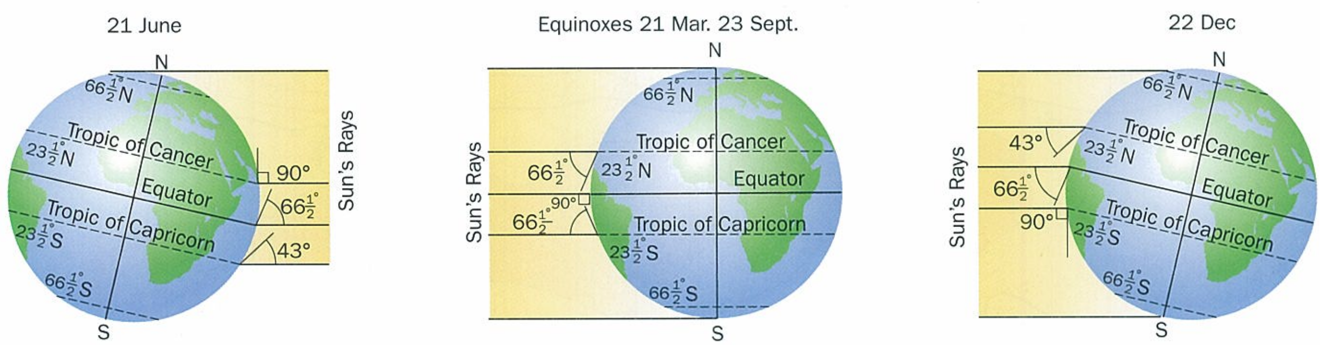


Fig. 2.22 The orbit of the Earth round the sun causes varying angles of the noonday sun on different dates in the year

Latitude influences the length of daylight and darkness. This alters the temperature in mid-latitudes but has little effect in equatorial regions where there are twelve hours daylight and twelve hours darkness all year. Lengths of daylight increase towards the pole in the summer hemisphere that is tilted towards the sun, until at latitude 66½° there is one day with 24 hours of daylight at the summer **solstice**. The number of days of total daylight continues to increase to the pole, which has six continuous months of daylight. However, the angle of the sun is so low there that it has little heating power.

Meanwhile, in the other hemisphere, the Earth is tilted away from the sun and it is the winter season. Towards the pole, the days become increasingly shorter and the nights longer. At latitude 66½º there is one day of complete darkness at the winter solstice and the number of days of total darkness increase with increasing latitude until at the pole there are six months of continuous darkness.

At the **equinoxes** (21 March and 23 September) all latitudes are bisected equally by the circle of solar illumination, so all have twelve hours daylight and twelve hours darkness. However, the mean annual **thermal Equator** (the zone of maximum heating) is at latitude 5°N, not the Equator, because of the greater heating of the northern hemisphere continents.

The position of the overhead sun changes with the seasons because the Earth orbits the sun with its axis at an angle of 23½° to the plane of orbit. The sun is never overhead further north than 23½°N or further south than 23½°S. In different seasons the thermal Equator, pressure and wind belts shift slightly in the direction of the position of the overhead sun.

The sun is overhead at 23½°N, the Tropic of Cancer, on 21 June (summer solstice for the northern hemisphere) and at 23½°S, the Tropic of Capricorn, on 22 December (summer solstice for the southern hemisphere), so it might be expected that these months would be the hottest for those hemispheres. They are not. The hottest months are July and January respectively, a month later, because there is a temperature lag as the ground heat builds up.

Similarly, in the northern hemisphere, the coolest month is January, and in the southern hemisphere it is July, a month later than when the sun is at its lowest in the sky at noon because net cooling continues, giving a temperature lag.

For any latitude, summer is the warmest period when the noonday sun is at its highest angle in the sky and winter is the coldest period when it is at its lowest angle. There are no seasons based on temperature in equatorial regions.

11. Use Figs 2.24a and 2.24b to compare the locations with the highest radiation gains in January and July. Explain why these locations have the greatest gains.

12. Compare the areas with the greatest loss in January and July. Explain their locations.

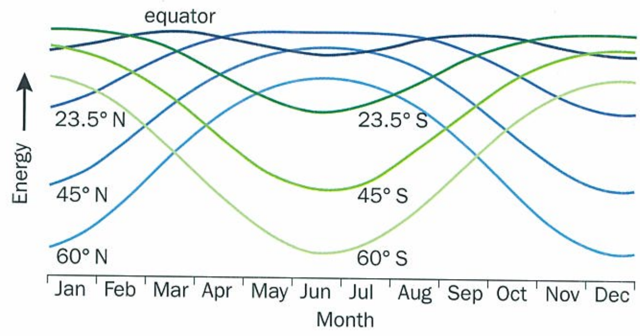


Fig. 2.23 Multiple line graph showing how solar energy received at noon varies with latitude during a year

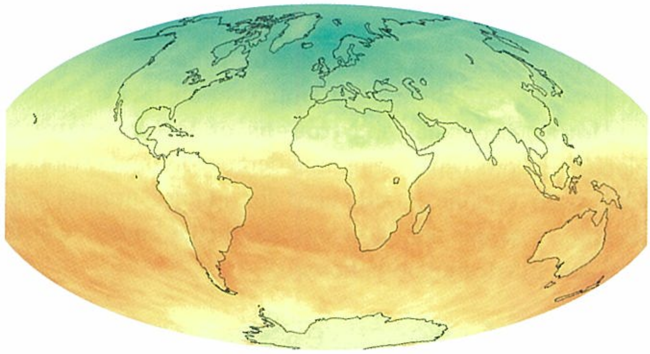
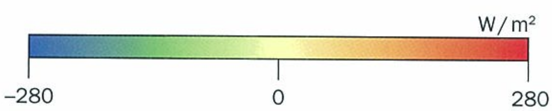


