7. Radiation and Global Climate

Earlier in this course, we discussed some simple ideas on radiative transfer from the Sun and the Earth, and developed a radiative equilibrium model for the Earth-atmosphere system. We are now going to consider how the surface radiation balance varies as a function of latitude, explaining why the temperature at high latitudes is lower than in the tropics. We will also consider the seasonal variation and the diurnal variation of radiation.

Review of surface radiation balance

In section 3 of the course, we developed a radiative equilibrium model for the Earth's surface, where the incident solar and long wave radiation from the atmosphere balances the emitted long wave radiation from the surface. This was developed as a global average model, giving

Absorbed radiation *Emitted radiation*

$$\textit{Surface}: \quad (1-b)(1-\alpha)\frac{S}{4}+\varepsilon_a \sigma T_a^4 = \sigma T_s^4$$

where S is the solar constant, α is the surface albedo, b is the atmospheric absorption and scattering of short-wave radiation, ε_a is the atmospheric absorption/emission of long-wave radiation, T_a and T_s are the temperatures of the atmosphere and the Earth, respectively, and σ is the Stefan-Boltzmann constant.

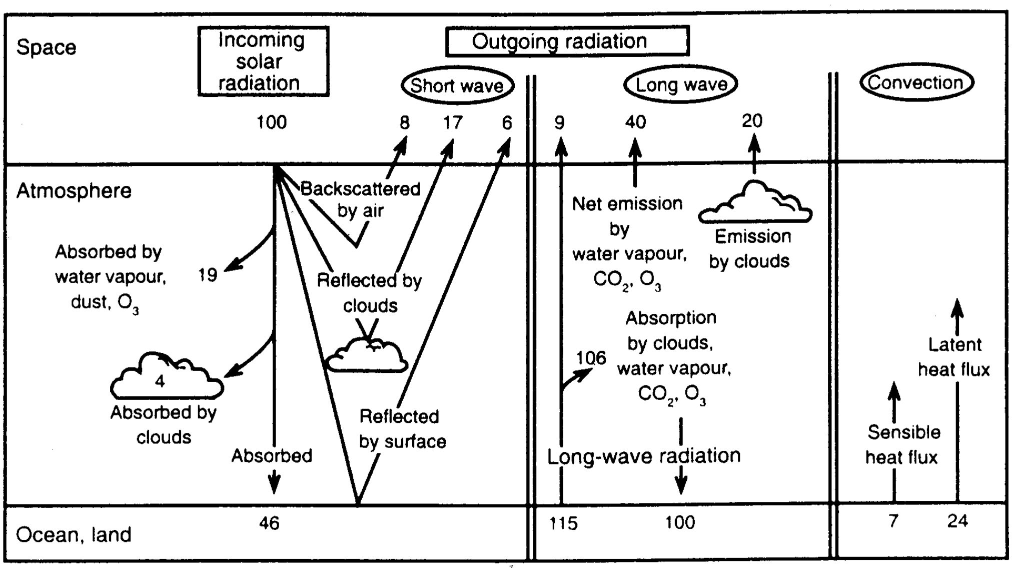

Figure 7.1 (reproduced from Fig. 3.3 earlier) Global radiation and energy balances. The units are percentages of incoming solar radiation to the Earth-atmosphere system.
(from Sturman and Tapper, 1996).

Figure 7.1 (reproduced from Section 3) shows the different processes affecting radiative transfer in the Earth-atmosphere system. In this figure, the incident solar radiation at the top of the atmosphere of 100 units is equal to $S/4 = 345\ \mathrm{Wm^{-2}}$ in the global average model. We shall now consider how this incident solar irradiance varies as a function of latitude, time of year or time of day, noting that the solar constant *S* does not change.

Variation of radiation fluxes with latitude

The reason for the lower temperatures at higher latitudes is the reduction of net incident solar radiation at the surface. The main reason for this is the variation of solar irradiance with zenith angle (the angle that the sun is away from the local vertical), but there are two other important factors; increasing atmospheric absorption with increasing zenith angle and increased albedo.

Let *I* be the incident solar irradiance on a horizontal surface at the top of the atmosphere. Then $I = S \cos Z$, where *S* is the solar constant and *Z* is the solar zenith angle. The solar zenith angle depends on latitude ϕ, time of year (which determines the solar declination δ) and time of day. At local solar noon (when the sun is nearest to being overhead) and at the equinoxes (when the sun is over the equator), $Z = \phi$ and $\cos Z = \cos \phi$.

More generally, $\cos Z = \cos\phi \cos\delta \cos h + \sin\phi \sin\delta$, where δ, the solar declination, is the latitude at which the noon sun is overhead and *h* is the hour angle of the sun away from solar noon. The solar declination varies from +23.5° on 21 June to -23.5° on 21 December and is associated with the tilt of the Earth's axis away from the vertical.

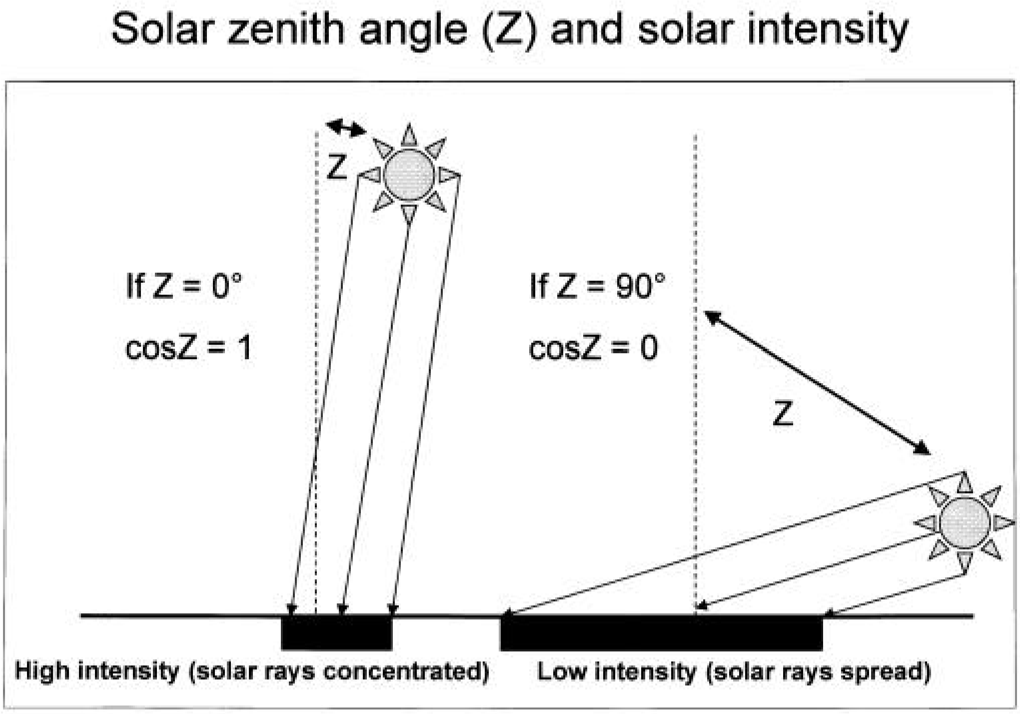

Figure 7.2 Variation of solar irradiance depending on the solar zenith angle.

