2 Atmosphere and weather

In this chapter you will learn:

→ How energy from the sun is gained, lost and transferred in the earth-atmosphere system at a local scale and how it varies from day to night.

→ About variations in the global energy budget, how energy is transferred from areas of surplus to areas of deficit and how it is linked to seasonal variations in temperature, pressure and wind belts.

→ How atmospheric moisture processes cause different types of precipitation.

→ How human activity is having an impact on weather and climate at both global and city scales.

The atmosphere

The **atmosphere** is a mixture of gases held to Earth by gravity; it increases in density, and therefore pressure, towards the Earth's surface and is divided into zones based on temperature variations. Only the lower two zones are relevant to our study.

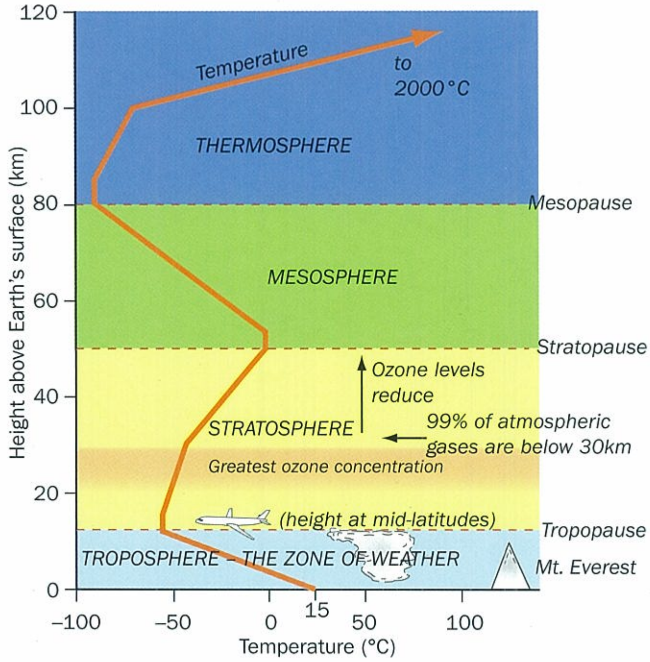


Fig. 2.1 The vertical structure of the atmosphere

Weather results from processes at work in the **troposphere.** At the **tropopause** a **temperature inversion** prevents air rising into the stratosphere. In the troposphere the air normally cools with increased altitude but the air above the tropopause is warmer than the air immediately below it. Cooler air is denser and cannot rise into warmer air. The tropopause varies in height from about 8 km at the poles to 18 km at the Equator.

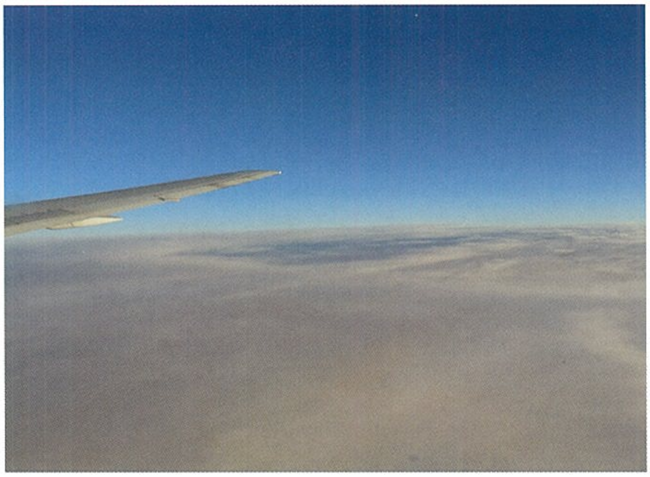


Fig. 2.2 Flat upper surfaces of cloud at the tropopause indicate the temperature inversion

Both troposphere and stratosphere consist of 78 per cent nitrogen and 20 per cent oxygen but trace amounts of other gases, such as methane and low-level ozone, also occur in the troposphere, whereas the stratosphere has important concentrations of ozone. Almost all the water vapour and suspended **aerosols** are in the troposphere.

Local diurnal energy budgets

Factors affecting the daytime energy budget

The sun provides the energy source to drive all atmospheric processes. The atmosphere derives little heat from the sun's rays passing through it. Most atmospheric heat is gained from the Earth. The daytime energy budget shown in Fig. 2.3 is a model of the average situation, based on 2013 revised estimates by NASA.

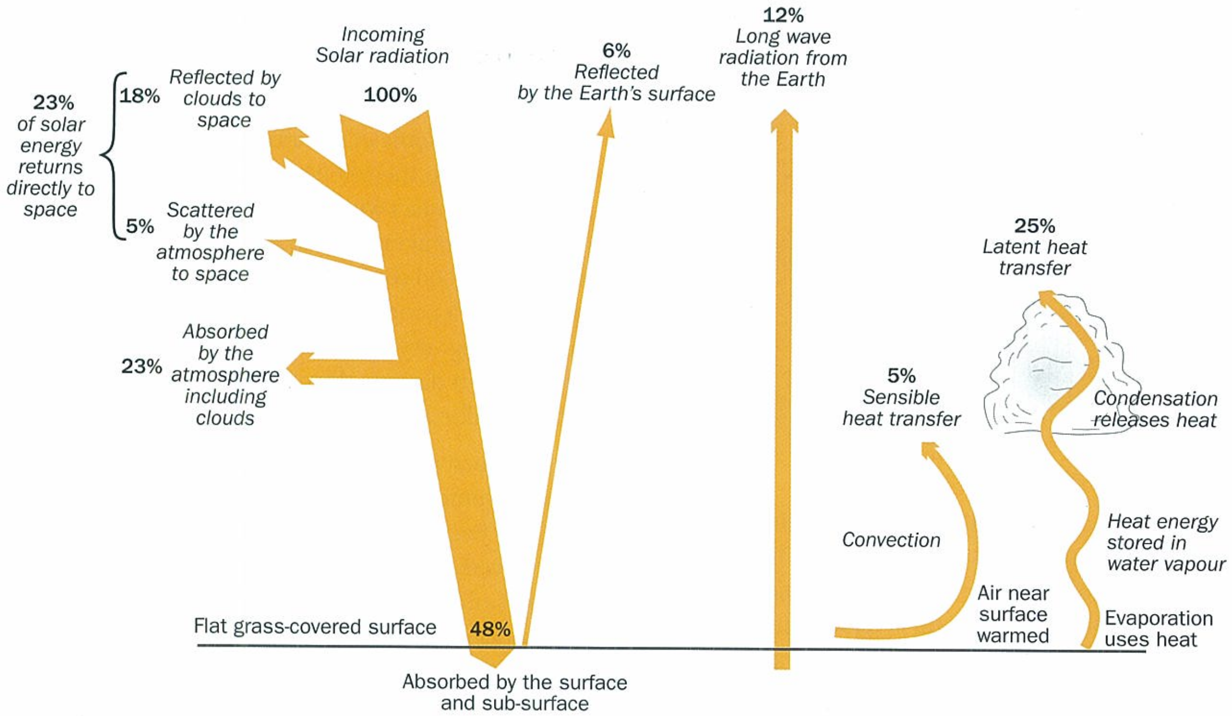


Fig. 2.3 The daytime energy budget

Incoming solar radiation (insolation)

The very hot sun emits short-wave (ultraviolet) **radiation**. Radiation is the transfer of heat from one body to another by electro-magnetic waves. At any time, the half of the Earth facing the sun is in daylight and being heated while the other side remains unheated.

Not all the energy emitted by the sun reaches the Earth's surface. During its passage through the atmosphere it is estimated that:

- 5 per cent is **scattered** straight back to space by dust and smoke particles.
- 24 per cent is **reflected** back to space, 18 per cent by the white upper surface of clouds and the water droplets within them and 6 per cent by the Earth's surface, mainly by snow, ice and water surfaces.
- 23 per cent is **absorbed** by atmospheric gases, mainly by ozone and oxygen at high levels, with small amounts by carbon dioxide and water vapour near the Earth's surface.

The remaining 48 per cent reaches the Earth's surface directly and heats it. The intensity of heating by incoming solar radiation depends on the angle of the sun's rays, being greatest where they reach the surface at 90° and reducing as their angle becomes smaller.

Dust and smoke particles also scatter another 5 per cent of solar radiation within the atmosphere. The short-wave bluer light rays are more easily scattered than the longer-wave red rays so, when the sun is low near the horizon, passing through a thicker atmosphere, more scattering occurs and only red rays remain. The thick stratus cloud in Fig. 2.4 limits radiation received at the surface to about 10 per cent.

Fig. 2.4 Red evening sky caused by scattering of the sun's rays

Solar radiation reflected by the Earth's surface

The percentage of solar radiation that is reflected back to space by the Earth's surface is known as its **albedo**. Lighter-coloured surfaces reflect more solar radiation, while darker-coloured surfaces absorb more of it.

Surface	Average albedo (%)
Thick cumulonimbus cloud	92
Fresh snow	80
Thick stratus cloud	65
Sandy surfaces	40
Thin cloud	32
Concrete	22
Deciduous forest	18
Green grass	15
Coniferous forest	12
Asphalt	10
Dark soil	7

Table 2.1 Average albedo values

The albedo of oceans varies remarkably according to the time of day; when the sun is at a high angle near midday it has very low albedo (about 4 per cent) but reflection can reach 80 per cent when the evening sun is very low in the sky.

The total amount of energy lost to space by scattering and reflection, from both Earth and atmosphere, is the **planetary** or **global albedo**.

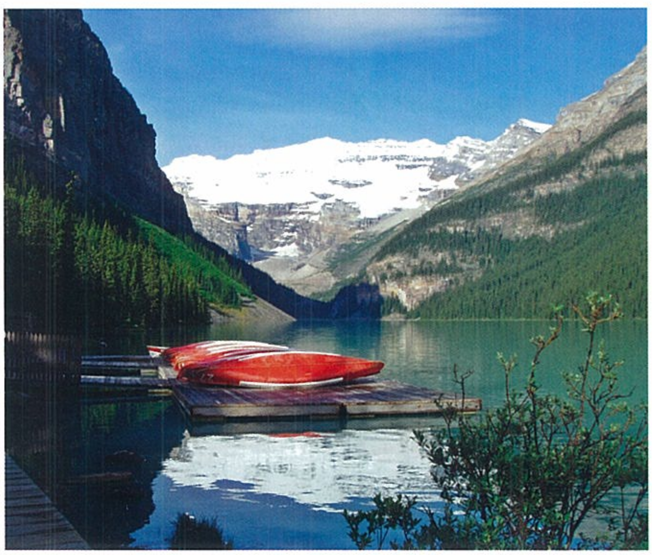


Fig. 2.5 Local variations in albedo in the Canadian Rocky Mountains

1. Rank, from highest to lowest, the variations in albedo of the different surfaces shown in the sunny areas on Fig. 2.5. The sun was at an angle of about 50° above the horizontal.
2. Explain why:
   (a) dirty snow melts faster than fresh snow
   (b) the albedo of crops can vary from 15 to 25 per cent
   (c) some parts of urban areas will have lower than the average albedo for an urban area (15 per cent) and others will have higher albedos. Give examples to illustrate each.

Energy absorbed into the surface and sub-surface

Dark surfaces absorb much more radiation than surfaces with a high albedo. Some of the absorbed energy is transferred a short depth into the soil and rocks by **conduction**. This is achieved by contact with the heated surface in the same way as heat transfers along the handle of a spoon left in a hot liquid, as metal is a good conductor of heat. Light-coloured rock, like limestone, is a poor conductor, so heating is confined to the surface, giving very high rock surface temperatures (45 °C) in hot deserts in daytime. By contrast, darker rock like granite, with a low albedo, absorbs heat well.

The conductivity of soils also varies according to their moisture content. Anyone who has walked barefooted on a dry sandy beach in early afternoon in low latitudes will have experienced great heat on the soles of their feet. The air in the pores of dry sand is a poor conductor, so heat remains concentrated at the surface, whereas water in soil increases heat flow. In a wet sandy soil conduction transfers the heat down and the surface is cooler.

Long-wave Earth radiation

Short-wave radiation from the sun is absorbed by the Earth and re-radiated as long-wave (infra-red) radiation because the Earth is a cool body. This is much more easily absorbed by 'greenhouse' gases in the atmosphere - mainly by water vapour and carbon dioxide - than short-wave radiation, and is the most important way in which the atmosphere is heated. Clouds absorb long-wave radiation very efficiently and continuously re-radiate it back to Earth - keeping heat in by the **greenhouse effect**. Heat loss is greatest in dry air but, in general, only small amounts escape directly to space through '**radiation windows**'.

Sensible heat transfer

Sensible heat transfer occurs when heat energy is transferred by direct conduction or **convection**.

→ Air is a very poor conductor of heat, so only a thin layer next to the surface is warmed by conduction.

→ Warming causes the air molecules to expand, become lighter and rise through air that is cooler and denser. This process of convection transfers heat to higher altitudes and, on very hot summer days, the strongly rising air currents can reach the tropopause. Cooler air moves down to replace the rising air and is, in turn, heated.

→ Warm winds near the surface can be deflected upwards by an obstacle and can reach 600 m above the surface if the wind turbulence is very strong.

Latent heat transfer

Latent heat transfer occurs when water on the Earth's surface **evaporates** to water vapour or ice melts to water vapour.

The heat needed to make these changes is absorbed from the air, leaving less energy for heating at the surface. This latent heat is stored in the water vapour and may be carried upwards in convection currents until it cools sufficiently for the water vapour to condense into water droplets or change into ice crystals. During this change the stored heat is released into the air, warming it. This is known as the **latent heat of condensation** and increases the speed and extent of convection. Much solar radiation is lost by latent heat being used to convert snow and ice back to water in high latitudes in spring and early summer.

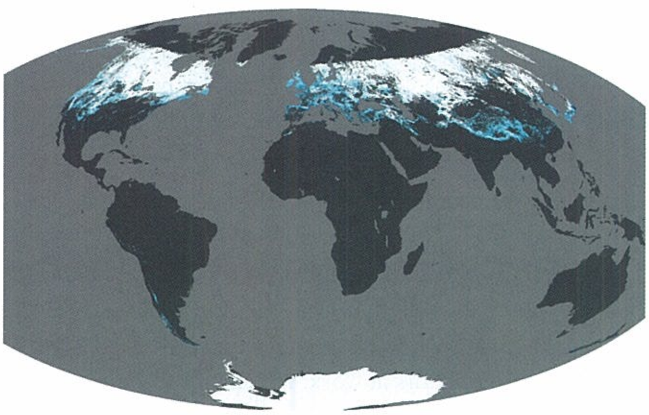


Fig. 2.6 Satellite image showing snow cover in December

Fig. 2.6 does not show the snow and ice in the most northerly latitudes which were in darkness at the time the image was received. Where snow cover is permanent, as in Greenland and Antarctica, its albedo is so great that the net radiation balance is zero or slightly negative, even on summer days when there is maximum insolation during the 24 hours of daylight.

The daytime energy budget has a surplus of energy, as shown in Table 2.2.

3 (a) Describe two types of latent heat transfer that would be occurring when the photograph, Fig. 2.5, was taken.

(b) Describe and explain the atmospheric process occurring in Fig. 2.7.

Fig. 2.7 Ice crystals after sunrise in winter in mid-latitudes

The influence of clouds on the daytime energy budget

The model shown in Fig. 2.3 (page 35) does not include the influences different clouds have on daytime energy transfers.

→ High thin clouds, such as cirrus, allow incoming solar radiation to pass through but absorb some long-wave radiation, so warming the Earth's surface.

→ Deep convective clouds, especially cumulonimbus, neither heat nor cool overall.

→ An overcast sky with complete cloud cover of low, thick clouds, such as stratus and stratocumulus, can reflect 80 per cent of solar radiation and cool the Earth's surface.

→ Clouds usually have higher albedos than the surface below them, so more short-wave radiation is reflected back to space than would be the case if there were no clouds. So, clouds have a net cooling effect.

4. Using a different example for each, describe how, and explain why, a daytime energy budget will vary:

(a) from time to time.

(b) from place to place.

The daytime energy budget			
Input		**Outputs**	
Incoming short-wave solar radiation	minus	Reflected solar radiation + outgoing long-wave terrestrial radiation + energy absorbed into the Earth's surface + sensible heat transfer + latent heat transfer	= Surplus energy available at the surface (variable from place to place and time to time).

Table 2.2 The daytime energy budget

Fig. 2.8 About 80 per cent of solar radiation is reflected back to space from the white upper surfaces of thick clouds

The night-time energy budget

Whereas the daytime energy budget model has six factors, the night-time model lacks two components - it has no short-wave radiation from the sun, so has a deficit of energy, and it has no reflected solar radiation, making it a four factor model. As insolation stops, the ground loses heat and cools and the air next to it also cools. At night the budget is in deficit.