Fig. 2.24a Image of net radiation in January compiled from satellite data



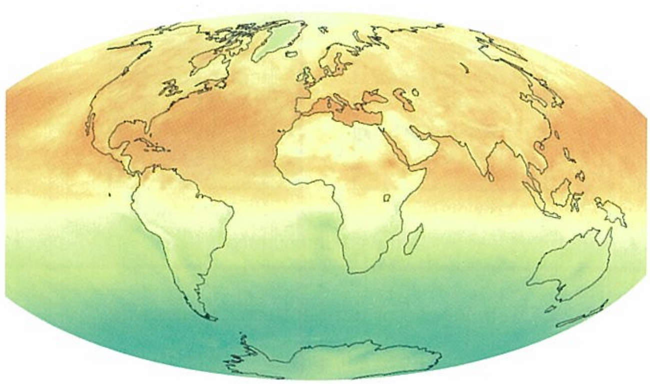


Fig. 2.24b Image of net radiation in July compiled from satellite data

In the tropics there is a net gain. In the middle and high latitudes of the southern hemisphere, lines of constant albedo and outgoing radiation are parallel to lines of latitude but this is not so in the northern hemisphere where differences between land and sea are clear. In middle to high northern latitudes in July there is greater outgoing long-wave radiation over the continents, which are warmer but albedo is greater over the oceans because of their greater cloud cover.

The influence of the distribution of land and sea

Land	Sea
Lower reflectivity (except when covered by ice and snow), so most radiation is absorbed.	Higher reflectivity, especially when the sun is low, so the sea absorbs less energy.
Heat is confined to near the surface because most surfaces are poor conductors of energy.	The sun's rays can penetrate deeper and convection currents in the sea also distribute the heat to greater depths.
Land has a low specific heat, so a certain amount of heating will raise the temperature of land much more.	The sea has a high specific heat, so the same amount of heating will raise the temperature of water less.
Little energy is used up in evaporation as there is less water.	Large amounts of energy are used in evaporation, especially in lower latitudes.

Table 2.7 Factors resulting in the differential cooling and heating of land and sea

Land heats and cools much more quickly than the sea which, like all water bodies, retains its heat for longer. Places near the sea have equable temperatures, with cooler summers and warmer winters than places inland. Small temperature ranges are typical of **maritime climates** where temperatures are moderated by proximity to the sea, while very large temperature ranges characterise **continental climates** with their seasonal temperature extremes.

The effects of continentality are especially marked in the northern hemisphere where large land masses stretch into high latitudes. In north-east Siberia annual temperature ranges are more than 60 °C. Verkoyansk, 67 °N, is the 'cold pole' of the Earth. It recorded a temperature of -68 °C in February 1982. Its long-term February mean is -45 °C and the July mean, its warmest month, is 14 °C, so its annual range is 59 °C. Fig. 2.25 shows the increase in annual temperature ranges eastwards from the Atlantic Ocean coast to the heart of Eurasia.

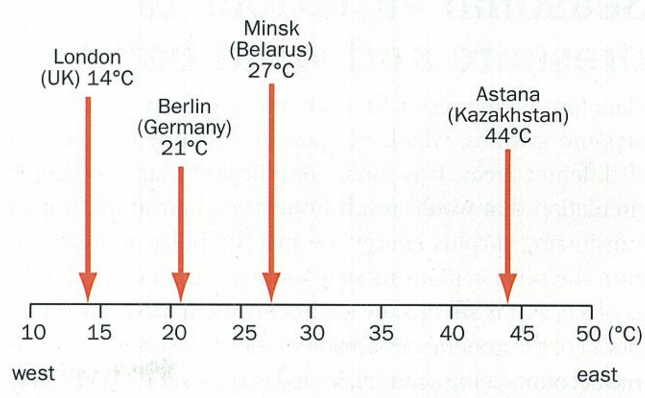


Fig. 2.25 Changes in annual temperature range along (or close to) 52½ °N

Winds are involved in sensible heat transfer. Warm winds move towards the poles, transferring heat to higher latitudes; cold winds move cooler air towards the Equator. The **prevailing wind** (most frequent wind) will influence the temperature of a place more than any other wind. The south-westerlies are Europe's prevailing winds and, as their source is the warmer ocean in winter, they bring warm air to make coastal western Europe's winter temperatures about 11 °C warmer than average for the latitude. By contrast, in the same latitudes, eastern Canada's prevailing winter winds are from the north-west and bring bitterly cold Arctic air over the area, resulting in January temperatures being at least 2 °C lower than average for the latitude.

The influence of ocean currents

Ocean currents change the sea temperatures and the temperatures of the air above them but can only affect temperatures on nearby land if an onshore wind blows over them.

13 How do the isotherms on Figs 2.20 and 2.21 indicate the influence of the Humboldt (Peruvian) and Benguela currents off the west coasts of South America and Africa?

14. Use information on Fig. 2.17 (influence of ocean currents) to explain the northwards curves in the isotherms over the northern hemisphere oceans in January, as shown on Fig. 2.20.

The influence of altitude

Air temperature decreases as altitude increases because the air becomes thinner and contains less water vapour to absorb the Earth's long-wave radiation, so at night there is very rapid heat loss. During the day, rock surfaces in the sun become very warm because almost all the insolation reaches the ground. The rate of decrease in temperature with altitude in *still* air varies but averages about 0.65°C for every 100 metres.