If we consider the annual average situation, then the incident solar irradiance varies as the cosine of latitude, decreasing rapidly towards the poles. In addition, the absorption of solar radiation increases as the length of the solar path through the atmosphere increases. The solar path increases proportional to 1/cos Z , so at high latitudes there is more absorption than at low latitudes, as the solar path through the atmosphere is longer. Finally, at high latitudes, the average albedo is higher because of permanent snow and ice cover. Hence, the net solar radiation at high latitudes is much less than at low latitudes.

In fact, as shown in Fig. 7.3, radiative equilibrium does not exist at high latitudes or at low latitudes. At low latitudes, the net solar radiation is greater than the outgoing long-wave radiation, implying a net heating due to radiation. At high latitudes, the net solar radiation is less than the outgoing long-wave radiation, even though the atmospheric temperature is lower, implying a net cooling due to radiation. In the next section, we will see that it is the heat transport by the atmospheric circulation and the oceanic circulation that balances the radiation and leads to negligible net heating or cooling.

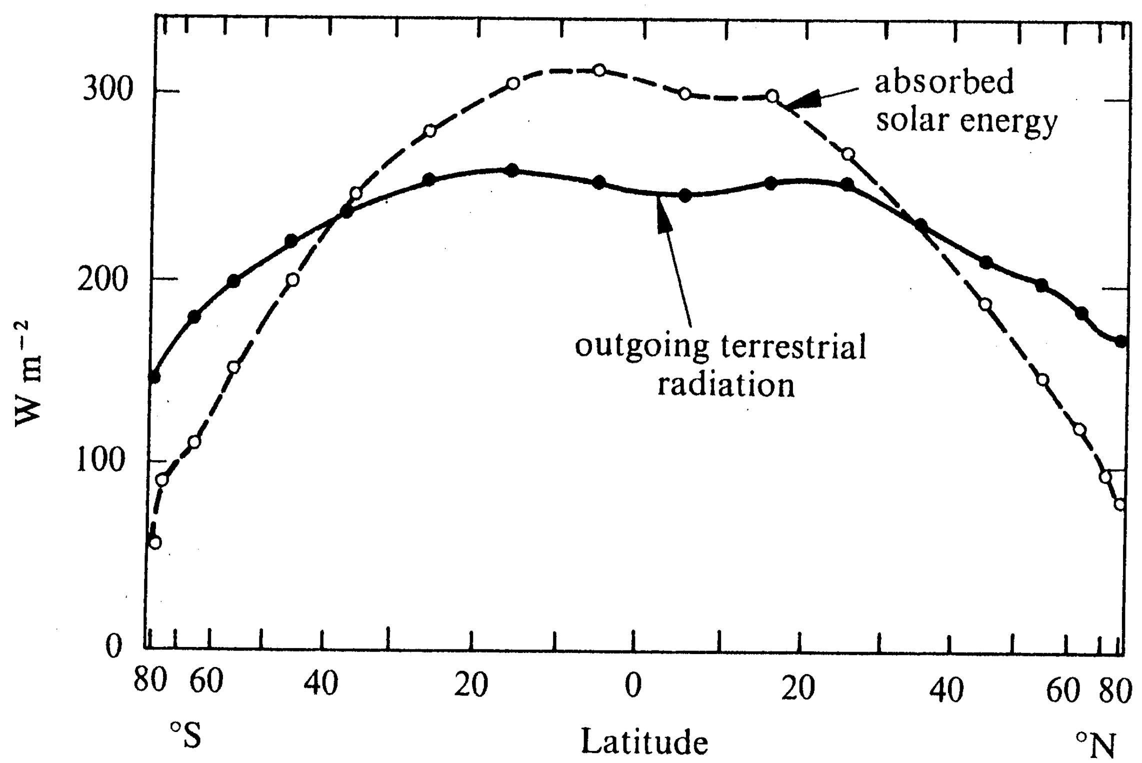

Figure 7.3 Variation of mean annual net solar and long-wave (terrestrial) radiation at the top of the atmosphere. (From Houghton, 1977)

To estimate the rate of cooling of the atmosphere associated with a radiative imbalance, we consider the heat equation

$$r c_p \frac{dT}{dt} = Q \, ,$$

where ρ is the atmospheric mass per square metre of the layer being heated (ρ = mass/m^2 = p/g for the layer),
c_p is the specific heat of air, c_p = 1005 J/kg/°C , and
Q is the net heating in Wm^{-2} or Jsec^{-1}m^{-2}.

Let us consider a radiative imbalance Q = -100 Wm^{-2} applied to the whole troposphere, from the surface to 250 hPa. This layer has a pressure of 750 hPa = 7.5×10^4 Pa = 7.5×10^3 kgm^{-2} . Hence, the rate of cooling of the layer would be about –1.3×10^{-5} °C/sec or about –1.1 °C/day. If this rate of cooling was to exist for even a month, it would lead to a catastrophic climate change. In practice, the cooling is offset by heat transport in the atmosphere and ocean.

Seasonal and diurnal variations of radiation

The seasons are caused by the Earth's motion around the sun and the tilt of the Earth's axis of rotation, as shown in Fig. 7.4. As the Earth orbits around the Sun and rotates about its axis, the axis points in the same direction in space, at an angle of about 23.5° to the orbital plane. On 21 December, the SH summer solstice, the South Pole is oriented towards the sun and the North Pole is in darkness. On 21 June, the SH winter solstice, the North Pole is towards the sun and the South Pole receives no solar radiation. In addition to the solar declination effects on the incident solar radiation, the length of day increases in summer so that the total solar insolation is almost constant from the equator to the South Pole at the summer solstice, as shown in Fig 7.5. The seasonal variations in net solar radiation are much greater than the variations in temperature or outgoing long-wave radiation, so there is net heating in summer and net cooling in winter. These temperature changes will continue until the outgoing long-wave radiation balances the net incident solar radiation. Hence, the maximum temperature in summer does not occur at the time of maximum solar radiation (summer solstice) but about a month later, as shown in Fig 7.6.

In practice, to calculate the actual heating or cooling at the surface, we should consider the surface energy balance and the heat capacity of the surface. The seasonal range of temperatures is much greater over land than over the oceans because the effective heat capacity over the ocean is greater than over the land. The heat input at the surface of the ocean is mixed over a layer of about 50m – 100m in depth, leading to a very large effective heat capacity. In contrast, the heat input to the surface over the land is conducted only slowly over a depth of order 10cm – 1m, so the effective heat capacity is much less and the temperature changes much larger.