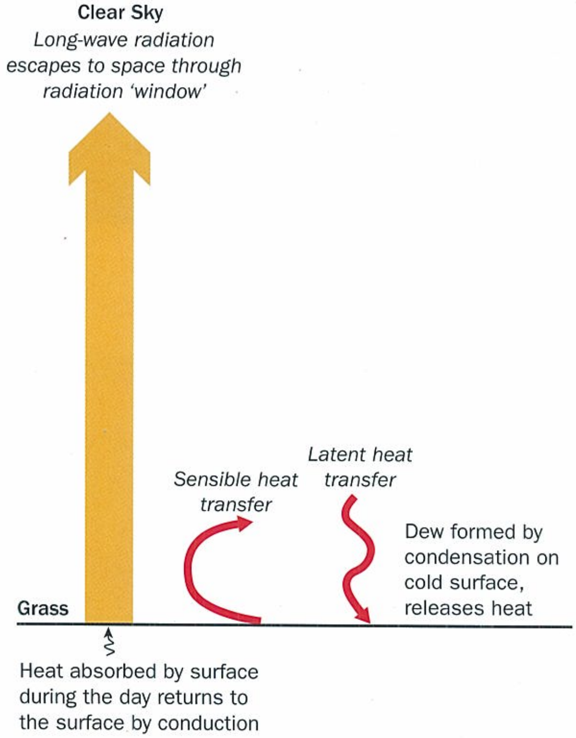


Fig. 2.9a The night-time energy budget

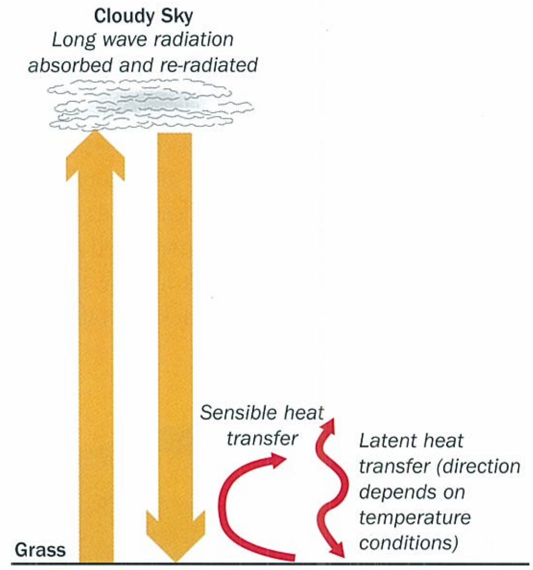


Fig. 2.9b How cloud changes the model

Conduction of heat to the surface

Heat that was absorbed into the soil and rocks during the day, returns to the surface at night and offsets to a small extent the other factors at work, which all cause heat loss.

Long-wave Earth radiation

The amount of long-wave radiation escaping to space from the Earth depends on the cloud cover. Clear skies result in very cold nights. Without cloud to stop the long-wave radiation escaping to space, temperatures fall quickly, leading to large temperature differences between night and day, especially if the daytime was also cloudless.

5. Look at Fig. 2.10 of temperature changes during a cloudless day.

(a) In mid-latitudes how many minutes after dawn is the minimum temperature and how many hours after noon is the maximum temperature of the day? What is the relationship between the incoming solar radiation and the outgoing long-wave radiation at these times?

(b) Describe and explain the two trends in temperature over 24 hours.

Sensible heat transfer

Although convectional uplift may continue after dark in the tropics and sub-tropics, it is unimportant in higher latitudes where air often sinks at night. There may also be some sensible heat transfer by **advection** - horizontal transfers of air from a warmer to a colder area.

Latent heat transfer

On cloudless nights, the Earth's surface rapidly loses heat by long-wave radiation. Cooling is very intense if the air is also

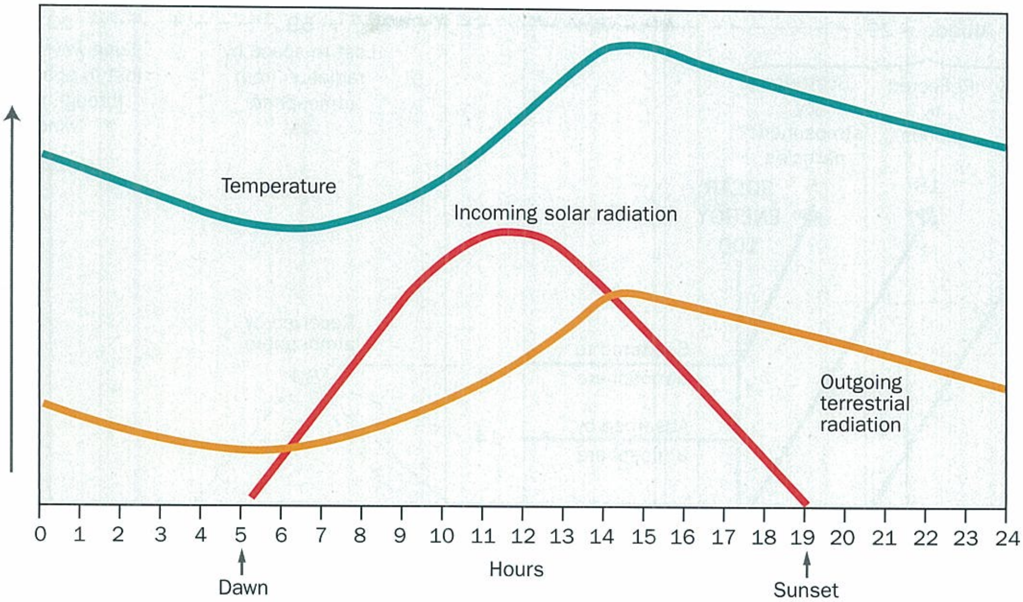


Fig. 2.10 Multiple line graph showing the influence of solar and terrestrial radiation on temperature during a day in mid-latitudes

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

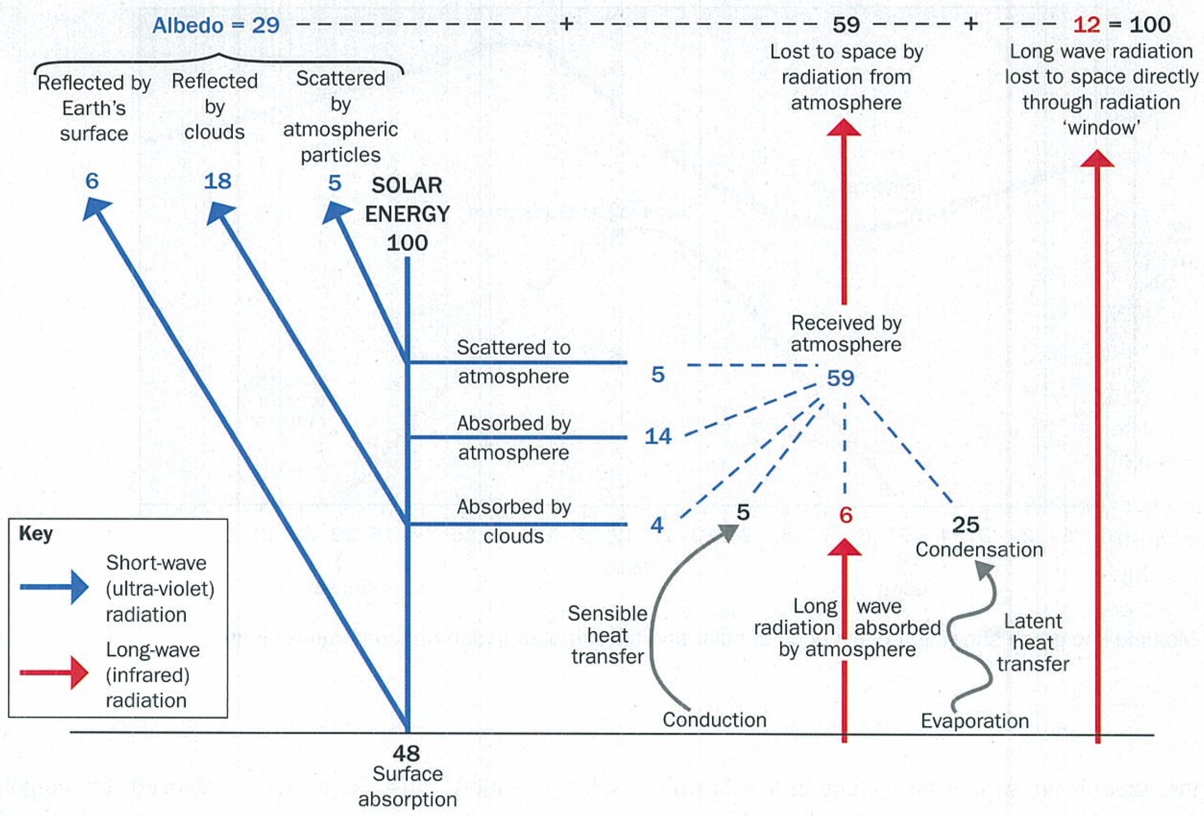


Fig. 2.11 The global energy budget (estimated figures)

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).

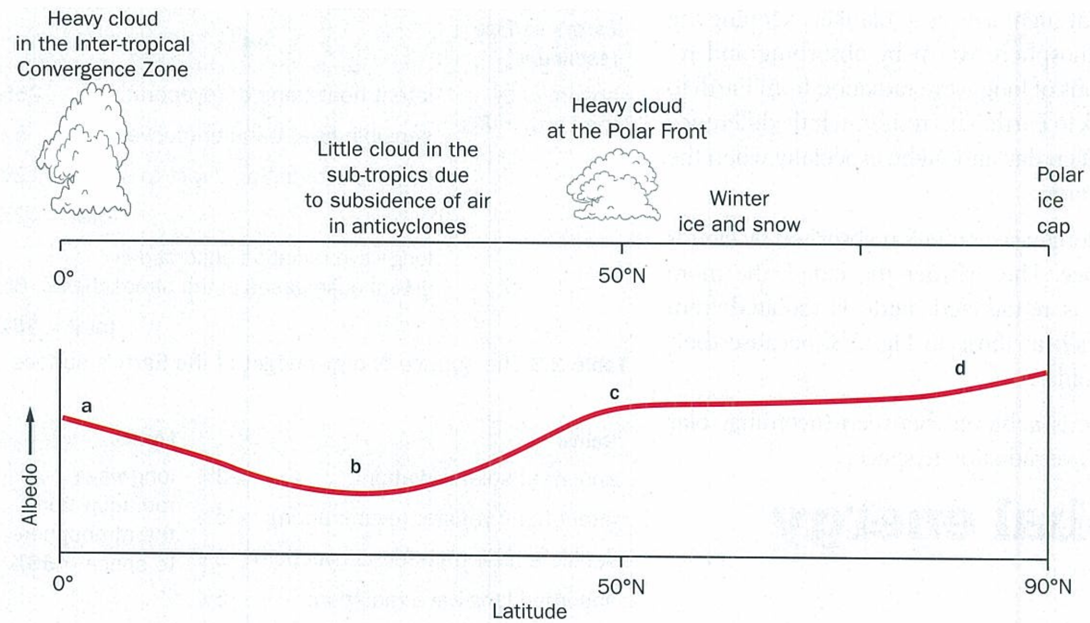


Fig. 2.12 The generalised pattern of albedo in the northern hemisphere

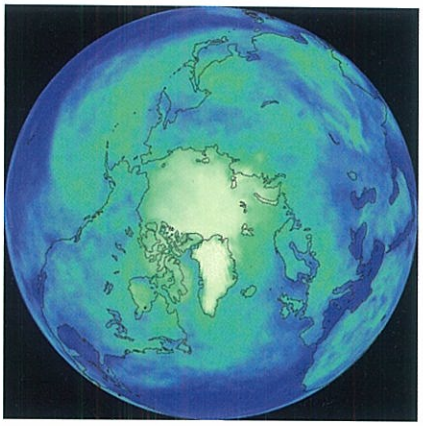


Fig. 2.13 Satellite image of reflected short-wave radiation in the polar region of the northern hemisphere in June 2009

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.

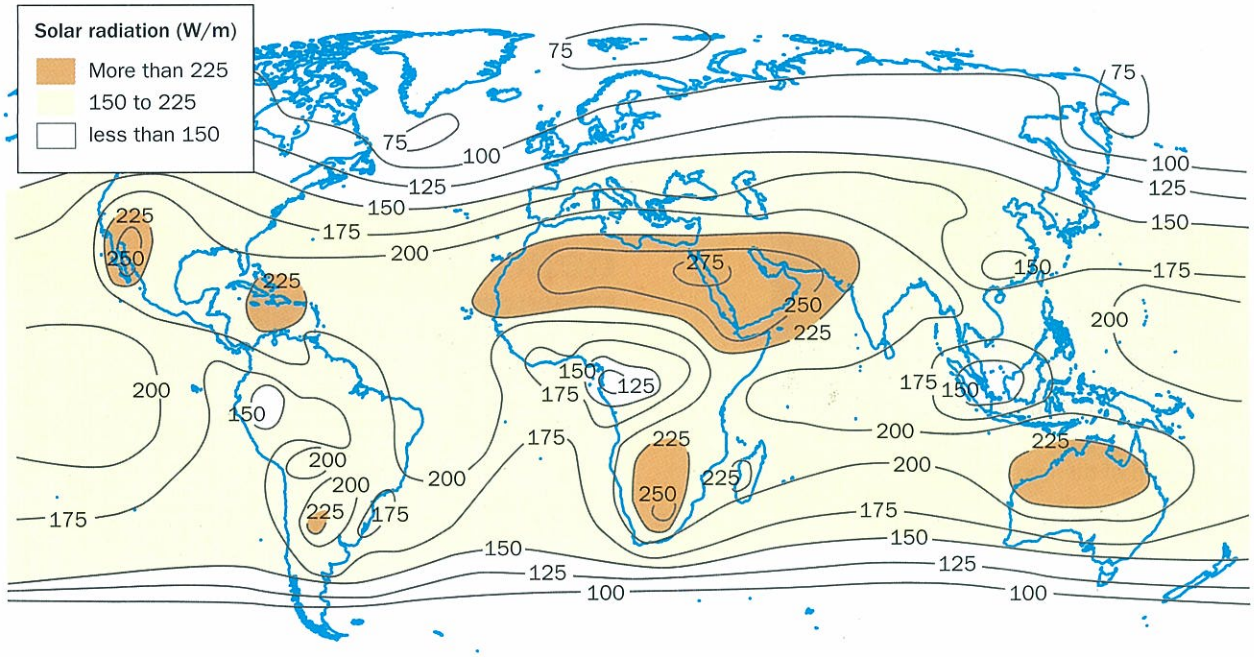


Fig. 2.14 Isoline map of the average annual distribution of solar radiation (W/m²) received at the Earth's surface

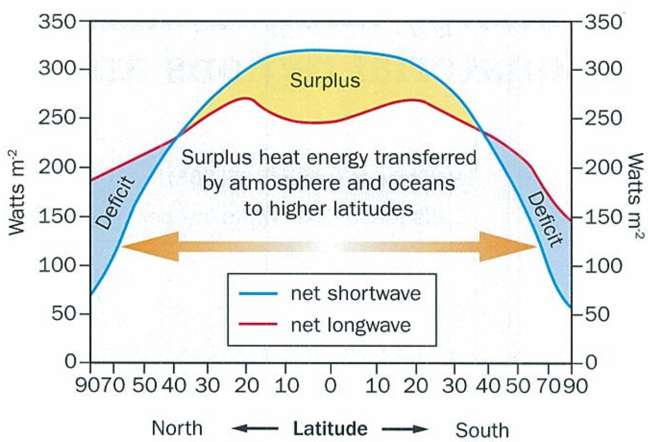


Fig. 2.15 The net radiation balance

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. Sub-tropical 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.