Seasonal variations in pressure and wind belts

Planetary winds result from air moving from high to low pressure systems, which are powered by unequal heating of different areas. It is more complicated than the simple circulation that would result from heated air at the tropics transferring surplus energy towards the poles and cold air from the polar regions moving towards the Equator to take its place, as it is affected by the Earth's rotation. A tri-cellular model of the general circulation of the atmosphere took this into account. Being a model, it tried to represent what usually happens, but there are areas where, and times when, it does not fit reality. In order to understand the model, knowledge of both the influences on pressure and on winds is necessary.

Seasonal variations in pressure

Pressure is the weight of the atmosphere and varies with height. Pressure maps usually show the pressure that is exerted at the Earth's surface reduced to sea level. Average pressure is 1013 millibars (mb) at sea level at latitude 45°. Pressure is always reduced to sea level because, if not, a pressure map would be like an inverted relief map – as there is less air in a column of air above a mountain than above the adjacent plain, so it would weigh less. Pressure on maps is shown by **isobars** (lines of equal pressure reduced to sea level). High pressure areas are surrounded by lower pressure and are not necessarily above 1013 mb. They are known as **anticyclones** if fairly circular or **ridges of high pressure** if they are elongated.

Low pressure areas are surrounded by higher pressures, and are not necessarily below 1013 mb. They are known as lows (**cyclones**) if circular and **troughs of low pressure** if elongated. Lows can develop into depressions and cyclones. (The term cyclone is also used for intense tropical low pressure systems that develop into fierce storms called cyclones, hurricanes or typhoons.)

Types of pressure change

Thermal pressure changes

When air is heated, it expands and becomes lighter than an equal column of cooler, heavier air because there is less air in the column.

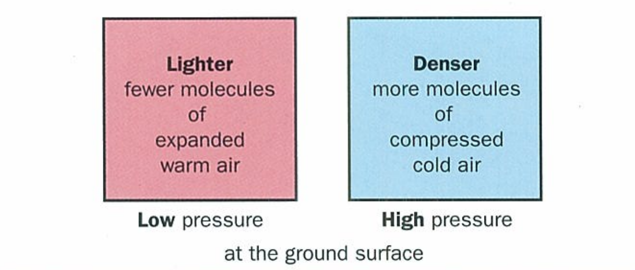


Fig. 2.26 The influence of temperature on pressure

Changes in temperature alter pressure at the surface; air pressure falls when its temperature rises and air pressure rises when its temperature falls.

Dynamic pressure change

If a glass of liquid is spun, **centrifugal force** sends the liquid to the sides and reduces the pressure in the centre. Pressure at the sides is increased and so is frictional drag. This type of pressure change affects winds because of the rotating earth. As the earth rotates, the atmosphere rotates with it but is also able to move freely. The shape of the Earth causes angular momentum to change with latitude. It depends on the distance from the axis of rotation of the Earth (a line between the two poles). Air moving towards the poles is getting closer to the axis of rotation so it speeds up (e.g. in the **jet streams**) while air moving towards the Equator slows down because it is getting further from the axis of rotation.

15. Explain why the permanent low pressure along the Equator and the permanent high pressure at the poles are thermally induced.

Seasonal variations in wind belts

Pressure and the pressure gradient force

High pressure areas have outflowing winds and low pressure areas have in-blowing winds. The speed of the air movement is determined by the pressure gradient - the difference between the high and low pressure systems. If the difference is great, the pressure gradient is high and the wind will be strong. Little difference results in weak winds. The pressure gradient is steep if isobars are close together and gentle if they are far apart.

The force of gravity

The wind will stay close to the ground if the pressure gradient force is balanced by gravity.

The Coriolis force

The rotation of the Earth causes an *apparent* deflection of wind direction. (The wind is actually blowing straight but its path when plotted on a map is a curve.)

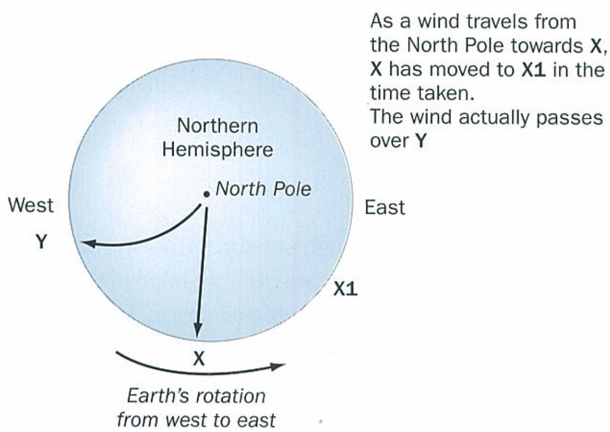


Fig. 2.27 The effect of the Coriolis force on wind direction

This effect is summarised by **Ferrel's Law,** which states that any moving body in the northern hemisphere will be deflected to its right and any moving body in the southern hemisphere will be deflected to its left, as shown on Fig. 2.27.

Fig. 2.28 shows the situation on an isobar map for the northern hemisphere. The wind cannot go on constantly curving because air can't flow from low to high pressure so, once it has started to flow parallel to the isobars, the effect of the pressure gradient force is balanced by the Coriolis force and the wind flows parallel to the isobars (provided there is no surface friction). It is then known as a **geostrophic wind**. Winds in the upper troposphere, such as the upper westerlies, are geostrophic.