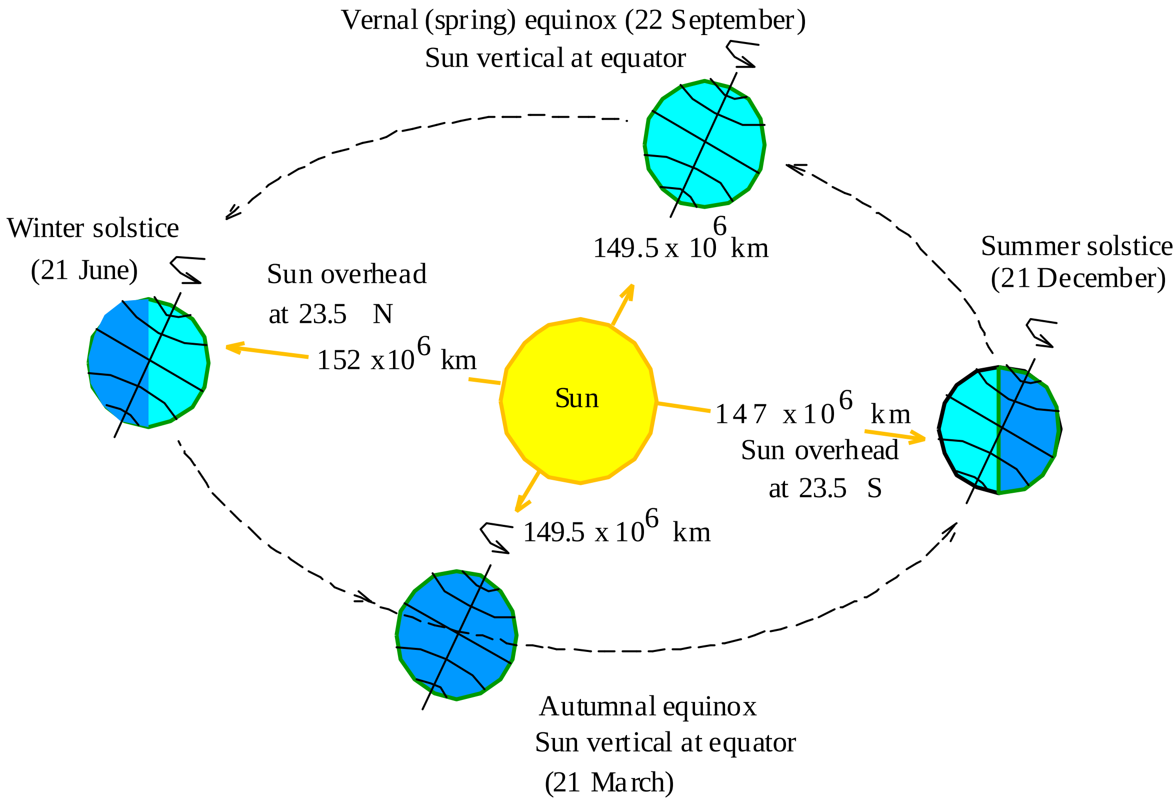

Figure 7.4 Daily and seasonal variations in solar radiation associated with the Earth's orbit around the sun and the tilt of its axis of rotation (from Sturman and Tapper, 1996)

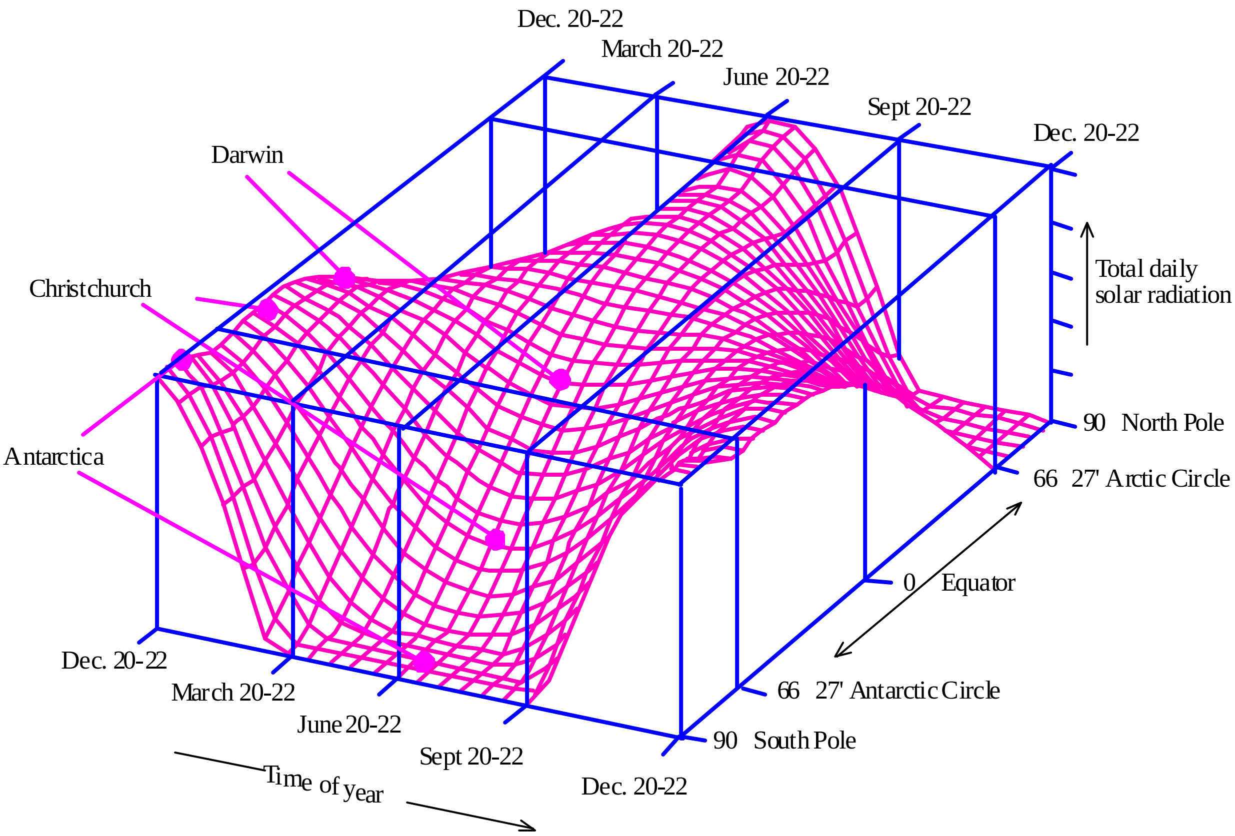

Figure 7.5 Seasonal variations in total daily solar insolation at the top of the Earth's atmosphere (from Sturman and Tapper, 1996)

The temperature variations associated with the diurnal variations of radiation can be estimated in a similar way. At night, there is no incident solar radiation and the radiation balance is between the downward long-wave radiation from the atmosphere and the upward long-wave radiation from the surface. During the day, the additional incident solar radiation causes a net heating at the surface. As for the seasonal cycle, the daytime maximum temperature does not occur at solar noon, but later in the day, when the surface has warmed further and the net outward long-wave radiation balances the incident solar radiation. Similarly, the night time minimum temperature occurs close to dawn. Again, the diurnal variations of temperature are much greater over land than over ocean. Also, the nocturnal cooling is less on cloudy nights, as the cloud provides an effective downward radiative emitter from warmer regions lower in the atmosphere. On clear nights, the effective long-wave emission comes from higher in the atmosphere, where the temperature is colder.