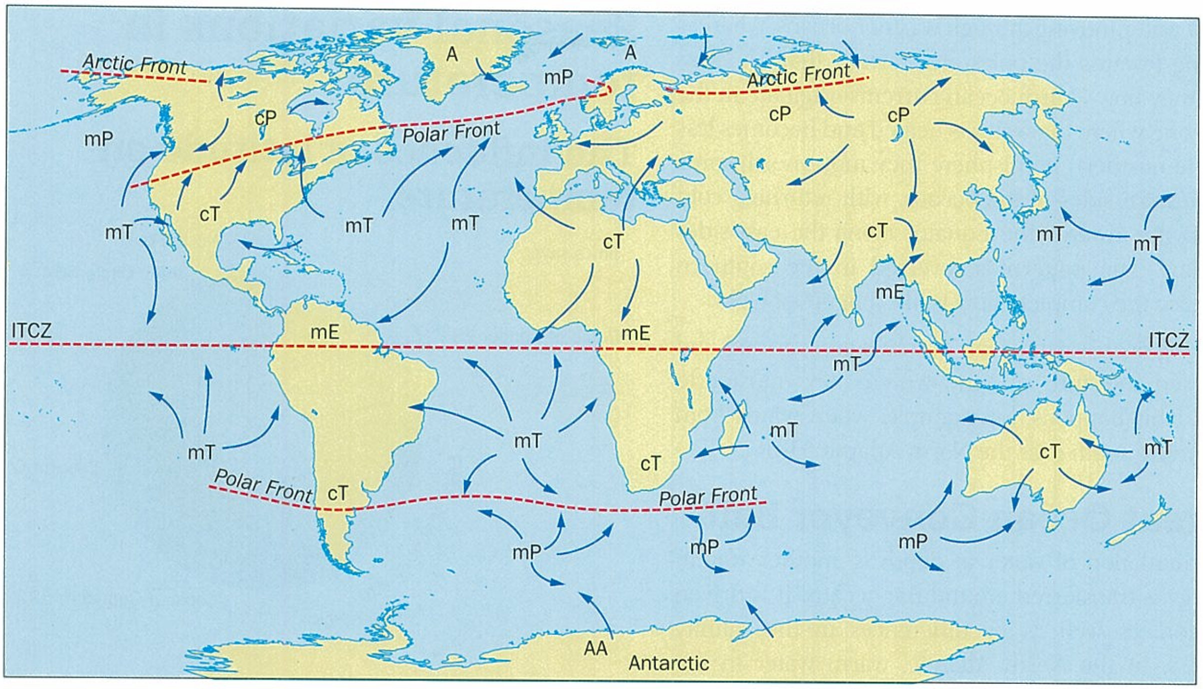
Source regions for polar air masses are high pressure systems over the continents. In winter the air becomes very cold by conduction cooling during contact with the cold land surface and in summer they are relatively cool. North-easterly winds in the northern hemisphere and south-easterly winds in the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

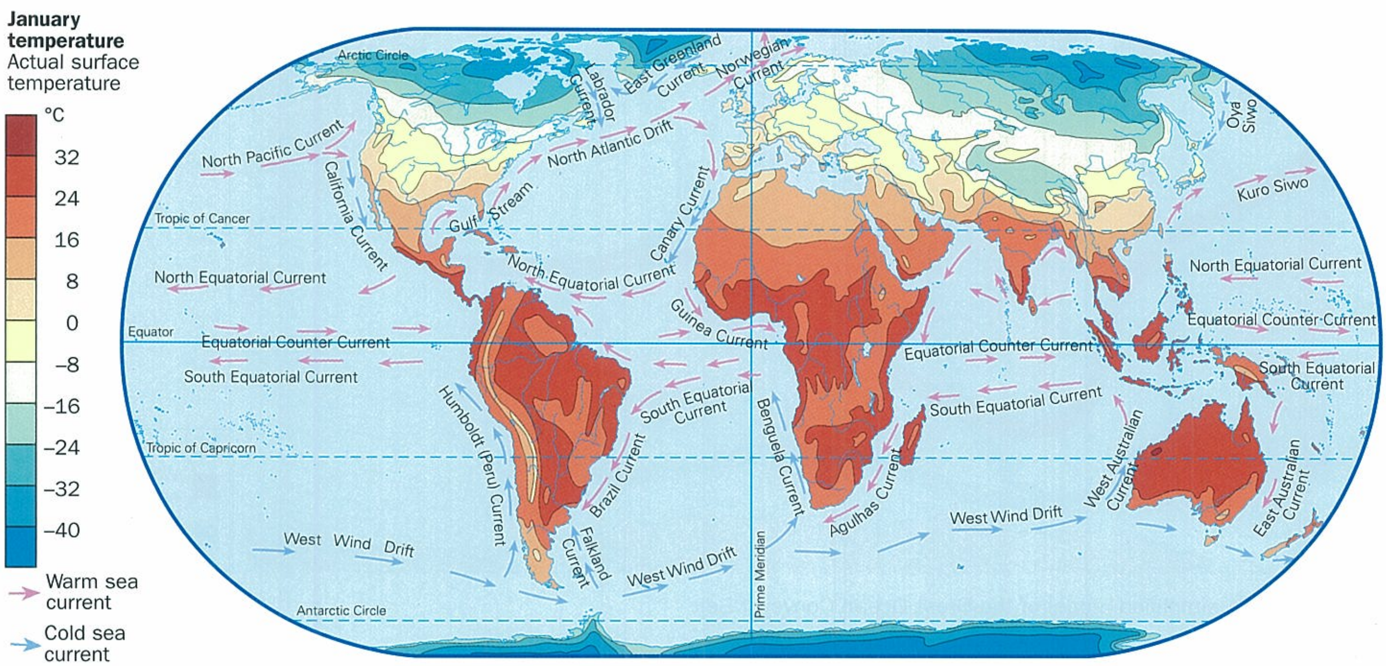


Fig. 2.17 Temperature and ocean currents in January

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

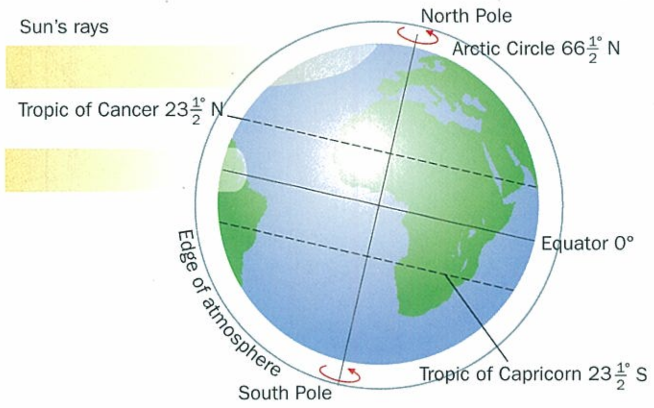


Fig. 2.19 The influence of latitude on temperature

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.

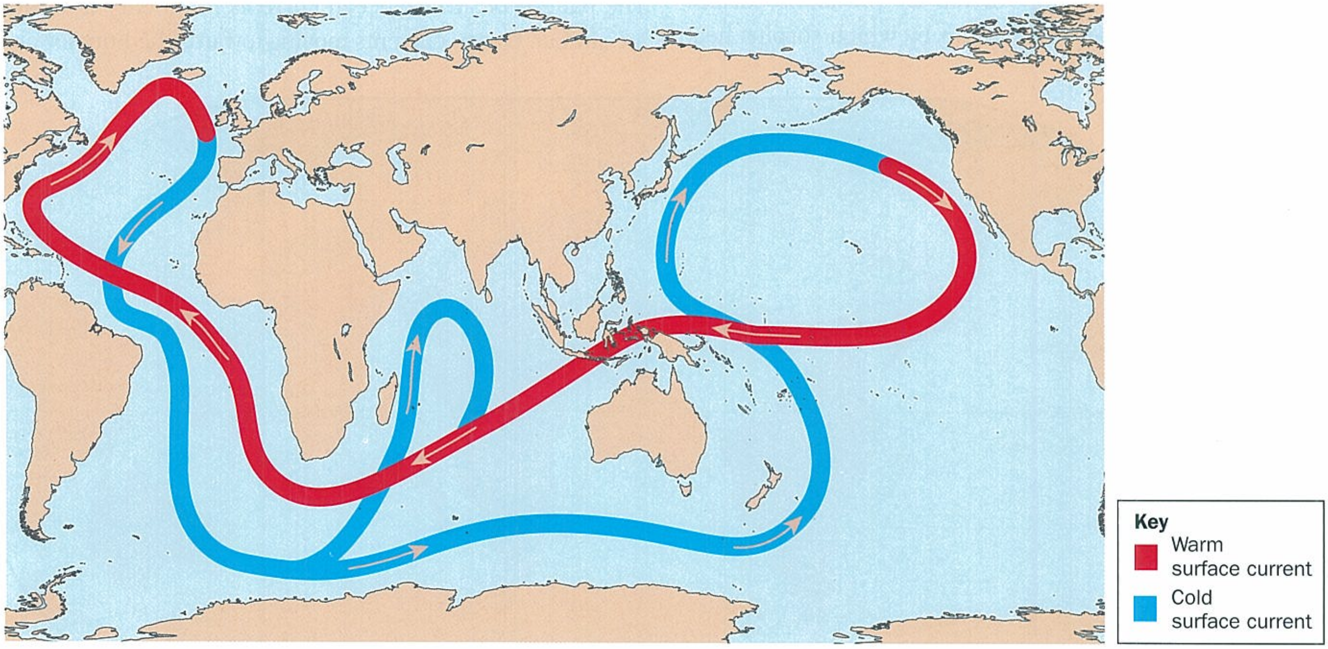


Fig. 2.18 A simplified model of the Great Ocean Conveyor Belt

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

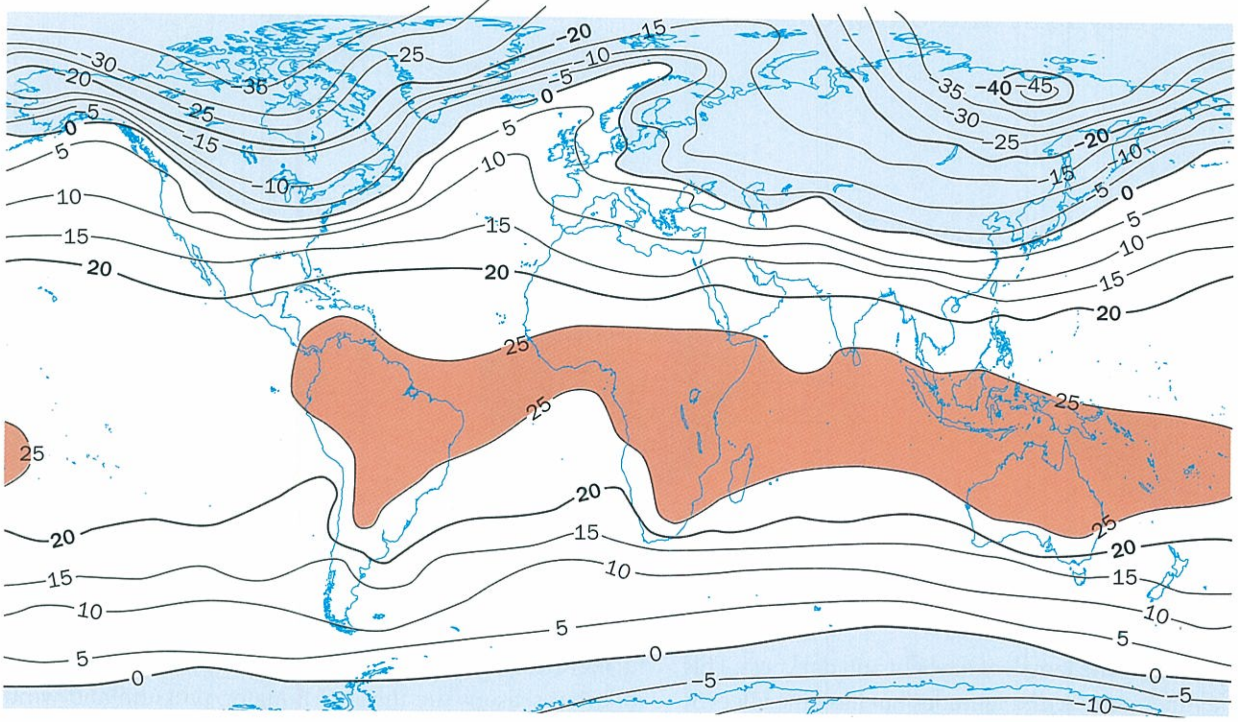


Fig. 2.20 Isotherm map showing the global pattern of temperatures in January

Key to Figs 2.20 and 2.21

- Above 25 °C
- 0–25 °C
- Below 0 °C

Mean Sea Level Temperatures (°C) – July

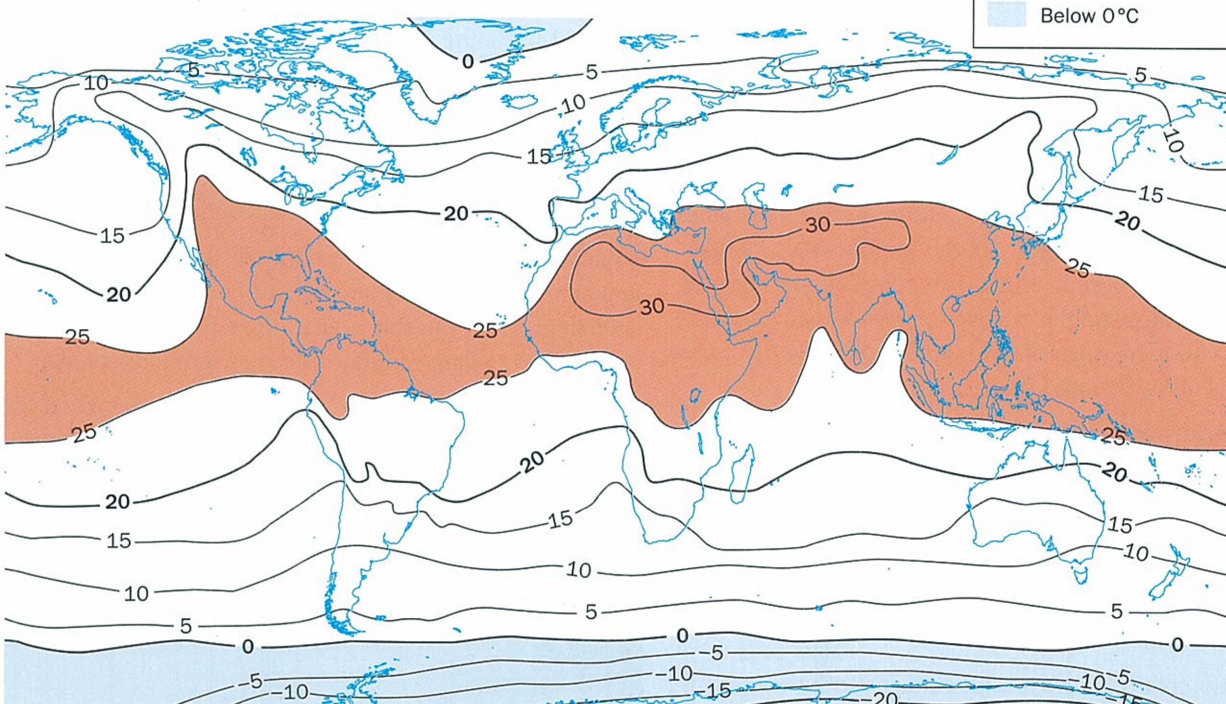
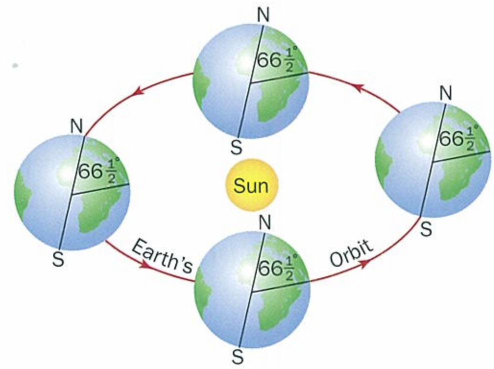


Fig. 2.21 Isotherm map showing the global pattern of temperatures in July

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



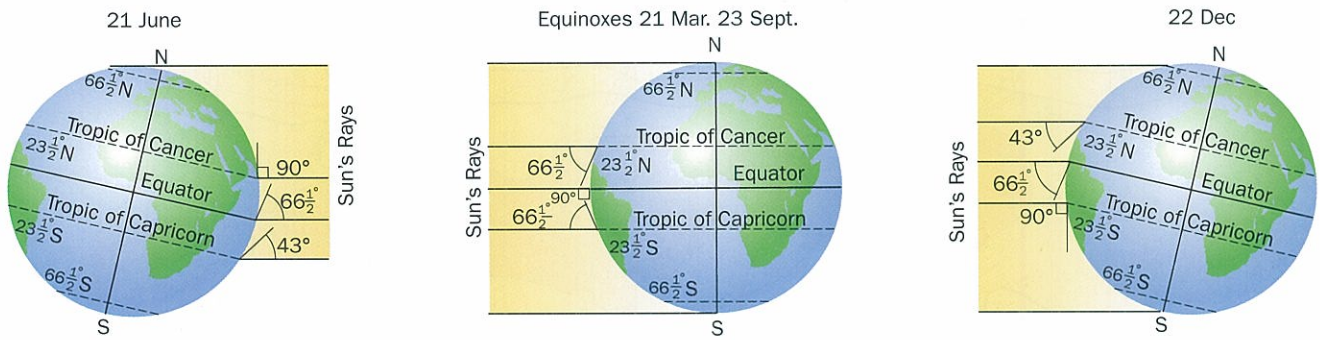


Fig. 2.22 The orbit of the Earth round the sun causes varying angles of the noonday sun on different dates in the year

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 66½° 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.