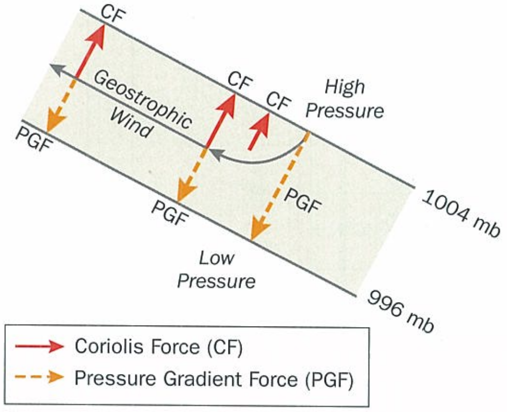


Fig. 2.28 Geostrophic wind

The force of friction

Winds near the surface lose energy and speed because of friction. This reduces the geostrophic force, so the pressure gradient force is no longer balanced by the Coriolis force. Friction causes the wind to blow across the isobars at an angle towards low pressure.

The global pattern of pressure and winds

The tri-cellular model shown in Fig. 2.29 demonstrates how warm air from the tropics could be transferred by an indirect route to add warmth to areas of insolation deficit, whilst these polar areas in turn send colder air back to replace the air from the tropics.

The Hadley cells

These low latitude circulations of air between the Equator and 30° of latitude in each hemisphere are the direct result of thermal differences. Insolation is intense at the warmest part of the Earth's surface (the thermal Equator). Air in contact with the hot land is warmed and rises, creating the permanent equatorial low pressure belt (doldrums). **Trade winds** are drawn into this low pressure belt and meet at the ITCZ (inter-tropical convergence zone). Beneath this zone of rising air, surface winds are light and variable.

Moist air, warmed in the ITCZ, rises in strong convection currents. As the air cools and the water vapour condenses, a lot of latent heat is released which is converted into potential energy and transferred in upper troposphere winds towards the poles. As the air travels towards the poles,