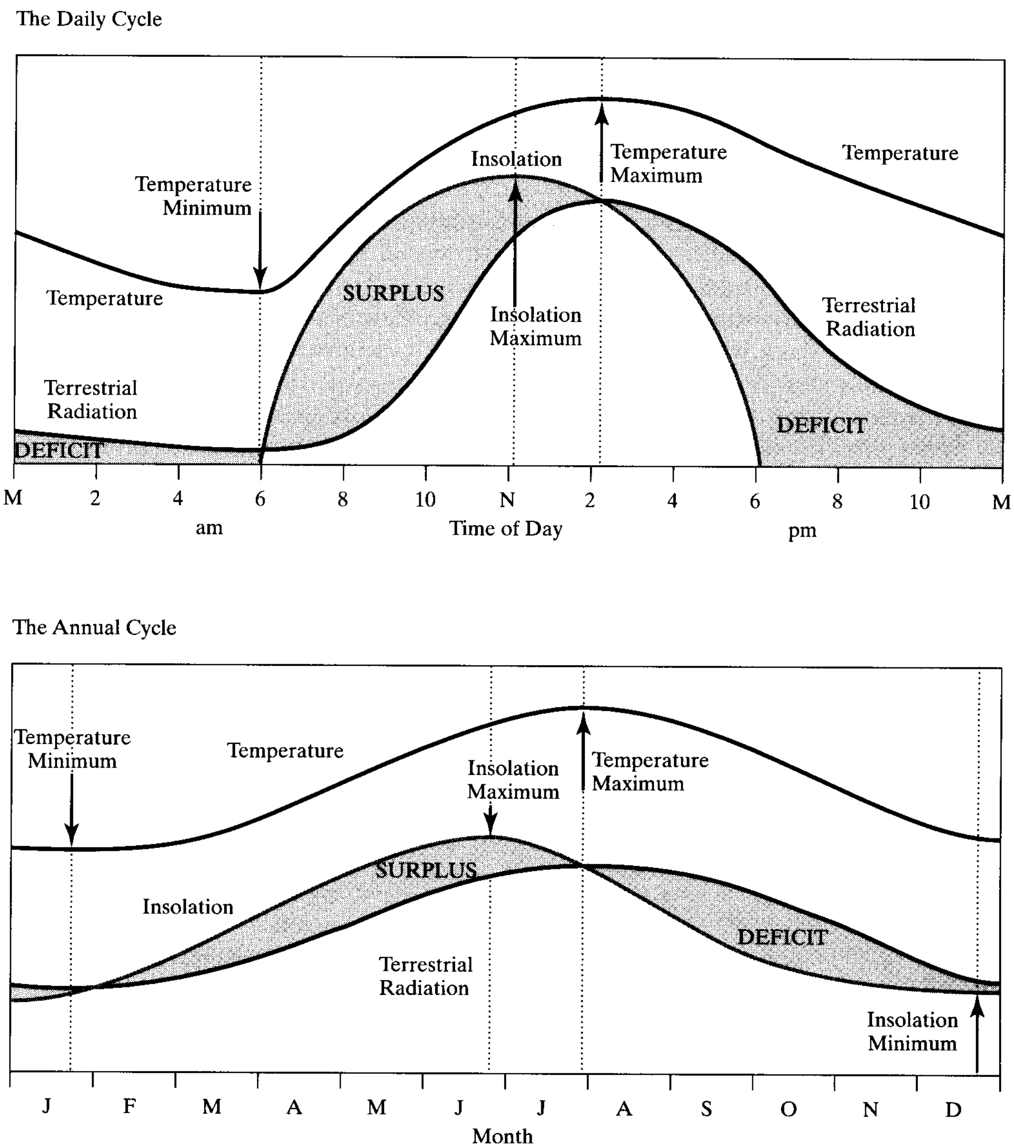

Figure 7.6 Diurnal and seasonal variations of solar insolation and long-wave radiation and of temperature. (From Anthes, 1997)

8. General Circulation of the Atmosphere

In the previous section, we showed that, on average, there is net radiative heating at low latitudes and net radiative cooling at high latitudes but the temperature trends are much smaller. It is the circulation of the atmosphere and the ocean that transports heat away from the topics and into low latitudes, as shown in Fig 8.1 below. This transport of heat is the main reason for the patterns of mean circulation (winds) in the atmosphere.

Sailors and scientists have been observing the mean wind patterns for many years, and the earliest theories for the atmospheric general circulation were developed in the 1700's and 1800's. We will describe and explain the large-scale flow of the atmosphere, usually averaged over many months or years.