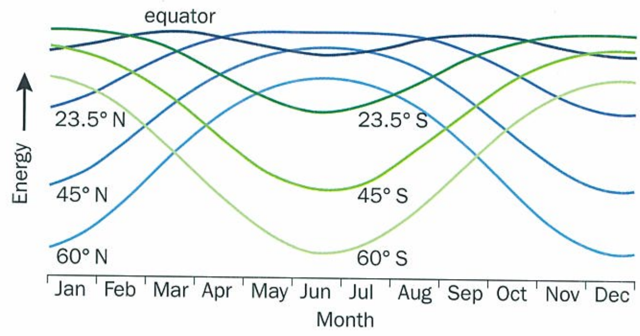


Fig. 2.23 Multiple line graph showing how solar energy received at noon varies with latitude during a year

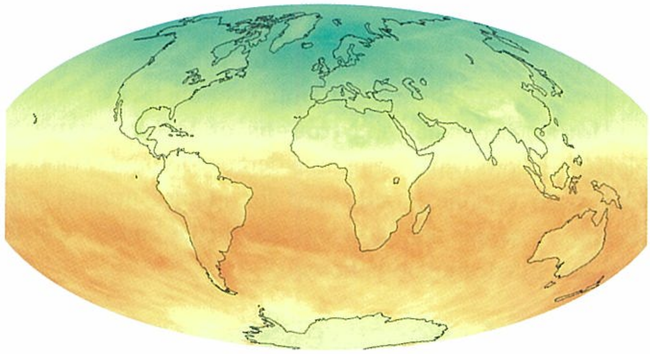


Fig. 2.24a Image of net radiation in January compiled from satellite data

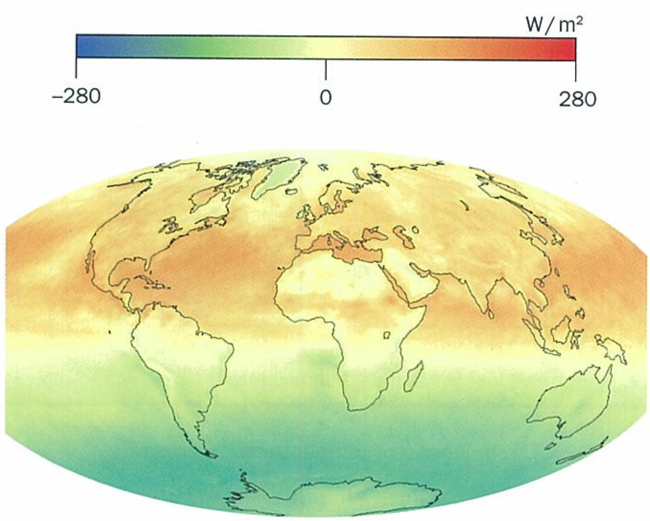


Fig. 2.24b Image of net radiation in July compiled from satellite data

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.

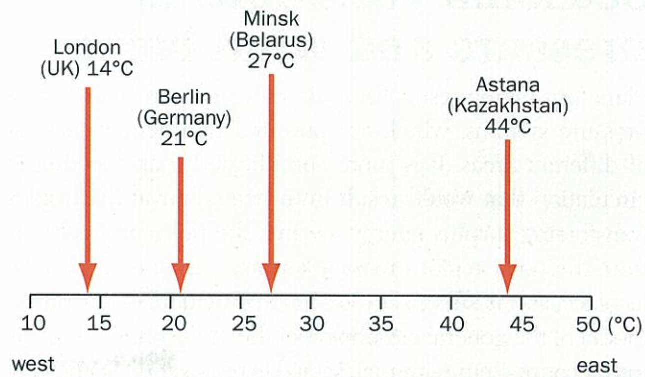


Fig. 2.25 Changes in annual temperature range along (or close to) 52½ °N

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.

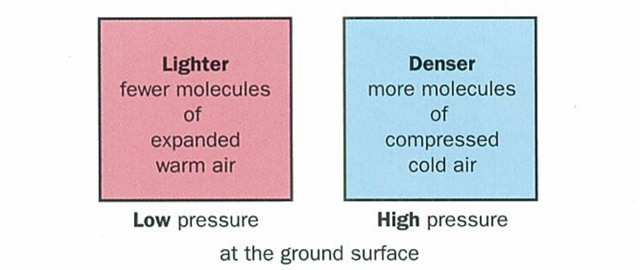


Fig. 2.26 The influence of temperature on pressure

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.)

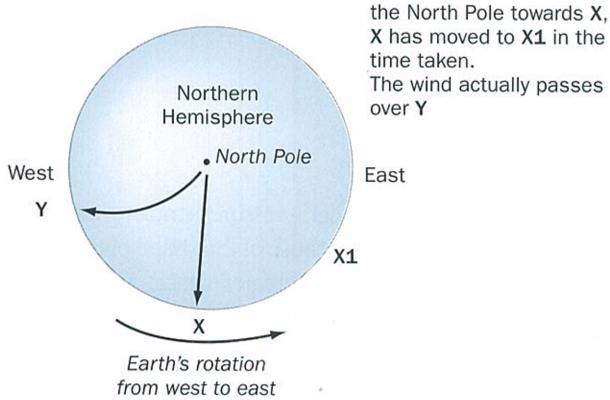


Fig. 2.27 The effect of the Coriolis force on wind direction

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.

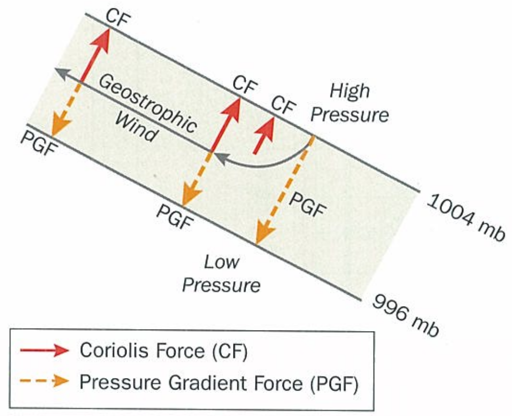


Fig. 2.28 Geostrophic wind

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,

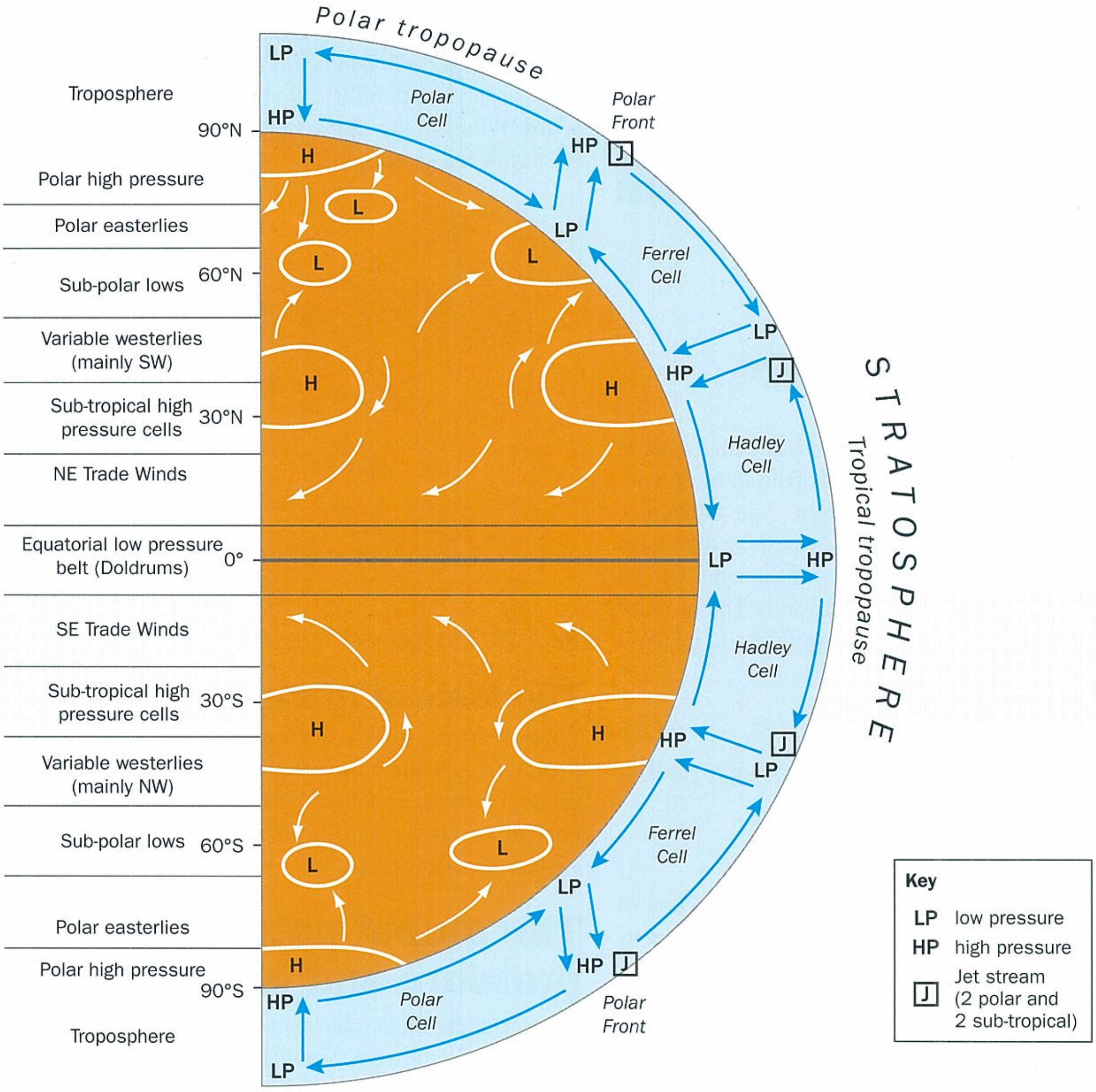


Fig. 2.29 The tri-cellular model of the Earth's atmospheric circulation

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.

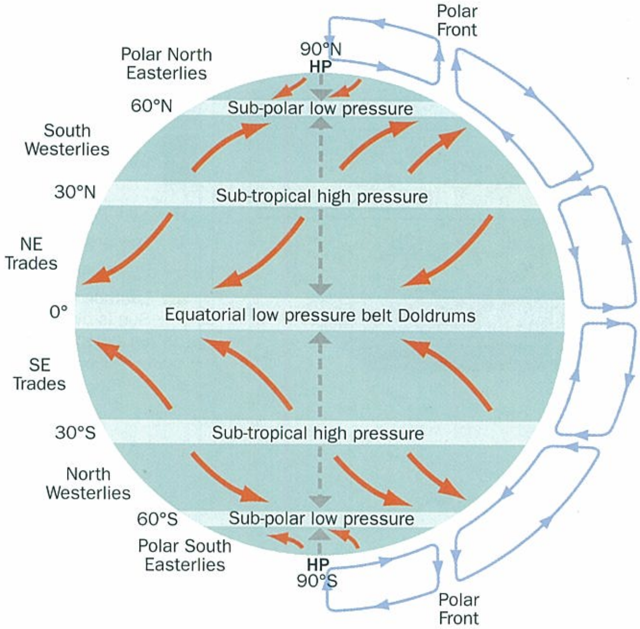


Fig. 2.30 Model of pressure belts and surface planetary winds at the Equinoxes

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.

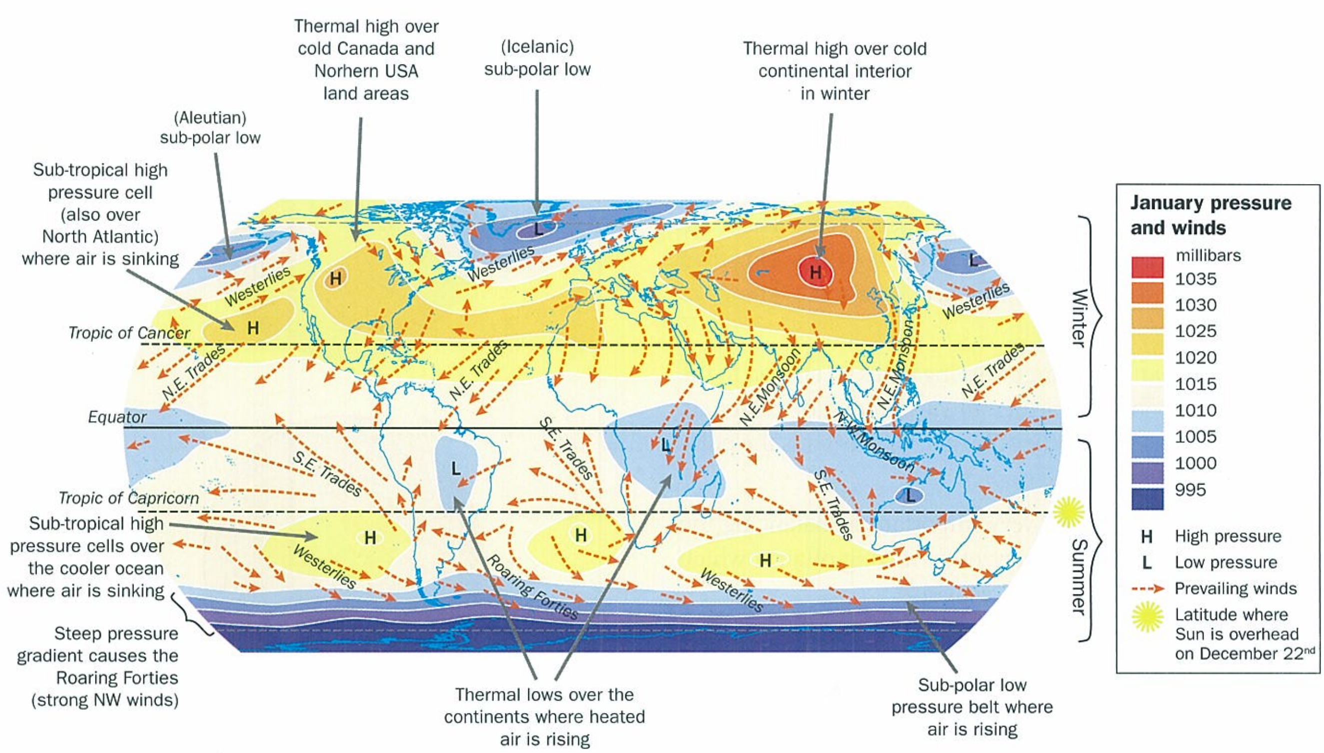


Fig. 2.31 Isobar map of surface pressure belts and their influence on winds in January

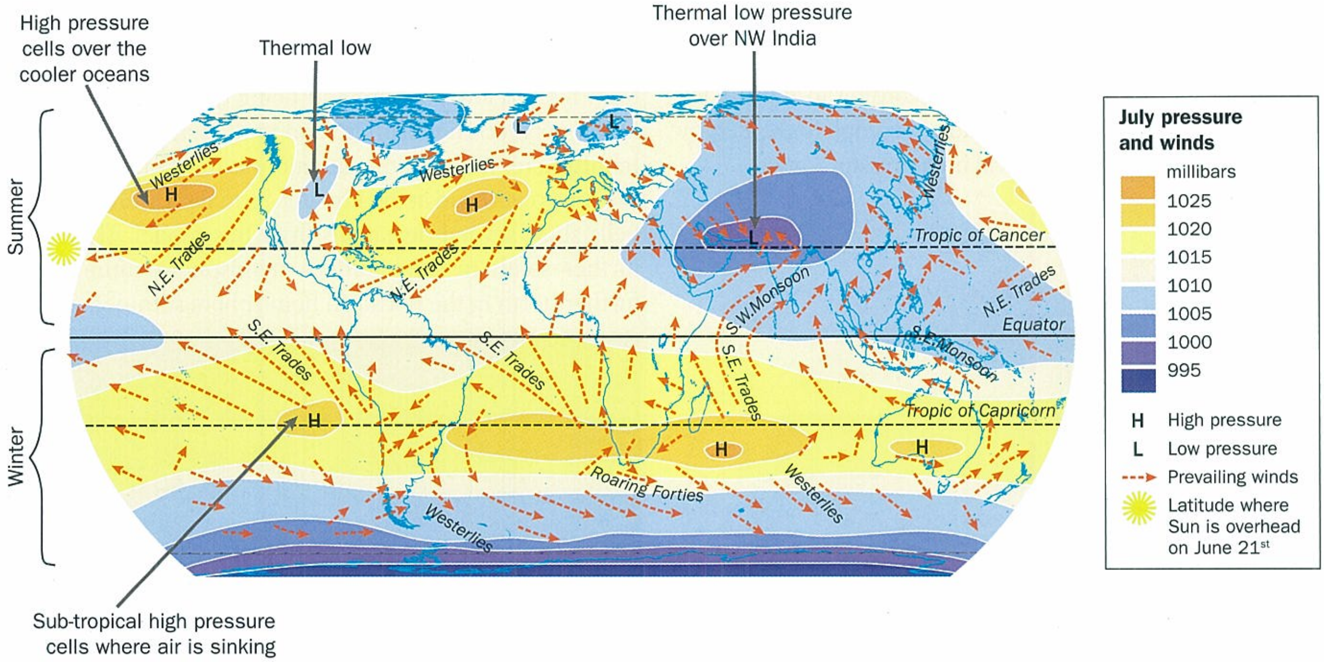


Fig. 2.32 Isobar map of surface pressure belts and their influence on winds in July

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

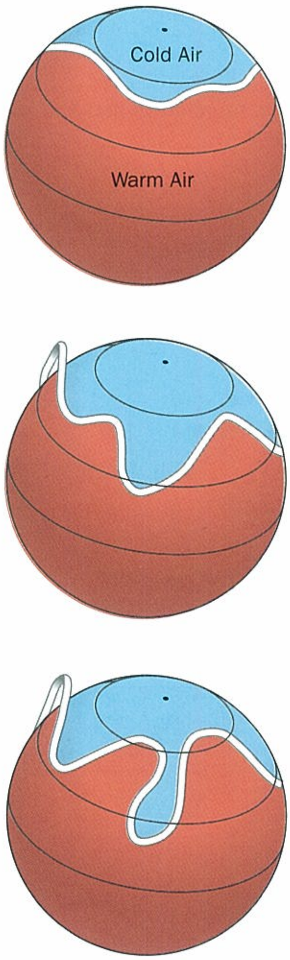


Fig. 2.33 The growth of a wave trough in the upper westerlies of the northern hemisphere until just before a pool of cold air is cut off to be mixed with warmer air. The growth of a ridge towards the north can cut off a pool of warm air in the same way

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

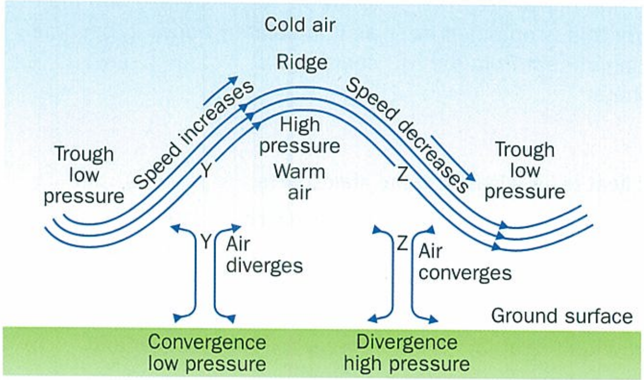


Fig. 2.34 The relationship between Rossby waves and pressure systems at the ground surface. The diagram combines a plan of a Rossby wave and vertical sections of airflow at points Y and Z

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.