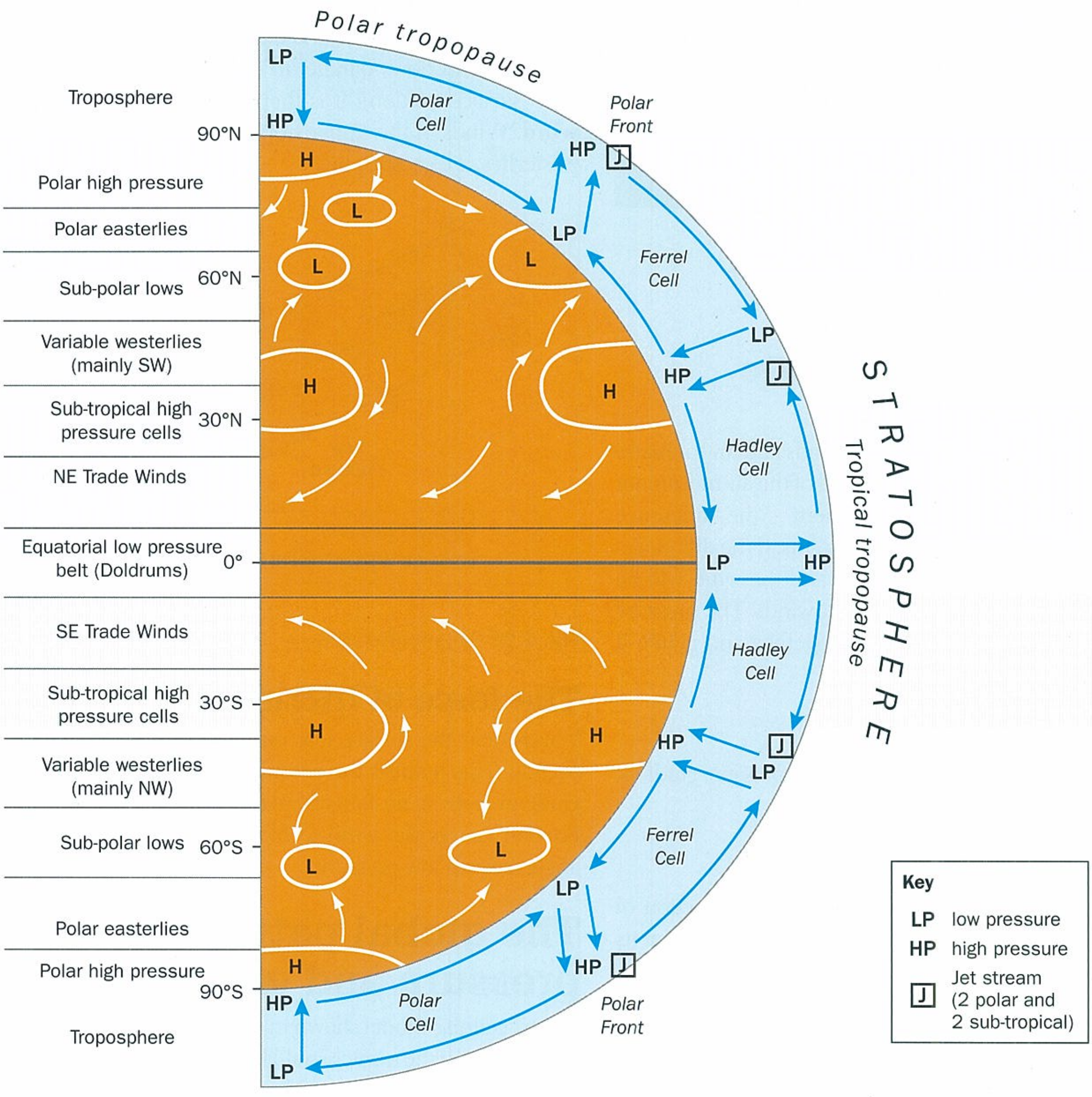


Fig. 2.29 The tri-cellular model of the Earth's atmospheric circulation

it is increasingly deflected to the right or left depending on which hemisphere it is in until, by the time it reaches 30 °N and S, it is strongly under the influence of the Coriolis force, and no longer travels towards the poles much. The air accumulates in the upper troposphere and moves eastwards as the sub-tropical jet streams. Air subsides at 30 °N and S beneath the sub-tropical jet streams, causing sub-tropical high pressure cells at the surface. The trade winds blow down the pressure gradient between these sub-tropical high pressure cells and the equatorial low pressure belt to complete the Hadley cell.

The polar cells

These are also directly thermally induced. Polar air, chilled over the ice caps, subsides to produce high pressure at the surface and then moves into temperate latitudes. The air moving away from the polar high pressure is moving into areas of increasingly wider space because of the shape of the Earth, so it spreads out to occupy the greater space. This reduces its pressure, causing low pressure belts at 50° to 60° north and south. In theory, some of the warmed air rises at these latitudes along the polar fronts and moves towards the poles to be chilled and sink over the polar ice caps to complete the cell. These theoretical upper-air south-easterlies do not occur; instead upper westerlies circulate in high latitudes.

The Ferrel cells

These cells are not directly thermally induced but are consequences of the adjacent thermally induced cells. Some of the air that sinks in the sub-tropical high pressure belts moves towards the sub-polar low pressure belts where it meets the colder air from the polar cell moving towards the equator. The surface between these two air masses is known as the **polar front** and is the boundary between the polar and Ferrel cells. Here the warmer air rises up the frontal surface to the tropopause where, according to the model, it moves back towards the Equator to complete the cell by sinking again in the sub-tropical high pressure zones.

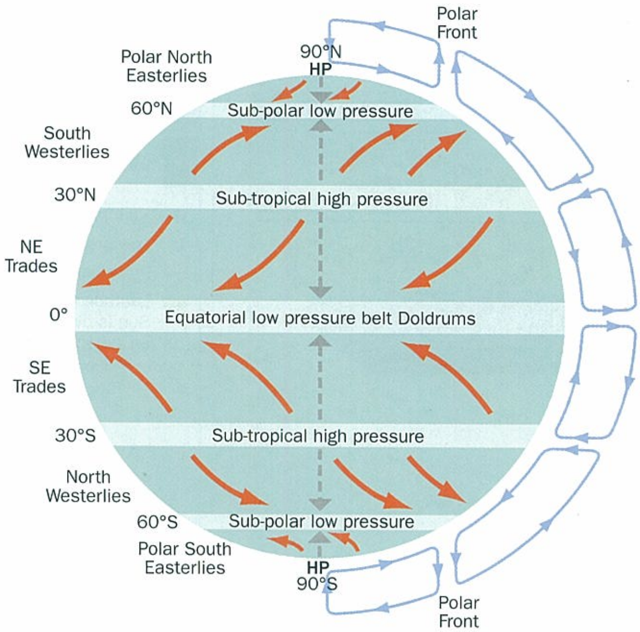


Fig. 2.30 Model of pressure belts and surface planetary winds at the Equinoxes

Fig. 2.30 shows the winds that would result from a tri-cellular model if the only other influences on them were the pressure gradient and Coriolis forces, with the winds moving from high to low pressure and being deflected to the right in the northern hemisphere and to the left in the southern hemisphere. It shows the situation at the Equinoxes when the sun is overhead at the Equator. Actual pressure and wind belts show several deviations from this model. They move further south in the southern hemisphere summer and further north in the northern hemisphere summer.

16. To what extent can you recognise the planetary wind belts, shown in the model on Fig. 2.30, on the map of the actual winds that blow over the Earth in Fig. 2.31?

17. Identify two major pressure systems in January that do not match the tri-cellular model, even after allowing for a slight southern shift expected then. How do they influence the actual wind pattern to make it differ from the model of planetary winds?

18. (a) Explain the thermal low pressure over northern India, shown on Fig. 2.32.

(b) Explain why there are no thermal high pressures over the southern hemisphere continents, although it is the winter season there in July.

Surface winds are only part of the general circulation of the atmosphere; upper air movements are also involved in the transfer of energy round the world.