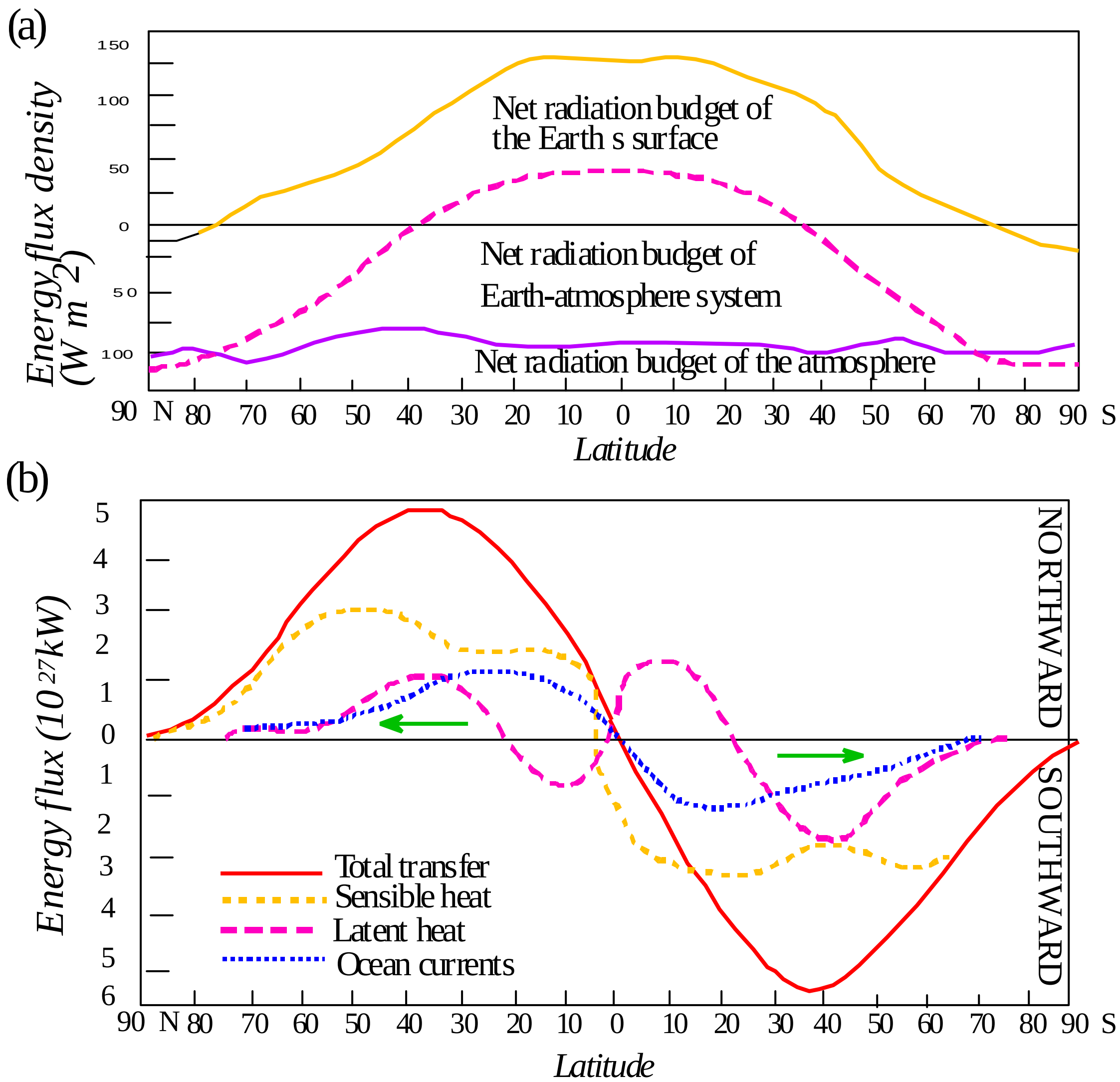

Figure 8.1 Components of the latitudinal radiation balance and heat transport.
(a) Annual average latitudinal variations in net radiation budget for the Earth's surface, the atmosphere, and the whole Earth-atmosphere system.
(b) Annual average latitudinal distribution of the northward energy transport in the ocean-atmosphere system. (From Sturman and Tapper, 1996)

Zonal average circulation

As before, we use the term zonal wind to mean the flow from west to east (westerly flow) and the meridional wind to mean the flow from south to north (southerly flow). We also define the zonal average [ ] to be the average of a quantity around a latitude circle ie the zonal average zonal wind is

$$[u]=[u(j,z,t)]=\frac{1}{2p}\int u(l,j,z,t)dl=\frac{1}{N}\sum_{i=1}^{N}u(l_i,j,z,t).$$

The zonal average zonal wind and temperature are show in Figure 8.2 for typical January and July conditions in the Northern Hemisphere. As expected, the temperature at low latitudes is greater than at high latitudes. In addition, the mean zonal flow increases with height, in thermal wind balance with the temperature decrease towards the pole.

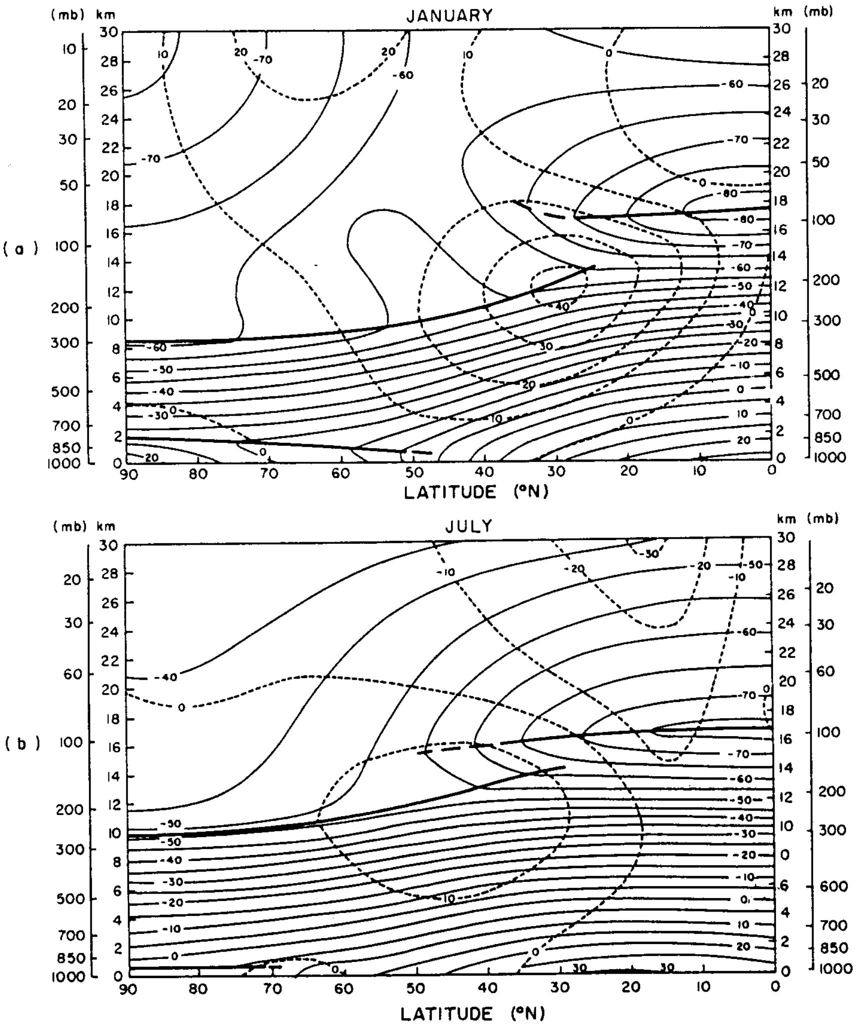

Figure 8.2 (Reproduced from Fig 6.1 earlier) Meridional cross-section of zonal-average temperature and zonal wind is (a) January and (b) July. (From Wallace and Hobbs, 1977.)

In a similar way to the definition of the zonal average, we define the time-average or time-mean of a quantity $\overline{(x)}$ to be its average over a given interval, such as a month, a season, or many years. Hence the time-average zonal flow is

$$\bar{u} = \overline{u(l,j,z)} = \frac{1}{T}\int u(l,j,z,t)dt = \frac{1}{N}\sum_{n=1}^{N} u(l,j,z,t_n).$$

Fig 8.2 shows the time-average zonal mean zonal wind $\overline{[u]}$ and temperature $\overline{[T]}$ in January and July

Now, let us consider the explanation for the general circulation. If we consider the simplest circulation in response to net heating at low latitudes and net cooling at high latitudes, we would expect rising motion at low latitudes and sinking motion at high latitudes, joined by a single large overturning circulation cell. This was first described by Hadley in the 1700's and is shown in Fig. 8.3 below. At low latitudes, adiabatic cooling associated with ascent and transport of heat away from the region balances the net radiative heating. At high latitudes, adiabatic warming associated with descent and transport of heat into the region balances the net radiative cooling.