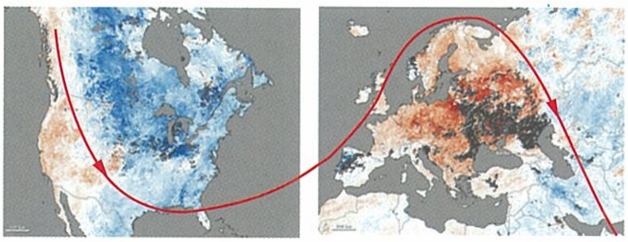


Fig. 2.35 Rossby waves superimposed on to satellite maps of resulting temperature anomalies for January 1–7 2014 in North America and Europe

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.

Condensation

Any *further* cooling *below* dew point temperature (the temperature at which the air is saturated), causes excess water vapour in the air to condense to water droplets, provided that there are **condensation nuclei** in the air for this to occur. These might be tiny dust, salt or smoke particles that form a nucleus around which condensation can begin. They are hygroscopic (have an affinity for water) and can sometimes even result in condensation in air with a relative humidity of below 100 per cent.

If the air humidity drops below saturation, the opposite will occur – the droplets will evaporate. Saturation can be attained either by the addition of water vapour into the air, as when the air moves over a warm, evaporating sea surface, or by cooling, which is the more important method.

Cooling of the air can be achieved in three ways:

1 Conduction (contact) cooling

Contact cooling leads to condensation when moist air comes into contact with a cold object whose temperature is below the dew point of the air. This may be a cold land surface which has lost heat rapidly by terrestrial radiation during a cloudless night, or a cold sea surface.

2 Radiation cooling

Air loses heat to space by long-wave radiation from clouds and gases in the atmosphere.

3 Expansion cooling

When air is forced to rise, it expands because it is rising into thinner air. When gases expand, their temperature falls, so the air cools. (Similarly, when air descends it is compressed because it is sinking into denser air, so it warms.)

Causes of precipitation

Precipitation is moisture that is deposited on the Earth's surface in liquid or solid form from the atmosphere. For it to occur, air has to cool to below dew point temperature, either by being forced to rise or by conduction.

Vertical movement can be triggered by convection, frontal uplift, orographic uplift and radiation cooling.

Convection

Convection occurs over a hot land surface or over a structure that emits heat like the cooling tower of a thermal power station. The air is warmed by contact with the heat source. Convection is common during summer anticyclones when air is sufficiently still to enable long contact with the hot land surface. The heated air expands, becomes less dense, rises and cools. Clouds form as moisture in the air condenses and the latent heat released adds further warmth to speed the ascent, which can reach the tropopause, forming cumulonimbus cloud.

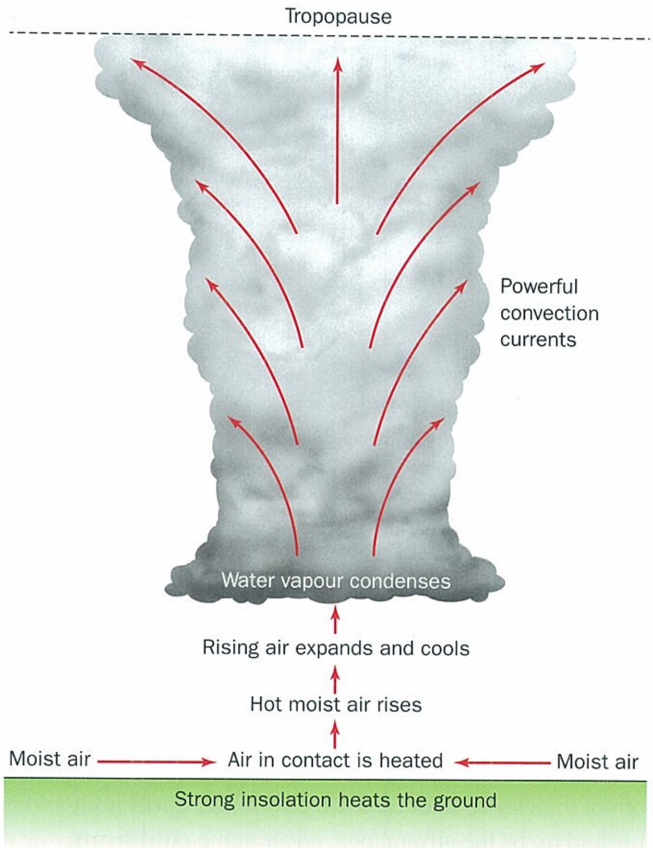


Fig. 2.38 Cumulonimbus cloud and rainfall

Fig. 2.39 Convection cells causing adjacent cumulonimbus clouds with horizontal spreading beneath the tropopause in the UK in summer

Frontal uplift

At the ITCZ moist tropical air masses of the same temperature meet and rise as a result of winds moving into the equatorial low pressure system, resulting in dense cumulonimbus cloud formation, especially in the afternoon and evenings.

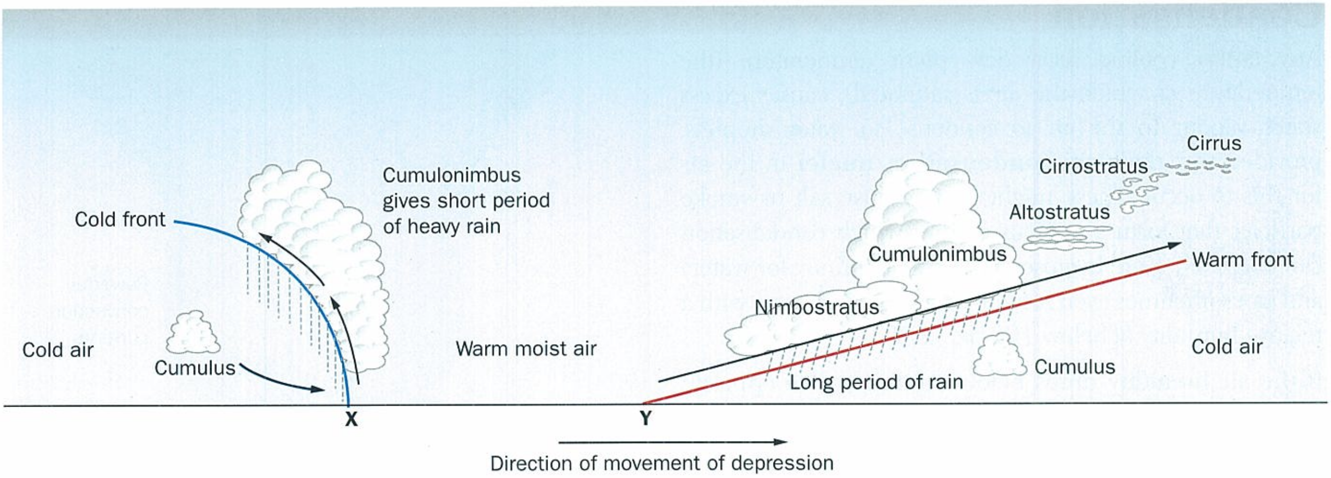


Fig. 2.40 Section through a mature depression showing warm air rising over cold air along the warm and cold fronts

Warm tropical air masses and cold polar air masses meet along the polar fronts in mid-latitudes. Low pressure systems called depressions form and, at their mature stage, have warm and cold fronts, the fronts of the advancing warm and cold air masses respectively. Being less dense, the warm moist air in the 'warm sector' of the depression can't push the cold air out of the way, so rises over the cold at the warm front and is pushed up by the advancing cold air at the **cold front**.

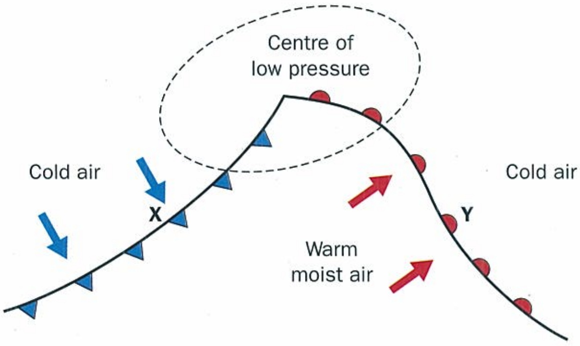


Fig. 2.41 Plan of a mature depression

A long period of precipitation occurs as the warm front passes over a place, followed by a shorter period of very heavy precipitation as the cold front moves through. Eventually, the cold front catches up with the warm front because the cold air in the rear moves faster. The warm air is completely lifted off the ground by the cold air, forming an occluded front with a long period of rain.

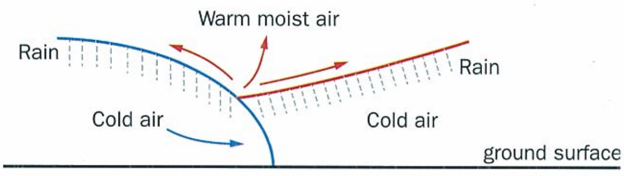


Fig. 2.42 Section through an occluded front

Orographic uplift

This occurs when the pressure force is strong enough to cause air to rise over a hill or mountain. As it rises up the windward slope, it cools and reaches dew point temperature at condensation level. Above this, condensation occurs, forming cloud from which precipitation falls if the cloud is thick. For this reason, the windward slope of a mountain is known as the rain slope while the lee slope is the rain shadow. The lee will be dry because the air is sinking and warming. Sometimes, conditions result in the cloud being too thin for rain or snow to fall from it.

Often orographic cloud gives heavy rain on the windward slope but, in certain conditions, forced ascent of air over a hill or mountain results in thin stratiform cloud, which is seen as hill fog. The situation in Fig. 2.43 occurs if the air at the top of the lee slope is cooler than the air around it at that level (a condition described as stable air), so it will be denser and sink. However, if the rising air is warmer than the air into which it is rising (a condition known as unstable air), the air continues to rise for a time instead of falling down the lee slope.

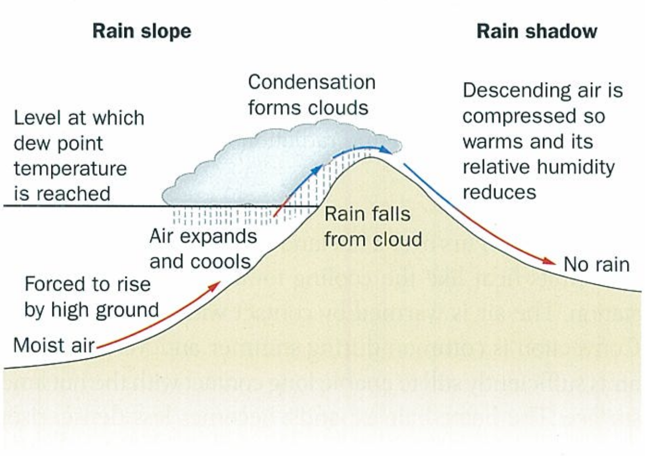


Fig. 2.43 Forced ascent of air over a hill or mountain

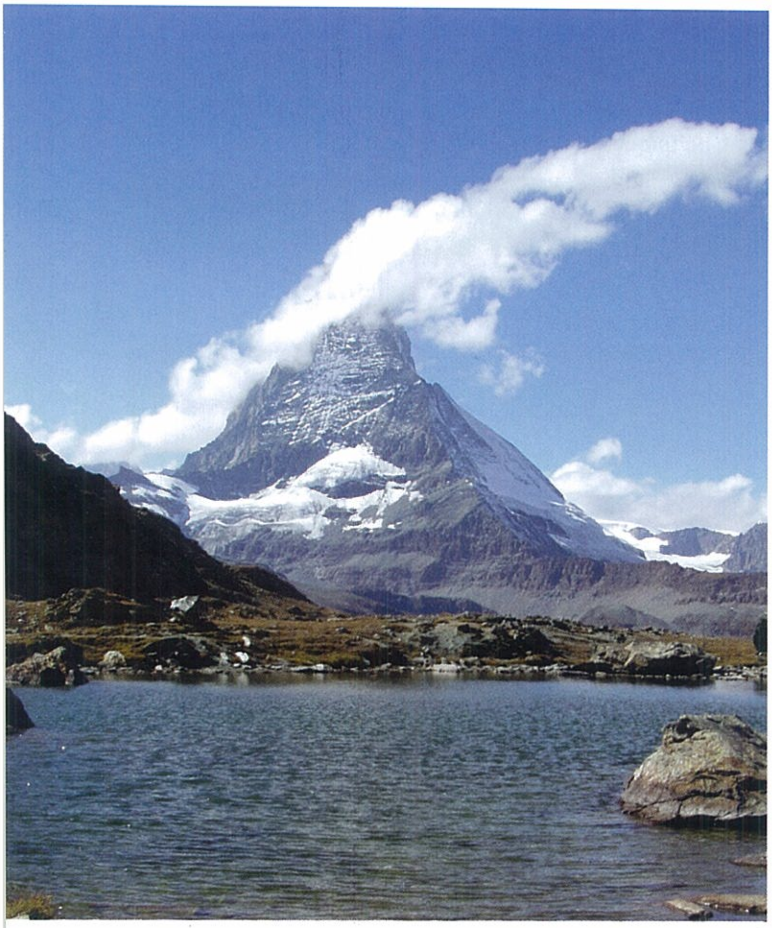


Fig. 2.44 Orographic uplift over the Matterhorn

Radiation cooling

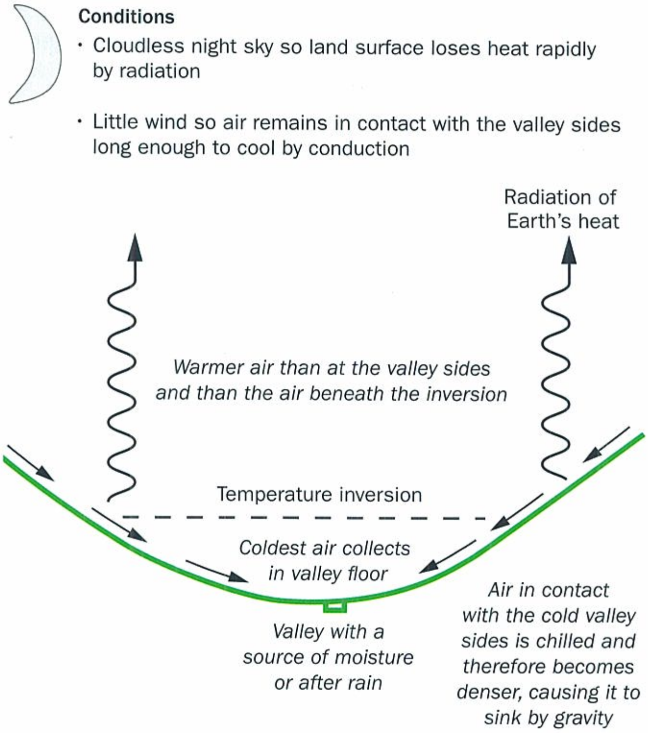


Fig. 2.45 Radiation cooling in a mid-latitude valley in winter

This is the reason why dew, **radiation fog** and ice form in valley bottoms. If the ground surface is cooler than freezing point, frost made of ice crystals occurs.

Types of precipitation

Cloud and fog are included here for convenience as weather phenomena formed of moisture, but they are not precipitation. Clouds *produce* precipitation and fog can add a little by fog-drip but most does not reach the ground.

Composed of water

Weather	Description of form	Origin
Fog	Tiny water droplets suspended in the air near the ground reduce visibility to less than 1 km.	Moist air is cooled to below dew point by contact with a cool land or sea surface and the water vapour condenses. Origin may be radiation fog or **advection fog**.
Dew	Small droplets of water on leaves, grass, spiders' webs, etc.	Water vapour condenses onto cold objects (e.g. grass). Still conditions allow prolonged contact with the object.
Clouds	At low altitudes, clouds are suspensions of tiny water droplets with an average diameter of 0.1 mm. These are super-cooled where they exist at temperatures below 0 °C and are too light to fall, so remain in the air until they are evaporated.	Formed above ground level when moist air rises, expands and cools adiabatically to below dew point temperature and the excess water vapour condenses around condensation nuclei.
Raindrops	Usually about 2 mm in diameter.	Formed when falling water droplets collide and coalesce. Some may be melted snowflakes or hailstones.