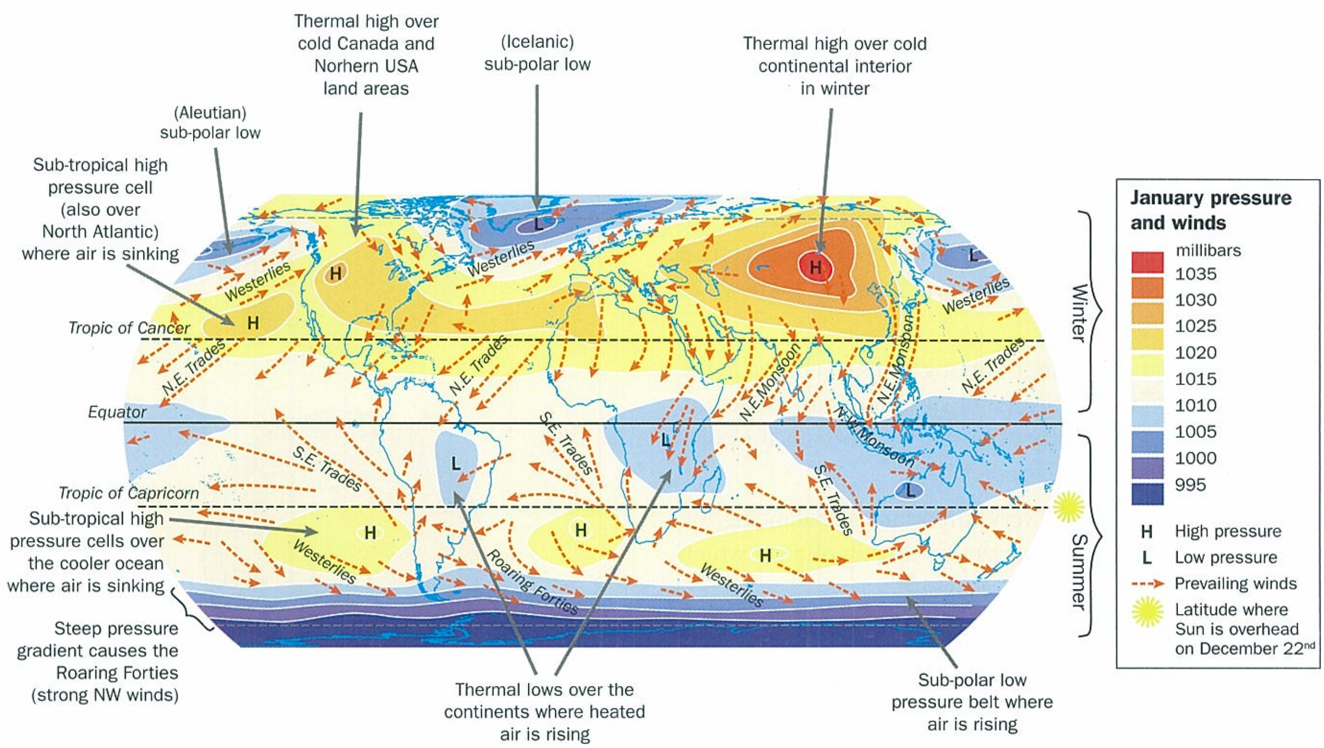


Fig. 2.31 Isobar map of surface pressure belts and their influence on winds in January

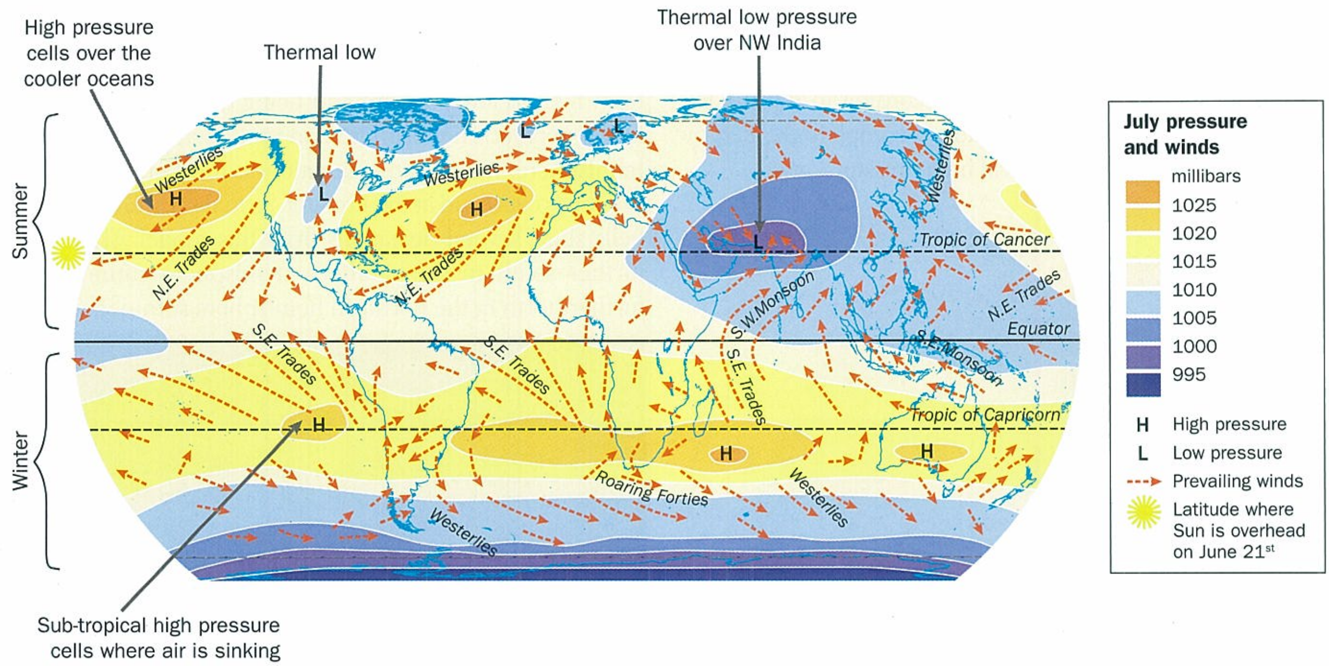


Fig. 2.32 Isobar map of surface pressure belts and their influence on winds in July

The upper westerlies and Rossby waves

These are fast moving westerly winds at a high level between latitudes 30° and 50° that result from:

- a very strong north to south temperature gradient which causes a very strong pressure gradient in temperate latitudes
- the Coriolis force increasing as the air flows towards the poles, causing the air to take a path towards the east and become geostrophic.