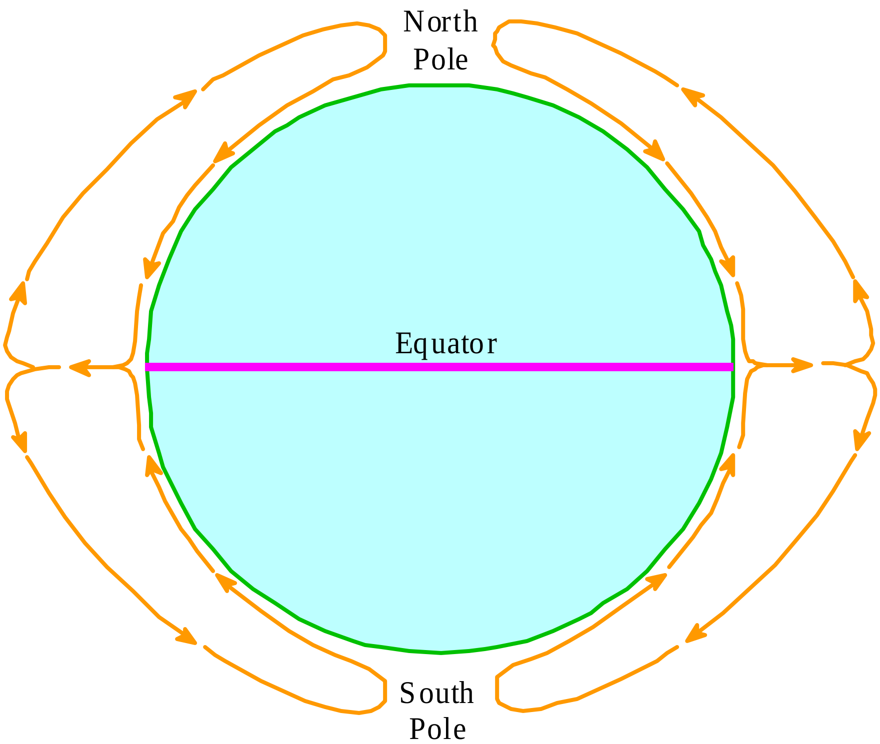

Figure 8.3 Idealised overturning circulation in response to heating at low latitudes and cooling at high latitudes on a non-rotating Earth (from Sturman and Tapper, 1996).

We can use conservation of angular momentum to help explain the pattern of zonal flow associated with this pattern of meridional circulation. Since the Earth is rotating around its axis once per day, an air parcel with no motion relative to the Earth's surface would be rotating around the Earth's surface with angular velocity $\Omega = 7.3\times10^{-5}\ s^{-1}$. The angular momentum is the mass of the air parcel times the distance from the axis of rotation times the tangential velocity, mRV_t. The distance from the axis of rotation depends on latitude, $R = a\cos f$, where a is the Earth's radius, a=6370 km. The tangential velocity is just the speed of rotation of the Earth's surface plus the zonal wind speed, $V_t = \Omega a\cos f + u$. Hence the angular momentum is $ma\cos f(\Omega a\cos f + u)$. Angular momentum is conserved for frictionless motion. If an air parcel, initially at the equator with no motion relative to the Earth's surface, is moved to higher latitudes then its angular momentum will remain constant at its initial value $ma^2\Omega$. Hence, the zonal wind at any latitude can be obtained by solving

$$ma\cos f(\Omega a\cos f + u) = ma^2\Omega\ .$$

Hence, $u = a\Omega(\frac{1}{\cos f} - \cos f)$ and the zonal wind would increase rapidly from the equator, to 55 m/s at 20° and 130 m/s at 30°. This helps to explain the increase of zonal wind in the poleward branch of the Hadley circulation but the observed jets are weaker than expected from conservation of angular momentum. Equatorward flow in the lower branch of the Hadley circulation would also conserve its angular momentum, apart from frictional losses. Hence, air parcels at rest at latitudes away from the equator would move westward as they approach the equator, explaining the surface easterlies in the tropics.

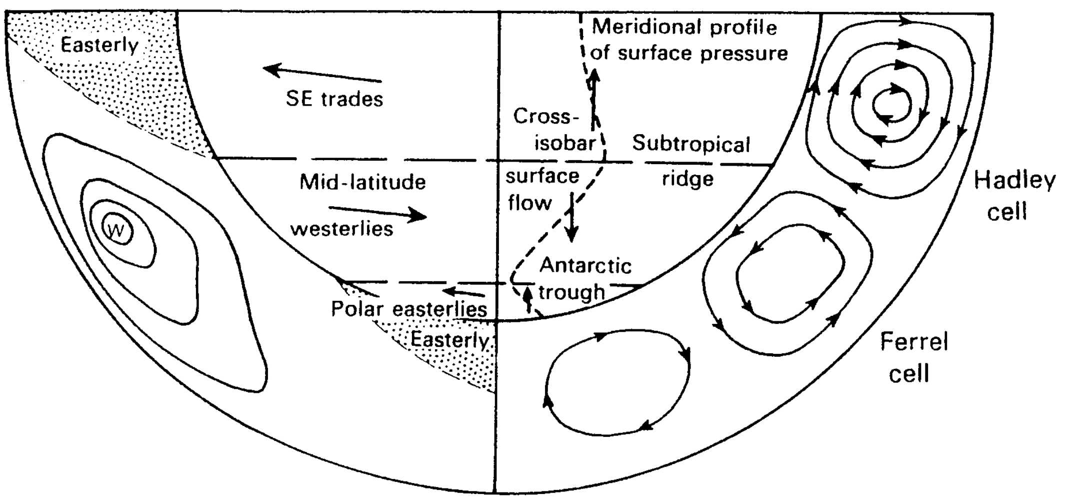

Figure 8.4 Schematic representation of the main features of the general circulation in the Southern Hemisphere, showing the zonal mean zonal flow (left) and the mean meridional circulation (right). (From Pittock et al., 1978)

We know from the observed upper air wind patterns that the zonal wind does not increase in strength all the way to the pole and the mean Hadley circulation does not extend from the equator to high latitudes. In fact, the Hadley cell terminates in the subtropics, just on the equatorward side of the latitude of the subtropical jet maximum.

Fig. 8.4 shows the observed mean pattern of zonal wind and meridional circulation in the Southern Hemisphere. In mid-latitudes, the mean meridional circulation shows a reversed cell, called the Ferrel cell, which is a thermally-indirect circulation ie air in warmer regions descending and air in cooler regions rising. Such a circulation cannot transport heat poleward. While the Hadley circulation is able to transport heat poleward, the mean meridional circulation in the Ferrel cell cannot provide a net poleward transport of heat.

Available Potential Energy

Another way to describe the general circulation is in terms of processes that convert potential (stored) energy into kinetic energy (associated with atmospheric motion). The mean temperature gradient between high and low latitudes means that the mean density surfaces in the atmosphere are not flat; they slope upwards to the poles, with higher density, colder air at high latitudes. These sloping density surfaces have more potential energy than the equivalent horizontal density surfaces, and this is called the available potential energy. Any process that can flatten the sloping density surfaces will release this potential energy and convert it to kinetic energy of atmospheric motion.