Table 2.10 Weather phenomena composed of water

Composed of ice

Weather	Description of form	Origin
Snowflakes	Aggregates of ice crystals arranged in hexagonal patterns.	Water vapour in cloud changes to ice and when the crystals fall, they collide and combine (process known as aggregation).
Hailstones	Roughly circular, made of alternate layers of opaque rime and clear ice glaze.	Formed when ice particles fall and rise in cumulonimbus clouds and super-cooled water droplets collide with and freeze around them (accretion).

Table 2.11 Weather phenomena composed of ice

Clouds

The main types of cloud

Clouds are classified into three main types according to their shapes:

- **stratus** are layer clouds that form when there is little vertical uplift but it is over a wide area
- **cumulus** occur where more vertical but localised uplift results in heaped clouds with flat bases and globular upper surfaces
- **cirrus** form where condensation occurs at very high levels, forming wispy clouds made of ice.

These are further sub-divided according to their altitude. Very high clouds are prefixed **cirro**, while middle-level clouds are prefixed **alto**. Low-level cloud is stratus or cumulus - or **stratocumulus** if it has the characteristics of both.

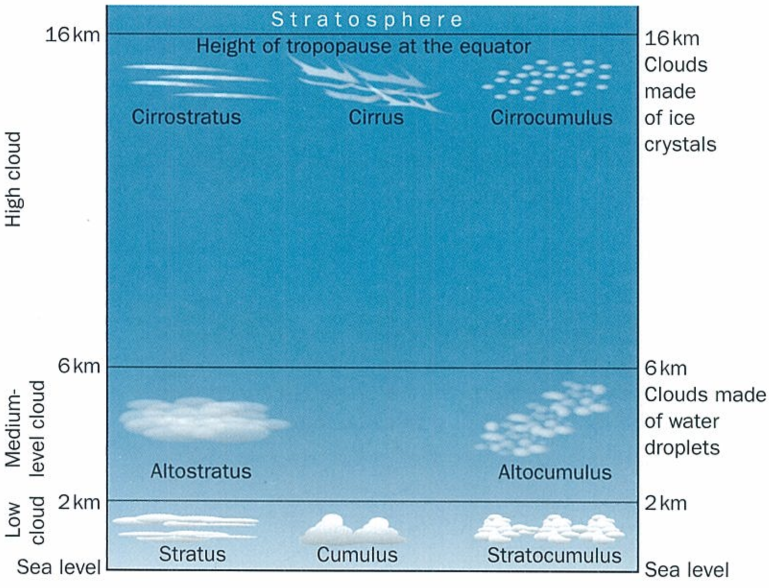


Fig. 2.46 Fair weather clouds, showing their heights in the tropics

If a cloud is sufficiently thin for sunlight to pass through, it is white and will not produce precipitation, although it could give fog if in contact with a valley side or mountain top, as the low-level layer cloud, stratus, in Fig. 2.47.

Fig. 2.47 Stratus filling a valley in the Rocky Mountains, resulting in hill fog

Clouds that produce precipitation

Rain cloud will have a large vertical extent which will prevent sunlight penetrating to its base and will appear to be black from below. **Nimbostratus** is a thick, dark grey layer cloud which can be 5000 metres deep, enough to produce steady rain. The biggest cloud by far is the towering **cumulonimbus.** It is a dense, dark grey cloud with a great vertical extent and can grow from near sea level to the tropopause, where it spreads out to form a distinctive flat anvil top because its cold air cannot rise through the warmer air above. If it does not reach the tropopause, its top is high and globular. It is composed of ice crystals at the top and water droplets lower down.

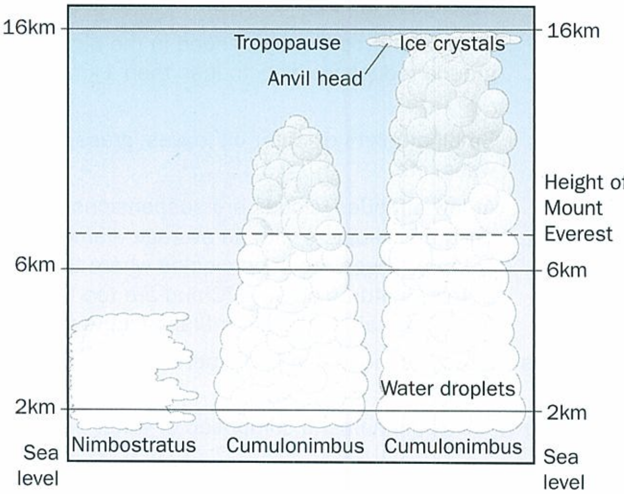


Fig. 2.48 Rain-bearing cloud, showing their heights in the tropics

Fig. 2.49 Cumulonimbus starting to develop on a sunny summer afternoon under high pressure conditions in the UK

Fig. 2.50 The base of this cumulonimbus cloud on Easter Island is at ground level because the warm air from the Pacific Ocean is very moist and cools to dew point temperature after a small uplift

Fig. 2.51 Nimbostratus is a dull grey, opaque cloud that covers the sky

Precipitation from cloud

Very few cloud droplets become raindrops. This can only happen in clouds with an abundance of moisture. For rain, hail and snow to reach the Earth's surface the droplets need to grow big and heavy enough to fall through the rising air currents that are responsible for the cloud's formation. There are two theories as to how this happens: collision theory and Bergeron-Findeisen theory.

Collision theory

Different sized water droplets have different falling rates and are carried in rising and falling air currents within cumulonimbus clouds. The droplets collide with others and join together to form a larger drop. Three processes occur:

- **coalescence** – two water droplets collide and join together. This is the main mechanism for the formation of rainfall
- **aggregation** – two ice crystals collide and join together to form snow
- **accretion** – an ice crystal collects a water droplet, leading to the formation of hail.

Bergeron-Findeisen theory

Air is saturated with respect to ice before it is saturated with respect to water. The Bergeron-Findeison process describes the growth of cloud droplets that occurs when the air is between ice and water saturation and has a temperature between -12 ° and -40 °C. In these conditions, water droplets evaporate and the resulting water vapour is deposited onto ice crystals. This continues until all water droplets are evaporated or until the ice crystals have aggregated into a snowflake and are large enough to fall. If they fall into warmer levels of the atmosphere they melt to reach the surface as rain. Most summer rain in mid-latitudes forms in this way. Snow falls if the temperature of the atmosphere remains below freezing point.

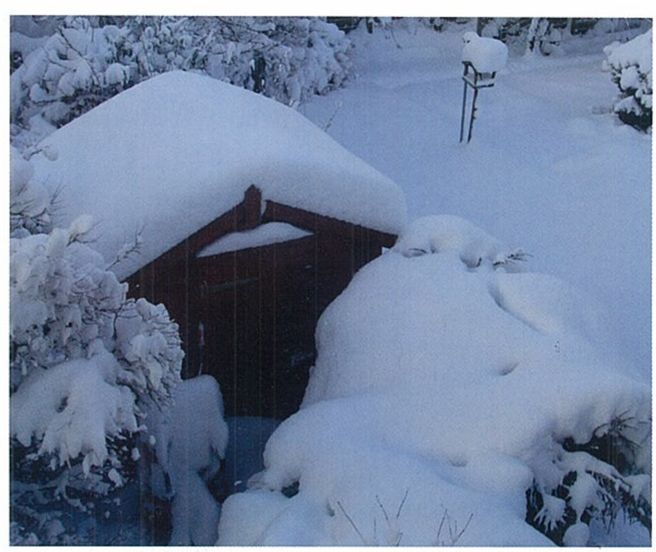


Fig. 2.52 Thick snow in the UK

Case study: A cumulonimbus experience

A parachutist practising for the **Commonwealth** Games opening ceremony in Brisbane in 1982 experienced the mighty up-draughts that are fuelled by the release of large amounts of latent heat during condensation, building cumulonimbus clouds to great heights. When he jumped from the plane into an up-draught, he was swept up 2000 m while being hit by falling hailstones. Knowing that the type of cloud could be more than 8000 m high and that at that height he would pass out through lack of oxygen, he cut his main parachute and free fell to a safe altitude. As he dropped, he experienced more hail coming up at him in the up-draughts.

Hail

The top parts of cumulonimbus clouds are below freezing point, so more latent heat is released when the water droplets that formed from condensation lower in the cloud are carried up and freeze round a freezing nucleus. This may be an ice crystal, snow pellet or frozen raindrop. Freezing is rapid in the highest parts of the cloud, so a layer of opaque ice is added to the frozen droplets. They fall again and are covered with another film of water which freezes more slowly in the warmer part of the cloud to form clear ice. Once again, they are uplifted to above freezing level where another coating of opaque ice is added. This continues until the hailstone has so many layers of ice that it is heavy enough to fall through the powerful up currents. Hailstones can be the size of golf balls and can cause considerable damage, but many melt to rain at lower levels.

Fig. 2.53 Hail from a cumulonimbus cloud at a Spanish holiday resort in May

Dew and fog

On calm, cloudless nights, air chilled by contact with the cold land surface becomes colder and denser, so sinks by gravity down the valley sides and valley floor.

Dew forms in the totally calm conditions of anticyclones if the temperature of the ground surface is above 0 °C. When there is a light breeze to mix the air near the surface, condensation takes place in the air, rather than at ground level, forming radiation fog. This often forms on winter mornings when convection just after sunrise causes air to mix.

Fig. 2.54 Early evening fog in Namibia

Fig. 2.55 Early morning fog in the UK

20. Use evidence from the photographs to explain the processes occurring in Fig. 2.54 and the reverse processes starting to occur in Fig. 2.55. Describe the effects these processes are having on the air temperature.

Advection fog occurs when:

- winds move towards the pole over a colder sea surface. Widespread advection fog forms if the air is chilled to below its dew point
- winds blow over a cold ocean current, forming advection fog over the current
- air crosses from the sea onto a cold land surface in winter. This also results in advection fog or in hill fog and stratus cloud if the air is forced to rise by the relief. Sometimes the cloud may be thick enough for a little drizzle to fall.

Fig. 2.56 Advection fog formed where warmer air from the Pacific Ocean blows over cold, icy water near a glacier in Alaska. Orographic cumulus and stratocumulus lie above the ridge

The human impact

The enhanced greenhouse effect

The natural greenhouse effect that keeps our planet warm enough to live in results from greenhouse gases, such as water vapour and carbon dioxide, which occur naturally in the atmosphere. Water vapour accounts for up to 85 per cent of the greenhouse effect in very cloudy conditions and about 50 per cent when there are no clouds. Large amounts of carbon dioxide are naturally present and are released during volcanic eruptions but they are equal to less than 1 per cent of the added anthropogenic carbon dioxide.

The **enhanced greenhouse effect** refers to the addition of greenhouse gases to the atmosphere by human activity. This is thought to lead to global warming and various types of climate change.

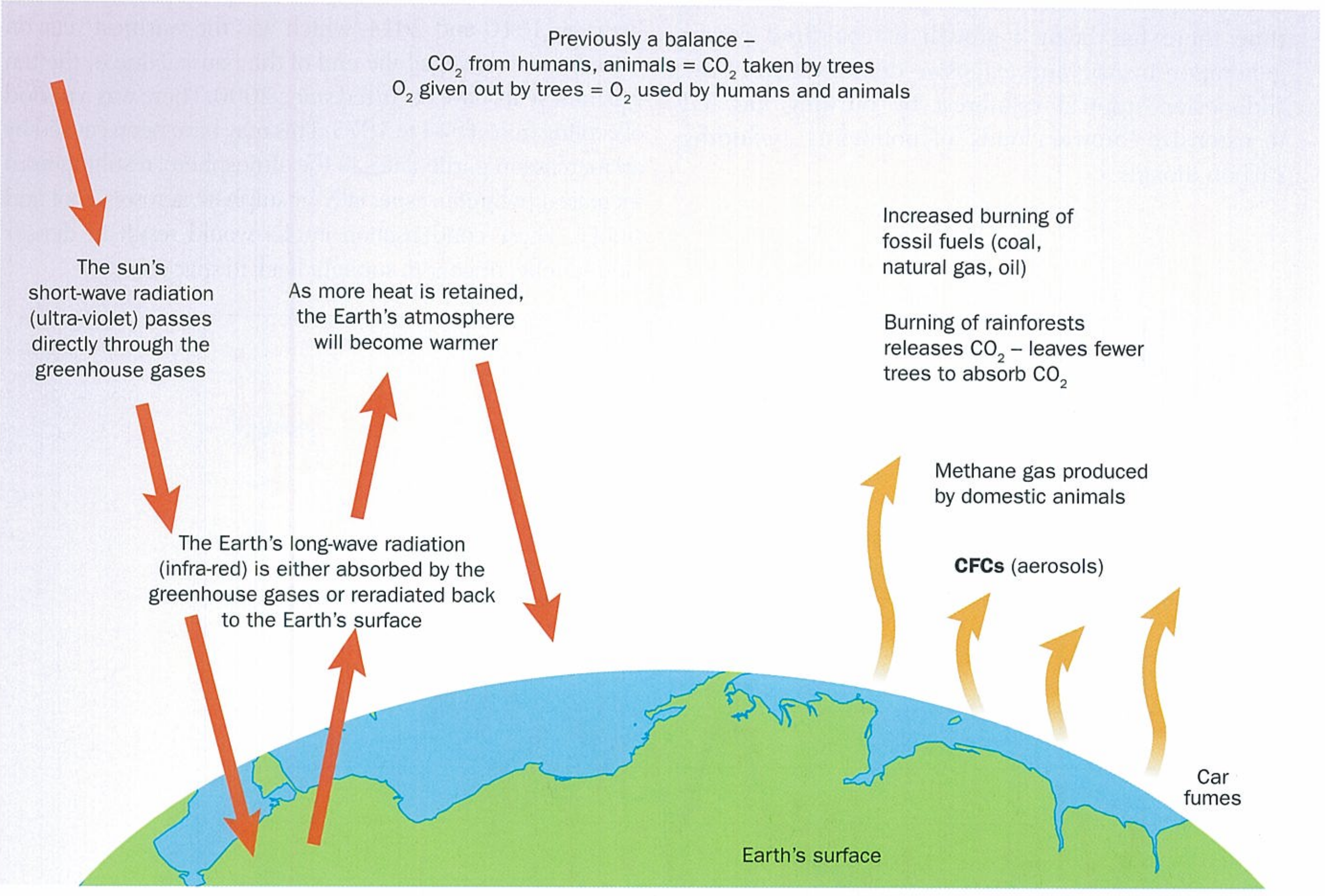


Fig. 2.57 The enhanced greenhouse effect

21. (a) Explain the meaning of 'greenhouse gas'.

(b) State which parts of Fig. 2.57 show the (natural) greenhouse effect.

(c) Explain why the effect of human activity is described as the *enhanced* greenhouse effect.

RESEARCH Find out about solar cycles and whether they can be responsible for the pattern of temperature increases in Fig. 2.58.

The warming effect of a gas is known as its **radiative forcing**. Although there is some disagreement about the degree of warming caused over a particular timescale, there is general agreement that:

- Current temperatures are 0.7 °C above pre-industrial (1880s) levels.
- Warming was less than expected during the 20th century.
- Global temperature rise since 1998 has been so little (0.05 °C in the first decade of this century) that it is described as having 'paused'. Possible reasons include changes in solar activity and a ban on CFCs. Also important are increased sulphur and ash emissions to the atmosphere from volcanic eruptions (Pinatubo in the Philippines caused cooling by 0.4 °C in 1992 by emitting 20 million tonnes of sulphur dioxide in 1991). Another possibility is that the climate is less affected by carbon dioxide than previously thought.
- There have been small-scale and large-scale periods of cooling since records began in 1880. At the same time there has been a growth of coal-fired power generation in Asia and extensive deforestation of the Indonesian tropical rainforest by burning has led to extensive 'brown clouds' of pollutants, including carbon dioxide.

Evidence for global warming

Increase in global temperatures

22. Use Fig. 2.58 to answer the following.

(a) Calculate the Earth's mean temperature change between 1910 and 2013. Is this a large amount per year? A graph should give a correct visual impression of the data it shows but there are good reasons why this is not always done. Justify the scale used for this graph.

(b) The 5-year running mean suggests that warming paused in 2002 but it is generally agreed that it paused in 1998. What is the evidence for this?

(c) Explain how the five-year running mean was calculated.

Fig. 2.58 shows that mean world temperatures rose overall between 1910 and 2014, which was the warmest year on record and may signal the end of the pause. Nine of the ten warmest years have occurred since 2000. There was a period of cooling from 1944 to 1975. This may have been caused by an increase in particulates in the atmosphere, resulting from increased pollution, especially by sulphate aerosols, soot and smoke. These condensation nuclei would result in denser clouds, reflecting more sunlight back to space.