Sometimes the upper westerlies only deviate a little from a west to east path but they can have three to six waves in each hemisphere which move slowly from west to east. These are known as **Rossby waves** and are much slower-moving than the air flowing through them which, at its fastest, is known as a jet stream.

When they meander in large curving paths it can lead to the separation of 'pools' of warmer air surrounded by colder air, or vice versa. In this way they transfer heat towards the poles and cooler air towards the tropics.

The upper westerlies are very important in balancing the Earth's energy budget by horizontal mixing of air - which the tri-cellular model did not recognise.

Rossby waves may be caused by a disturbance in the airflow. When the upper westerlies cross a very high mountain range, such as the Rocky Mountains or the Tibetan Plateau, the vertical column of eastwards flowing air is compressed to cross the high ground. As a result it is thrown frequently into wave troughs over north-eastern North America and eastern

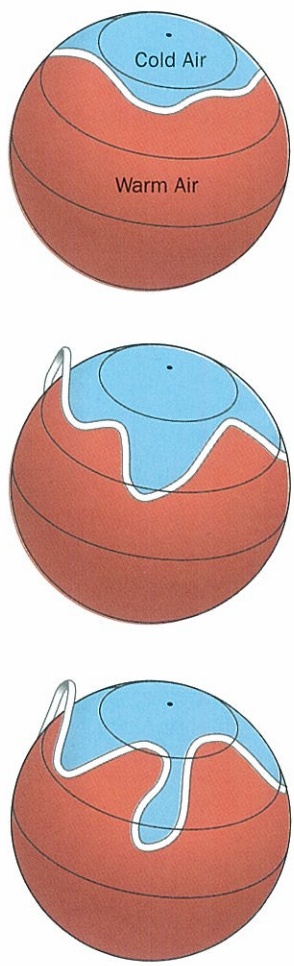


Fig. 2.33 The growth of a wave trough in the upper westerlies of the northern hemisphere until just before a pool of cold air is cut off to be mixed with warmer air. The growth of a ridge towards the north can cut off a pool of warm air in the same way

Siberia. This is especially so in winter when the cold of the continental interiors intensifies them, whereas ridges form over the warmer oceans.

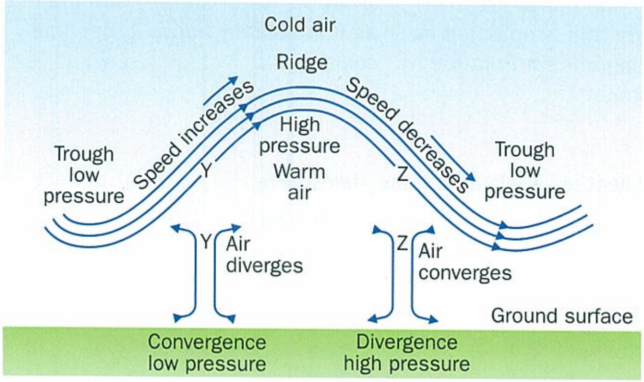


Fig. 2.34 The relationship between Rossby waves and pressure systems at the ground surface. The diagram combines a plan of a Rossby wave and vertical sections of airflow at points Y and Z

Ridges and troughs in the upper westerlies are closely linked to air movements and pressures at the surface. When there is a high pressure ridge in the upper airflow there is a low pressure trough at the surface and vice versa. This is because as the air flows towards the poles, it speeds up and divergence occurs. Then, as the air turns to move towards the Equator it slows, the air piles up, and convergence occurs. This leads to convergence and divergence at the Earth's surface which, in turn, results in low pressure and high pressure respectively at the surface.

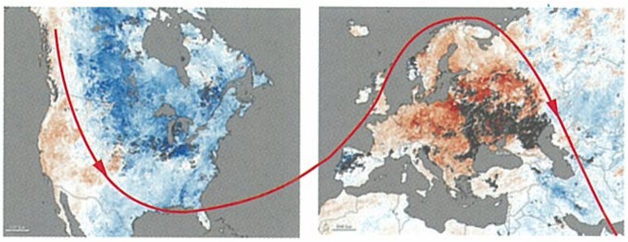


Fig. 2.35 Rossby waves superimposed on to satellite maps of resulting temperature anomalies for January 1–7 2014 in North America and Europe

The map images show how Rossby waves caused abnormal temperatures at the same time in different continents. Areas in blue had negative temperature anomalies (below the average temperatures), with the deepest blue-18 °C below average. Areas in red had positive anomalies, with the deepest red 18 °C above normal. A giant meander in the upper westerlies caused a trough to loop down over North America, allowing very cold air from the Arctic to reach much further south than usual. Temperatures were well below zero as a result. Meanwhile, the Rossby wave continued to form another giant meander creating a ridge over Europe. This brought a winter 'heat wave' with abnormally high temperatures of about 10 °C, as warm air from the south was able to move north. As a result, Russia had to provide artificial snow for the Sochi Winter Olympics, where it would normally have been well below zero in January.

Alaska experienced record-breaking warmth (some places were 22 °C above normal), as it was under the ridge of high pressure over the Pacific coast on the other side of the trough that brought the cold to continental USA. Between the trough over the USA and the ridge over Europe, the speeding limb of the ridge brought storms, heavy rain and intense flooding to western parts of the UK, which had its wettest January since records began.