In a non-rotating atmosphere, the mean meridional overturning circulation, like the Hadley circulation, can flatten the density surfaces. The available potential energy is converted into kinetic energy of the mean meridional circulation. Hence, this can occur in low latitudes, where the effects of rotation are weak. The Hadley circulation only exists in low latitudes.

In middle latitudes, however, the potential energy associated with the mean meridional temperature gradient is not transformed into kinetic energy through a mean meridional overturning circulation (which would occur if the Earth was not rotating). In middle latitudes, the increasing wind with height, which is in thermal wind balance with the meridional temperature gradient, is unstable to small disturbances. These can grow into weather systems, the highs and lows that we see on weather maps. The mechanism that produces the growing weather systems is beyond the scope of this course, but is called “baroclinic instability”. The potential energy associated with the mean meridional temperature gradient is converted into the kinetic energy of the transient weather systems. In middle latitudes, it is the movement of air in transient weather systems that provides a net poleward transport of heat, and not the mean meridional circulation. These mean transport of heat acts to reduce the mean temperature gradient and to flatten the density gradient, reducing the available potential energy.

So why doesn’t all the available potential energy disappear when the density surfaces have been made horizontal by the atmospheric motions? There is continual production of available potential energy associated with the net radiative heating at low latitudes and the net radiative cooling at low latitudes, which maintains the temperature and density gradient between low and high latitudes. Hence, the contrast in solar heating between high and low latitudes is the main forcing of the general circulation and the atmosphere responds to this. The large-scale circulation is constrained by conservation requirements, such as conservation of angular momentum and total energy, and by balance requirements, such as geostrophic balance and thermal wind balance.

Eddy Transport

So how can the variations of wind and temperature in weather systems provide a net poleward transport of heat when there is no mean meridional circulation? To understand this, we must consider the separate contributions to the poleward transport of heat from mean motion and from waves or weather systems.

Let us consider the mean northward transport of heat in the atmosphere, which is the heat content of a cubic metre of air $r c_p T$ J/m^3 multiplied by the rate of transport of air across a latitude circle, v m^3/m^2/s or v m/s. Hence, the northward transport of heat is $r c_p v T$ J/m^2/s.

We now consider the zonal mean transport $[rc_p vT] = r c_p [vT]$. We can represent any quantity, such as temperature T as the sum of its zonal mean $[T]$ and the departures from the zonal mean, called the eddy or wave components, $T^* = T - [T]$. Hence, we can write $T = T^* + [T]$ and $v = v^* + [v]$, so

$$[vT] = [(v^* + [v])(T^* + [T])]$$

$$= [v^* T^* + v^*[T] + [v]T^* + [v][T]]$$

$$= [v^* T^*] + [v^*[T]] + [[v]T^*] + [[v][T]]$$

But $[v^*] = 0$ from its definition, so $[v^*[T]] = 0$ and hence we obtain

$$[vT] = [v^* T^*] + [v][T]$$

A B C

where A is the total zonal mean heat transport (divided by $r c_p$, a constant);
B is the zonal mean heat transport due to eddies (due to correlated departures of meridional wind and temperature from their zonal mean); and
C is the zonal mean heat transport due to the zonal mean meridional circulation.

Note that this applies at any individual time. We can apply the same type of analysis to calculate the contributions to the time average heat transport from the time mean flow and the time variations of the flow.

We can represent any quantity, such as temperature T as the sum of its time mean $\overline{T}$ and the departures from the zonal mean, called the transient or time-varying components, $T' = T - \overline{T}$. Hence, we can write $T = T' + \overline{T}$ and $v = v' + \overline{v}$, with $\overline{v'} = 0$. Hence, the time mean heat transport can be written

$$\overline{vT} = \overline{(v'T')} + \overline{v}\,\overline{T}$$

A B C

where A is the total time mean heat transport (divided by $r c_p$, a constant);
B is the time mean heat transport due to transient variations (due to correlated departures of meridional wind and temperature from their time mean); and
C is the time mean heat transport due to the time mean meridional circulation.

Now, we can show that the time average zonal mean heat transport can be written

$$\overline{[vT]} = \overline{[v^*T^*]} + \overline{[v][T]}$$

$$= \overline{[v'^*T'^*]} + [\overline{v}^*\overline{T}^*] + \overline{[v]'[T]'} + \overline{[v][T]}$$

A B C D

where A is the time mean, zonal mean heat transport due to transient eddies (due to time varying departures of meridional wind and temperature from their zonal mean);
B is the mean heat transport due to standing waves (time mean departures from the zonal mean);.
C is the mean heat transport due to transient variations of the zonal mean flow; and
D is the mean heat transport due to the time mean zonal mean meridional circulation.

Normally, C is negligible and the mean heat transport has three components, due to transient eddies (weather systems), due to standing waves (which we discuss more later), and due to the mean meridional circulation (the Hadley cell or the Ferrel cell).

Fig 8.5 shows the observed variation of the northward transport of heat by the atmospheric circulation, as a function of latitude, and the different contributions from eddies and from the mean meridional circulation. As expected, there is a poleward transport of heat by the atmospheric circulation in both hemispheres, which partially balances the difference in net radiative heating between high latitudes and low latitudes. In low latitudes, the poleward heat transport is predominantly by the mean meridional circulation, by the Hadley cell. In middle latitudes, the mean meridional circulation (the Ferrel cell) transports heat in the wrong direction, from pole to equator, and the transport of heat in the weather systems, the eddies, dominates the poleward heat transport.

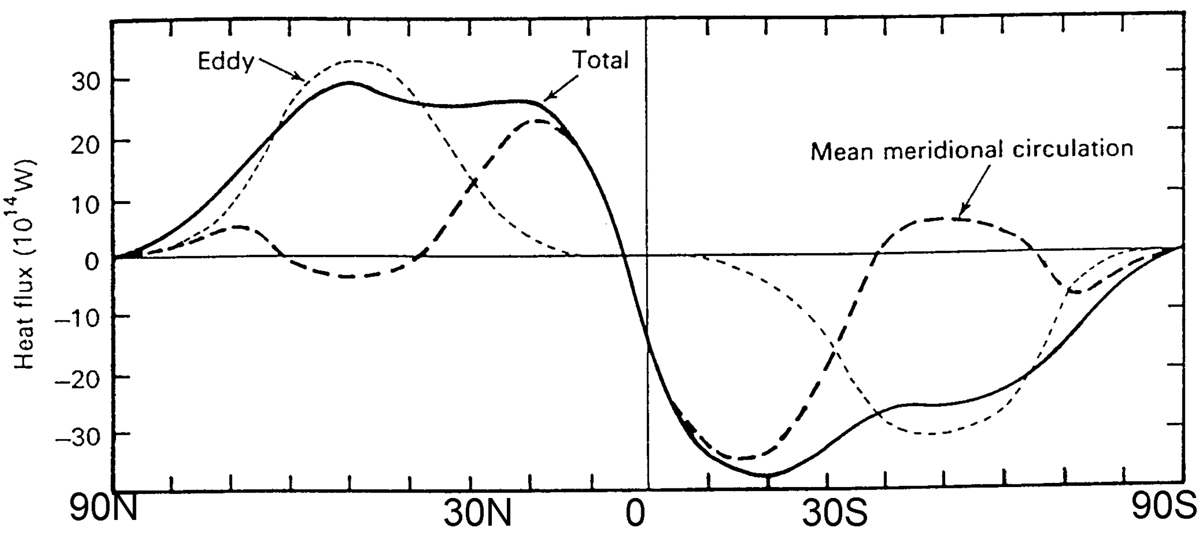

Figure 8.5. Annual mean zonal mean meridional heat transport in the atmosphere, showing the total transport (solid line), eddy transport (thin dashed line) and transport by the mean meridional circulation (thick dashed line). Note that positive transport is northward. From Pittock et al., 1978)