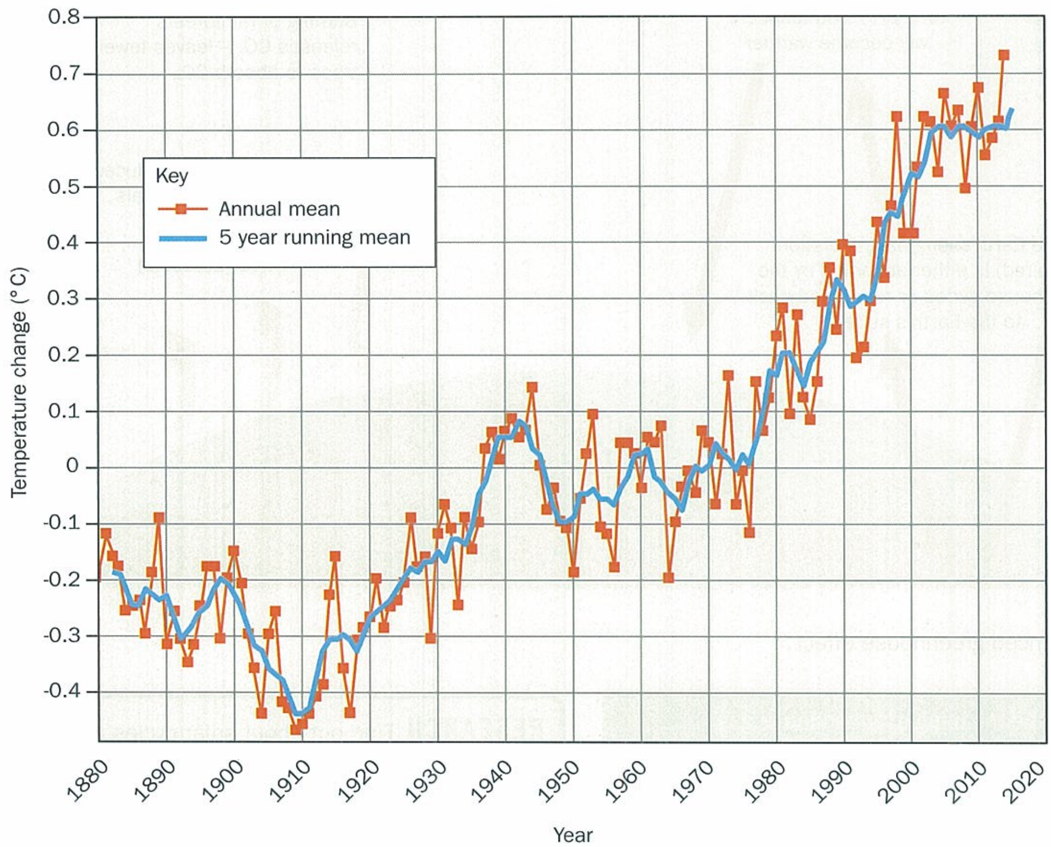


Fig. 2.58 Line graph showing how global annual temperature (in °C) and the 5 year running mean 1880–2014, varied from the mean temperature of the period 1951–1980, shown as 0 °C

Each decade since the 1970s has been a little warmer than the preceding one. This has been attributed to the increase of greenhouse gases in the atmosphere, whereas much of the increase between 1910 and 1940 was thought to be largely natural.

Upper ocean heat has increased steadily over the past 20 years as a result of the oceans storing an estimated 63 per cent of the heat added by enhanced global warming. Monitoring of the deep oceans suggests heat is accumulating there, rather than in the atmosphere.

Ocean salinity

Drier areas and areas of high evaporation, such as the western Indian Ocean, have become saltier whereas wetter areas, such as the North Atlantic, have fresher water. As the Arctic sea ice and glaciers melt and rainfall increases, more freshwater is entering the north Atlantic Ocean and diluting the salt water, making it less dense. This will cause it to remain on the surface and could prevent the sinking of north Atlantic surface waters which power the Great Ocean Conveyor Belt. Tropical water will then no longer move towards the poles because there will be no sinking water to replace. These ocean currents show signs of weakening, which would make Europe colder.

Melting sea ice

Arctic sea ice extent has declined since satellite observations began in the late 1970s. Its March maximum extent declined by about 2.4 per cent per decade since 1980 and was at a record low in 2012, but recovered considerably in 2013 only to fall to the third lowest June extent in the satellite record in 2015. The ice is also becoming much thinner but, according to the European Space Agency, thickened by up to a third in 2013.

By contrast, in June 2015 Antarctic sea ice reached its third highest June extent since satellite records began. It is believed to have increased by over 4 per cent per decade recently and has been well above the mean in 9 of the 13 years between 2001 and 2014, even though an active volcano is beneath it. To complicate matters, scientists have recently found two channels that may be allowing warm ocean water to move beneath the 75-mile-long Totten Glacier. If proved, it would contribute to rising sea levels by melting ice. Between 2010 and 2014 the extent of Antarctic ice increased from 19 million square kilometers to more than 20 million square kilometers.

Rising sea levels

Global sea level rose by an average of 3.1 mm a year between 1992 and 2010, twice the average rate for the last century. The record highest yearly average height was in 2014, when it reached 67 mm above the 1993 average. It is rising because melting glaciers send more water to the sea and also because of thermal expansion - water expands as it warms. As oceans store heat for longer periods than land and store it over a large body of water, a small increase in temperature could result in a significant rise in sea level.

RESEARCH Venice is regularly flooded to increasing depths. The rise in sea level there partly results from ice melting and partly from the Earth's crust sinking in that region. Find out about sea level changes caused by tectonic and isostatic movements.

Melting glaciers

With some exceptions, most mountain glaciers have been shrinking since the middle of the 19th century. In 2013, three Norwegian glaciers were advancing and all in Nepal were either stable or growing because snowfall was greater than average in 2013. Glaciers and ice sheets in Greenland and West Antarctica are melting rapidly at the highest speeds since satellite records began and increasing sea levels. Fortunately, the East Antarctic ice sheet is increasing.

Fig. 2.59 This glacier in Chile used to reach the sea

Increasing acidity of the oceans

Oceans are thought to have absorbed about 50 per cent of the carbon dioxide emissions released by human activity by dissolving carbon dioxide into them, forming carbonic acid. Since 1750, the pH of the ocean surface is thought to have fallen by 0.1, from 8.2 to 8.1. This is a logarithmic scale and is equivalent to a 26 per cent increase in acidity. The 1750 level was calculated by analysing the CO_2 content of air bubbles trapped in glaciers at that time and the knowledge that the ocean surface has the same concentrations of CO_2 as the atmosphere. As pH 7 is neutral, a more accurate description of the situation is that the ocean surface is a little less alkaline now.

Biological indicators of warming

Examples include:

- the bee-eater, a tropical bird, is now found in the UK every spring
- malaria, a tropical disease, is increasing in southern Europe where mosquitos have moved in with the increased warmth
- the blanching of some coral reefs is believed to result from increased acidity of the oceans.

Possible causes of global warming

Greenhouse gas	Atmospheric concentration	Sources resulting from human activity	Number of years the gas stays in the atmosphere	GWP*	Contribution to the enhanced greenhouse effect
Carbon dioxide	400 ppm (2014 at Hawaii and some other locations)	Burning fossil fuels and wood, deforestation; especially by burning.	Variable: up to 200 but averaging about 62	1	Thought to be the main greenhouse gas, it has increased from 280 ppm in 1850.
Methane	1800 ppb	Bacteria in wet padi fields, bogs, waste landfill sites and the guts of cattle and sheep.	12	25	In small quantities but 25 times more effective than CO2. Increasing by up to 2 per cent p.a.
CFCs	1863 ppt	Old aerosols and refrigerators (CFCs are no longer used)	Variable – up to 50 000	Variable – most more than 3500	Very efficient absorbers of long-wave radiation.
Nitrous oxides	323 ppb	Nitrate fertiliser, burning fossil fuels (especially diesel engines) and burning vegetation.	114	nearly 300	In small quantities but impact nearly 300 times that of carbon dioxide.

*GWP is the global warming potential of the gas compared with carbon dioxide (1) shown for a 100 year period

Table 2.12 Greenhouse gases

The radiative forcing of the greenhouse gases increased by nearly a third between 1990 and 2012, with carbon dioxide contributing 80 per cent of this increase in warming, according to the World Meteorological Association.

Carbon dioxide has been of most concern because it stays in the atmosphere for much longer and is much more abundant than methane. Measurements at the Mauna Loa Observatory in Hawaii show that CO_2 concentration in the atmosphere increased fairly steadily from 1960 to 2014 with a slightly increased rate from about 1995, which is about the time that warming appeared to have paused.

People are being made aware of their '**carbon footprint**', a measure illustrating the impact human activities have on the environment. The carbon footprint will be proportional to the amount of greenhouse gases (such as carbon dioxide) produced.

23. Use the information in Table 2.13 to construct a percentage divided bar chart to show country shares of the world total CO_2 emissions. Add 13 per cent from other MICs and 29 per cent from other LICs.

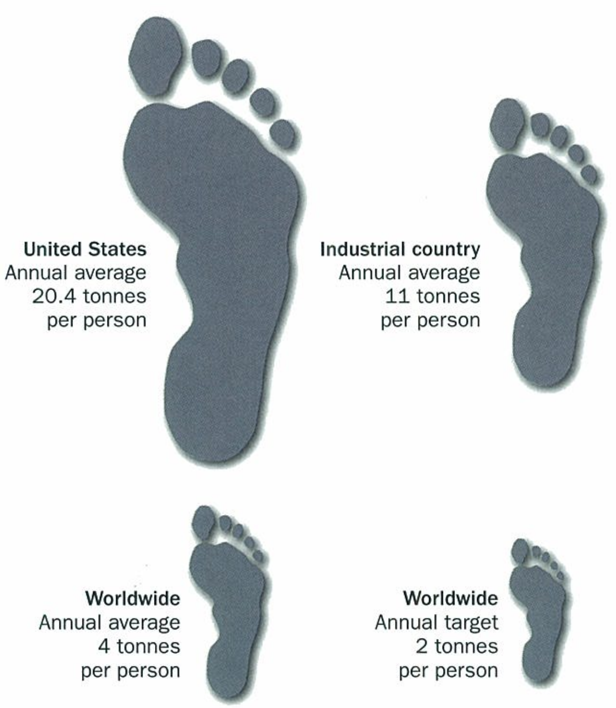


Fig. 2.60 Carbon footprint awareness can be raised by using easy to interpret diagrams. The target is what is thought necessary to limit global warming to 2 degrees C this century.

Country	2009 (million tonnes)	2011 (million tonnes)	Change from 2009	2011 emissions per person (tonnes)	2011 country share of world total (per cent)
China	7037	8715	increase	6.2	27
USA	5657	5490	decrease	17.2	17
Russian Federation	1716	1788	increase	12.2	5
India	1802	1725	decrease	1.4	5
Japan	1208	1180	decrease	9.2	4

Table 2.13 Emissions of carbon dioxide (Source: CDIAC, US Department of Energy)

RESEARCH Find out the main sources of carbon dioxide emissions for one of the countries in Table 2.13.

El Niño

Global temperature changes appear to be linked with natural events in the equatorial Pacific Ocean known as **El Niño** and La Niña. These are characterised by different pressure patterns and reversals of wind and ocean water movements. El Niño is also known as ENSO (El Niño Southern Oscillation).

There are discrepancies between data shown on Figs 2.58 and 2.61: for example, the anomaly for 2014 was +0.74 on Fig. 2.58, whereas on Fig. 2.61 it is +0.69. Both diagrams were obtained from NASA sources - the first from the Goddard Institute of Space Studies and the second from NOAA. These institutions use their own methods of research, leading to minor variations, but are in broad agreement, as are results from research by the Japanese Meteorological Agency and the UK Meteorological Office's Hadley Centre.

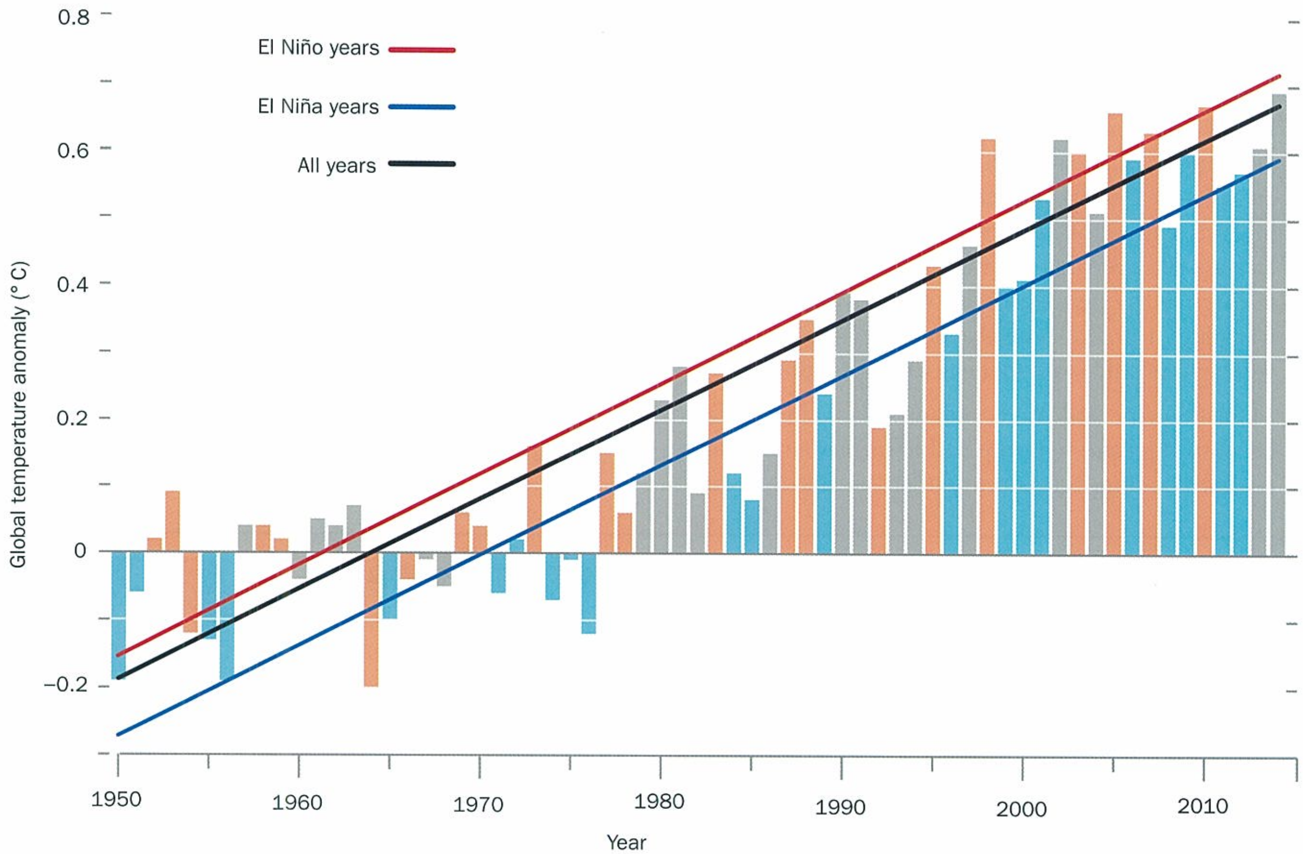


Fig. 2.61 Bar graph showing the relationship between global temperature anomalies calculated from the 1951 to 1980 mean temperature, shown as °C, and El Niño and La Niña events

During El Niño events, such as in 2010 (the second hottest year on record), but not 2014, the warmest, the world is warmed by heat from the Pacific Ocean. The Pacific warms because the normal westward surface flow of ocean currents reverses and warm water moves east across the Pacific from Indonesia. Large areas have sea surface temperatures in excess of 28 °C, causing low pressure over the eastern Pacific Ocean and heavy rain in Peru and northern Chile. Meanwhile, eastern Australia experiences high pressure, drought and extreme heat. In El Niño years, sea levels rise because there is more rainfall over the warmer oceans and warmer waters expand.

La Niña brings cooler years. Cold upwellings of water off the coast of Peru make the central and eastern part of the Pacific Ocean up to 2 °C cooler than average while eastern Australia has heavier than normal rainfall. Some scientists are linking the 'pause' in global warming to La Niña cooling the eastern Pacific.

Atmospheric impacts of global warming

→ The melting of Arctic sea ice and loss of snow cover will reduce the albedo, resulting in less reflection of solar radiation and more heating of the Earth and atmosphere, so accelerating global warming.

→ Warmer temperatures will lead to more evaporation from the oceans, increasing moisture in the atmosphere, giving the potential for increased cloud and rainfall in places, which would cause local cooling.

- Warmer temperatures in places with high pressures, such as south-west USA, South Africa, Australia and the Mediterranean, will cause less rainfall and more severe droughts.
- Heat waves will occur especially in cities because urban temperatures are always higher than those of surrounding rural areas. 30 000 people died in Europe in the hot summer of 2013.
- Western Siberia is warming quickly because, as the covering of snow and ice melts, the darker rock and soil surfaces absorb much more radiation. If all the permafrost melts, massive quantities of methane will be released into the atmosphere, accelerating global warming. Methane releases from the melting permafrost are already causing numerous wildfires in Alaska and Siberia. Permafrost also stores more than twice the amount of carbon that is in the atmosphere, which will be released as the permafrost decomposes and accelerates warming. If it accelerates beyond the 'tipping point', it will be impossible to stop the warming continuing.
- More frequent and more violent storms are expected because of greater moisture in the warmer air and more coastal flooding from storm surges are likely because of higher sea levels. However, no connection has been found between the number of tropical storms and warmer years. Despite being the fourth warmest year on record, 2013 was the quietest year for North American hurricanes with only one reaching category 1 status.