Fig. 2.36 Rossby waves brought storms to the western UK in January 2014. The south-west airstream was so moist and unstable that only a small amount of orographic uplift was needed to trigger heavy cumulonimbus cloud and torrential rain

As there has not been a change in the amount of atmospheric energy circulating in the world, these weather extremes cannot be explained in that way. Research suggests that the unusual pattern of the Rossby waves in winter 2013-14 was driven by abnormally warm sea surface temperatures in the western tropical Pacific Ocean, extending along the west coast of North America into the north Pacific.

Jet streams

A jet stream is a narrow ribbon of very fast-moving air that runs through the centre of the Rossby waves. Whereas the upper westerlies travel at 50-100 km/h, the jet streams often reach speeds of 250 km/h. They are discontinuous but can be thousands of kilometres long and meander from west to east. There are two in the upper westerlies - one in each hemisphere. Jet streams are often seasonal.

Jet streams form at a high level at the polar front, at the meeting of very cold polar air with warm tropical air, where there are the greatest differences in temperature and pressure in a narrow horizontal distance. The jet stream is located in the warm air about one kilometre below the tropopause. The polar front jet streams are fastest and most frequent in winter when the temperature differences are most marked. They are very variable and can extend latitudinally from 35°

Changes that result from an increase in temperature and cause absorption and storage of heat from the atmosphere			
Change	**Description**	**Reason**	**Consequence**
Melting	Solid to liquid (ice to water).	All involve an increase in the speed of the molecules. The energy for this is obtained by absorption of heat from the atmosphere.	The immediate surroundings (both the water surface and the air immediately above it) become cooler.
Evaporation	Liquid to gas (water to water vapour).		
Sublimation	Solid to gas (ice to water vapour).		

Table 2.8 Phase changes taking heat from the atmosphere

Changes that result from a decrease in temperature and cause latent heat to be released to the atmosphere			
Change	**Description**	**Reason**	**Consequence**
Condensation	Gas to liquid (water vapour to water).	All involve a decrease in the speed of the molecules, so less energy is required, therefore latent heat is released.	The released heat warms the surroundings. This release of heat is very important in providing energy for depressions and storms.
Freezing	Liquid to solid (water to ice crystals).		
Deposition	Gas to solid (water vapour to ice).		

Table 2.9 Phase changes adding heat to the atmosphere

to 70° in each hemisphere. They are at the lowest latitudes in the winter.

Sub-tropical jet streams occur at about 25°N and S where the Hadley cell and mid-latitude circulations meet. They are also westerly flows (moving from west to east) but do not meander as much as the sub-polar jet streams.

Weather processes and phenomena

Atmospheric moisture processes

Atmospheric moisture exists in three **phases** or states:

- *gas* – water vapour, an invisible gas, is the stable phase of moisture when temperatures are above 100 °C, but it can exist at temperatures down to well below freezing point.
- *liquid* – water is the stable phase between 0 °C and 100 °C but exists as super-heated water above 100 °C and as super-cooled water down to –40 °C (it then freezes as soon as it touches ice).
- *solid* – ice is the stable phase of moisture at temperatures below 0 °C.

Phase changes can occur in two ways: a change of temperature and a change in the amount of water vapour in the air.

19. Describe and explain the phase change occurring in Fig. 2.37 and the effect this has on the air temperature.

Humidity

Humidity refers to how moist the air is because of the water vapour it contains. **Absolute humidity** is the *actual* amount of water vapour in a given volume of air.

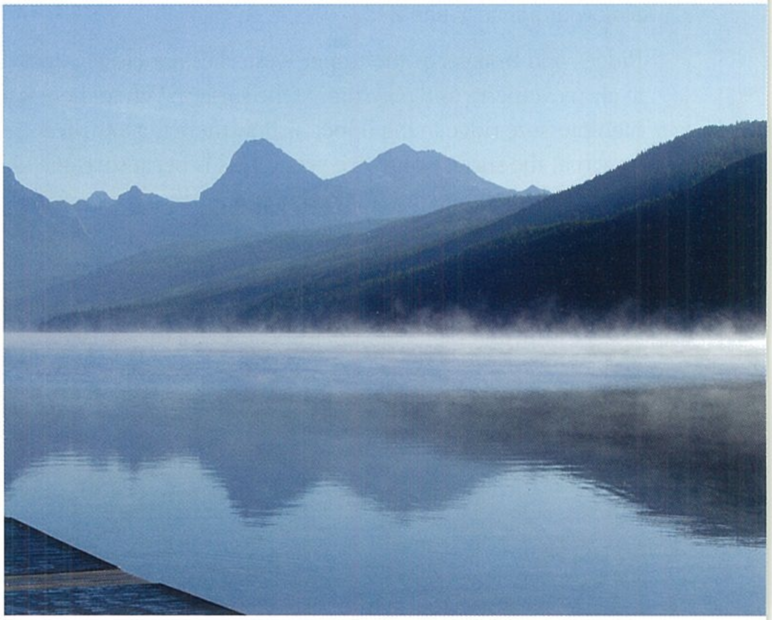


Fig. 2.37 A phase change occurring in early morning in Glacier National Park, USA

Relative humidity is more useful, as it measures how near the air is to saturation. It indicates how much water vapour the air is holding (i.e. the absolute humidity) compared with the *maximum* amount that it could hold *at that temperature and pressure*.

$$\text{Relative humidity \%} = \frac{\text{Actual moisture content} \times 100}{\text{The saturation moisture content at the same temperature and pressure}}$$

Air is saturated when it has 100 per cent relative humidity. Warm air can hold more moisture than cold air.

Evaporation

Water changes to gas when it is heated and the air is unsaturated. Rates of evaporation increase when temperatures rise and the air is very dry, conditions are calm and there is a water source available.