9. Conservation of vorticity and Rossby waves

In the previous section, we used the concept of conservation of angular momentum to deduce the variations of zonal wind associated with the mean meridional circulation away from the equator in the Hadley cell. In practice, this concept was helpful in explaining the existence of the subtropical jets but it did not apply to the mid-latitude jets, because the Hadley cell did not extend into mid-latitudes.

We now apply the concept of angular momentum to rotating air masses, such as high and low pressure systems, to derive another conservation relation. Following section 7.14 of the recommended text (McIlveen, 1992), we consider the large, relatively-flat air masses of weather systems as they rotate about a local vertical axis in their centre. As the horizontal scale of weather systems is large (about 1000 km) relative to their vertical scale (about 10 km), we can consider weather systems as thin rotating disks of air. The angular momentum of the air is $w\,r^2$, where w is the angular velocity (rotation rate) of the air at distance r from the centre of the weather system. In the Southern Hemisphere, w is positive (negative) for a high (low) pressure system and opposite in the Northern Hemisphere.

Conservation of angular momentum means that $w\,r^2$ is constant as the weather system moves. For geostrophic flow, the wind is parallel to the pressure contours and there is no flow into or out of the weather system. Hence, the distance r of an air parcel from the centre of the weather system remains constant, so w must be constant. In practice, friction and other factors can mean that the assumption of purely geostrophic flow is broken. However, the flow is close to geostrophic in middle latitudes away from the surface boundary layer.

In practice, as the Earth is rotating, we must consider the background planetary rotation as well as the rotation of the weather system when considering the conservation of angular momentum in a weather system. The local vertical component of the background planetary rotation at latitude ϕ is $\Omega\sin\phi$, as shown in Fig. 9.1. This must be added to the local angular velocity of the weather system, so $w+\Omega\sin\phi$ is constant as an air parcel moves in the weather system.

To estimate the local rotation rate in a weather system, we use the vertical component of the vorticity $z = (\nabla \times \boldsymbol{u}).\boldsymbol{k} = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}$, which is equal to twice the local angular velocity, $z = 2w$. The Coriolis parameter $f = 2\Omega \sin f$ is the vorticity associated with the background planetary rotation, called the planetary vorticity. The total vorticity of the flow, $z + f$, is constant for geostrophic flow. For cyclonic flow ie a low pressure system, the vorticity z has the same sign as the planetary vorticity, f ie positive in the NH, negative in the SH.

We can use the conservation of vorticity of the flow to help explain the existence of large-scale waves in the atmosphere, which are often called Rossby waves. They owe their existence to the variation of the planetary vorticity with latitude. Consider straight-line flow, with zero initial vorticity, in a south-eastward direction at some latitude in the Southern

Hemisphere. Let the planetary vorticity at this initial latitude be $f_0 < 0$. As the air moves southward, it enters a region with more negative planetary vorticity $f < f_0$. To maintain the total vorticity, $z + f$ constant, the vorticity of the flow must increase, $z > 0$, which means an anticlockwise turning of the flow. The flow will turn anticlockwise and return to its original latitude, this time in a north-eastward direction. It will overshoot its original latitude, as shown in Fig 9.2, and enter a region where the planetary vorticity $f > f_0$. To maintain the total vorticity constant, this time the vorticity of the flow must decrease, $z < 0$, which means a clockwise turning of the flow, which will again return the flow to its original latitude. This generates the wavey flow shown in Fig 9.2, which is due to the conservation of total vorticity and the variation of the planetary vorticity with latitude. These are Rossby waves, which can be seen in the atmospheric circulation in middle latitudes almost all the time.

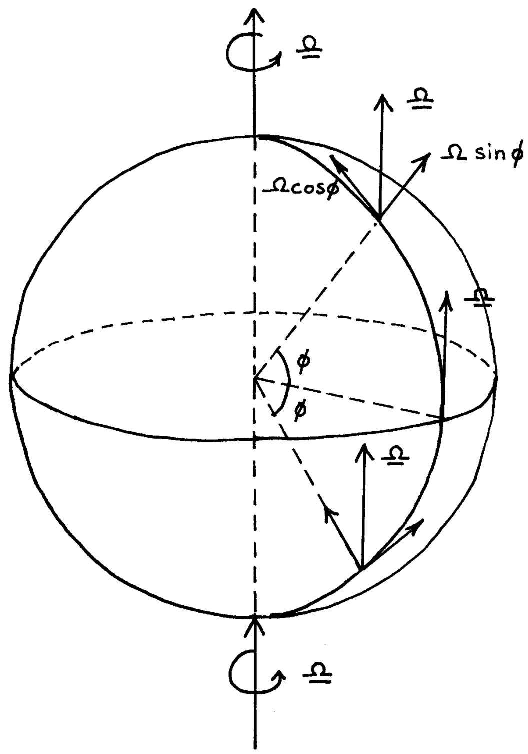

Figure 9.1. The variation of the horizontal and vertical components of the Earth's angular velocity with latitude.

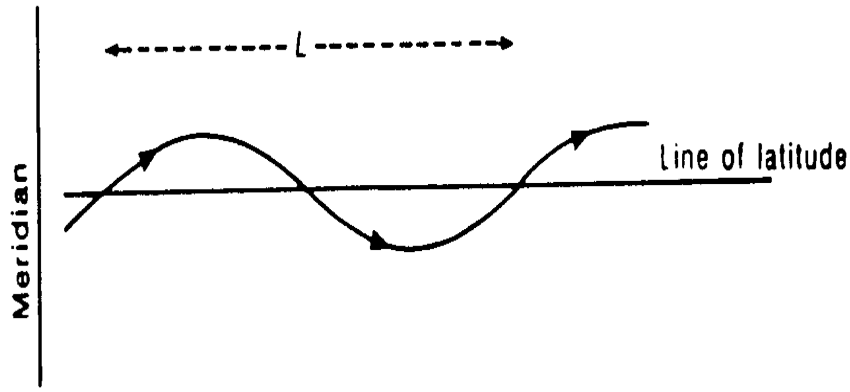

Figure 9.2 Schematic view of wave-like flow around a line of latitude