The amount of CO_2 in the atmosphere has not increased as much as expected. Carbon dioxide that is not absorbed by the atmosphere is stored in other carbon 'sinks', such as the oceans and vegetation.

The changes will not be uniform and are difficult to predict. Some areas will be hotter and others colder. Some will be wetter and others, such as the Amazon basin, drier. Areas of unequal pressure will still exist to drive the winds but they could be in different locations. The greatest changes will be to areas that are near climatic boundaries.

Difficulties of predicting climate change

Predictions of the degree of warming made by the IPCC, a panel of experts, were revised down a little in their 2014 report. They depended on computer models, so were only as good as the weightings attached to the various inputs.

Case Study: Vancouver's microclimates

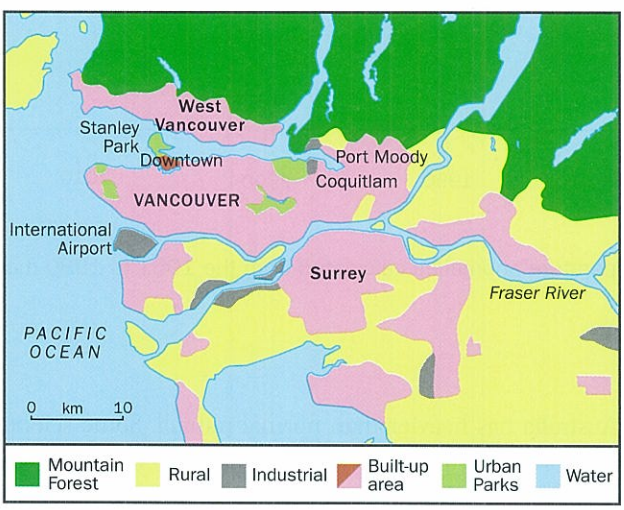


Fig. 2.62 Vancouver's land use zones and locations in the case study

Vancouver, Canada's third largest city, is open to the sea but backed by mountains broken by the valley of the Fraser River, which enters the coastal lowland to the east of the city.

Temperature

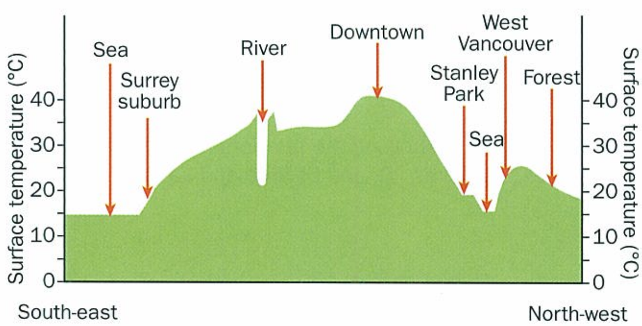


Fig. 2.63 Temperature transect across Vancouver's heat island on a very hot day in July

On sunny summer days a surface temperature traverse across the city shows a marked heat island. The sea has the lowest temperature, followed by the river. Stanley Park also has relatively cool temperatures, similar to those in the rural farmland of the Fraser Valley. The international airport is not shown on the diagram but would record a higher temperature than the well-vegetated Stanley Park because of the terminal buildings and runways. Temperatures are higher in the suburbs of West Vancouver and Surrey but the 'dome' of the heat island is centred on the downtown (CBD) area where there is a high density of skyscrapers.

When sea breezes blow on some summer afternoons, air temperatures near the coast can be 5–10 °C cooler than in the Fraser Valley beyond their influence, whereas in winter it is the inland areas that are cooler, by about 2 °C.

	Downtown	Airport
Daily mean temperature (°C)	11	10.4
mean annual precipitation (mm)	1587	1189
Mean annual sunshine hours	1818	1937

Table 2.14 Some climatic differences within Vancouver

Fig. 2.65 shows atmospheric daytime temperatures typical of the summer temperatures in the area. Again, the CBD is the warmest area at 27 °C and is 10 °C warmer than the rural area to the south of the city, where the temperature is only about 17 °C. There is a steep rise in temperature at the edge of the built-up area, then it increases more gradually to 25 °C before peaking in the CBD. There are two peaks of temperature, separated by cooler temperatures over a small inlet of water.

The difference in temperature between urban and rural areas is also high on summer nights under calm and cloudless anticyclonic conditions when there are no winds to remove the heat. In those conditions,

Fig. 2.64 The view north-west from downtown Vancouver, showing the forested mountain rim with the suburb, West Vancouver, in the background and part of Stanley Park on the left

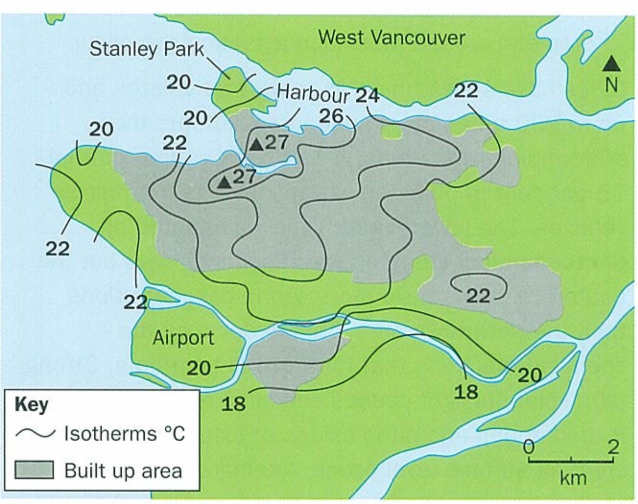


Fig. 2.65 An isotherm map of air temperatures in the Vancouver area

temperatures can decrease steadily outwards from 20 °C in central Vancouver to 15 °C at the international airport. All large urban areas have earlier springs, later autumns and fewer frosts than nearby rural areas. This heat island effect is caused by several factors:

- Brick, concrete and dark tarmac surfaces have a large specific heat capacity and low albedos. They rapidly absorb solar radiation by day, especially in summer. The heat is conducted away from the surface, stored and re-radiated steadily into the air at night as long-wave radiation.
- Tall buildings turn streets into 'canyons', giving a reduced sky view and a smaller angle of direct heat loss at night. Large urban parks, such as Stanley Park, and open countryside are cooler because the sky is much 'wider', so radiational heat losses are greater. The greater the building density, the greater the heating, so city centres have the highest minimum temperatures at night and temperatures gradually cool through the more spacious suburbs to the countryside.

- Central heating, gas and electric fires, together with heat from industrial sources, all warm the urban atmosphere.
- Although pollutant particles and gases (such as sulphur dioxide from the combustion of fossil fuels and carbon monoxide from vehicle exhausts) reduce the amount of short-wave radiation reaching the surface by day, their main effect is reducing the loss of re-radiated energy at night by absorbing and re-radiating long-wave radiation.
- Higher rates of evapotranspiration in rural areas lower their temperatures by absorbing heat.

Wind

Skyscrapers in the CBD can cause turbulence and vertical uplift with eddies in the lee of buildings, but winds are generally less strong in the city centre because the tall buildings obstruct and deflect airflow and increase friction. There are more calm days in the built-up area than in the surrounding countryside. However, canyon-like streets have the strongest gusts when winds are funnelled between their tall buildings.

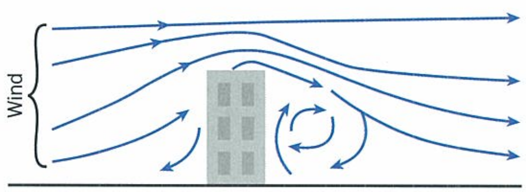


Fig. 2.66 The effects of a building on airflow

Humidity

Absolute and relative humidity are both lower during the day in the city than in the rural area because less surface water and vegetation is present for evaporation and transpiration to occur. Rainfall runs off rapidly down gutters and drains. However, humidity rises over the harbour, Burrard Inlet and the Fraser River where more evaporation is possible. Daytime humidity is also greater over Stanley Park where moisture is increased by transpiration.

At night, humidity tends to be slightly higher over the city because in rural areas more dew is deposited from the air, making it less humid.

Precipitation

An orographic effect is present as, for every increase in height of 100 metres within Vancouver, precipitation increases by about 100 mm. The mountain edge of West Vancouver receives double that of the main city. Snow falls there in winter but is rare in the lower areas, the parts of the city nearer the warmer ocean and in the central city, where the warmth melts most snow as it is falling.

Relief is the most important cause of variations in rainfall within the city but the greater amount of cloud and precipitation also results from:

- the development of convection currents on calm, clear nights when the urban-rural temperature differences are largest. Warm air rises over the central city and the airflow moves towards the rural area aloft and then sinks. Meanwhile, air moves from the rural area over the surface to the lower pressure in the city. These convection currents result in higher rainfall in the central city, with more thunderstorms. It can be 10 per cent cloudier than in the surrounding rural area
- storm cells intensifying by contact with the warm surface as they pass over
- buildings forcing moist air to rise, triggering convection
- the increased density of hygroscopic (condensation) nuclei strongly encourages condensation.

Fog and smog

Fogs are more frequent in the city, particularly in winter, because:

- lower wind speeds and more frequent calm periods allow the air to remain in contact with the ground and to cool by conduction
- there are a greater number of hygroscopic nuclei (e.g. smoke and dust particles and sulphuric acid droplets) around which condensation can occur.

Cities have seven times as many particulates and up to 200 times more gaseous pollutants than surrounding rural areas. It has been estimated that 75 per cent of pollution in Vancouver is from motor vehicles. The lower Fraser Valley is affected by **photochemical smog** on warm summer days but the eastern suburbs have more ozone concentrations than the downtown area. This is because the chemical reactions that form ozone take time. Strong ultraviolet sunlight causes chemical reactions in the hydrocarbons and nitric oxides of exhaust fumes, creating surface level ozone, the ingredient of photochemical smog. It is gradually produced as sea breezes move the pollutants inland. Mountains to the north and east trap the pollutants in the eastern suburbs of Port Moody and Coquitlam.

The worst pollution occurs during anticyclones because the subsiding air warms and causes a temperature inversion above the city, which acts as a lid on rising air and traps the pollutants below it. The light winds in anticyclones do not disperse the pollutants, so assist fog to form and mix with pollutants to form smog.

Fig. 2.67 Aerial photograph of photochemical smog. The level of the temperature inversion 'lid' and trapping effect of the mountain rim can be clearly seen

Key concepts

The key concepts listed in the syllabus are set out below. For each one a summary of how it applies to this chapter is included.

Space: throughout the chapter, space is evident in the global distributions of radiation energy, temperature, wind belts and ocean currents. Collectively, they use all the space the world has. The patterns of radiation energy, temperature and winds are evident in maps covering all land and sea surfaces, while ocean currents not only span ocean surfaces but also great depths down to and along the ocean floors. This space is dwarfed by the immensity of atmospheric space; radiation energy from the sun crosses 149.6 million kilometres to warm the Earth.

Scale: influences on temperature vary from those operating at the global scale which are controlled by latitude, to others which complicate the global model at a local scale, such as distance from the sea and cloud cover. Human impact is considered in this chapter at both the global and city scales in the studies of climate change and the microclimates of cities. Timescale is also important: for example, the speed of movement of a weather system determines whether it rains for a short or a long time, or the late arrival of the wet monsoon which has severe consequences for the people dependent on it for growing crops.

Place: location has a very important influence on the weather and climate a place experiences. Proximity to the ocean compared to an inland location results in differences in atmospheric pressures, temperatures and precipitation amounts. Similarly, planetary winds and pressure systems are complicated by local differences resulting from the unique physical characteristics of a locality. Very localised small-scale changes are illustrated in the study of microclimates where temperatures and humidity vary between riverside and downtown locations or urban parks and areas of high building density.

Environment: the interactions between people and their environment do create the need to manage the environment sustainably. This is well seen in studies of climate change; research strongly suggests that sustainability depends upon limiting the emissions of greenhouse gases into the atmosphere. The wide-ranging impacts of not doing so, such as the melting of polar ice caps resulting in substantial rises in sea level, will impact adversely on many people. A vicious cycle starts with humans interfering with atmospheric composition and ends with the atmosphere interfering with human activity. In addition, the study of microclimates in cities illustrates that variations result from differences in human use of the various physical environments within the city boundary. The problem of photochemical smog in Vancouver and many other cities is caused by humans using polluting vehicles and combustion. It needs managing sustainably to reduce this important cause of ill-health.

Interdependence: weather is the result of complex interacting systems and processes. Differences in radiation amounts lead to different pressure systems which, in turn, result in winds which bring particular moisture and temperature characteristics to their destination areas from their source regions. Human systems then modify the temperature and rainfall by interfering with the composition of the atmosphere, which has the effect of modifying the weather, causing increased storminess and rainfall in some places and drought in others.

Diversity and **change**: these are evident in the variation of weather from hour to hour, day to day or season to season and also from place to place, as the controlling energy inputs and pressure systems change.

Exam-style questions

1 Fig. 2.68 shows information about the weather in the northwest of the USA on a day in August when there was a westerly wind.

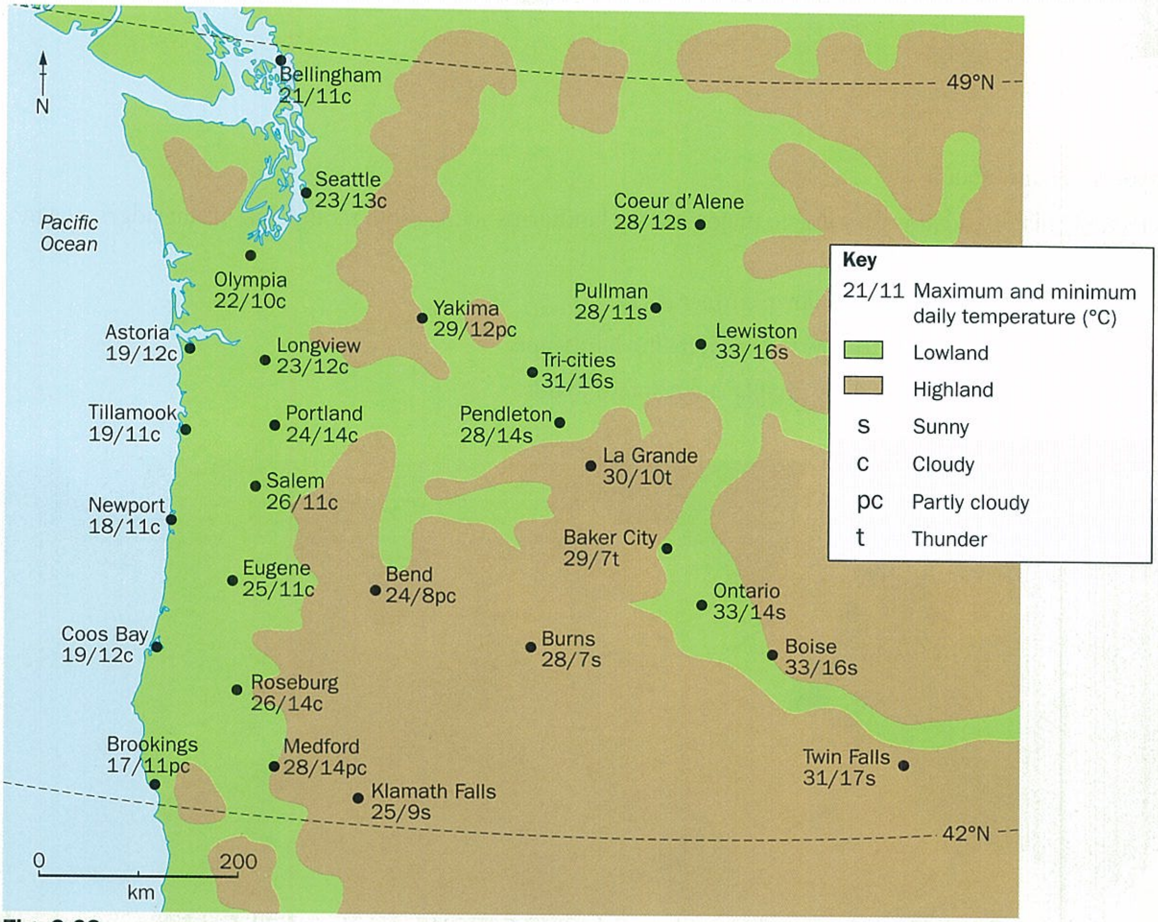


Fig. 2.68

(a) (i) Calculate the diurnal temperature range at Astoria on the coast. [1]

(ii) Using words only, compare the maximum temperatures at Astoria and Lewiston. [1]

(b) Using evidence from Fig. 2.68, describe the relationship between maximum temperature and distance from the sea. [4]

(c) Explain differences in cloudiness within the area that is south of Eugene, Bend and Ontario. [4]

2 (a) (i) Define the term enhanced greenhouse effect. [3]

(ii) Describe the differences between solar radiation and terrestrial radiation. [4]

(b) Describe and explain the air circulation in the Hadley cell of the northern hemisphere. [8]

(c) Assess the extent to which precipitation is dependent on air movement. [15]