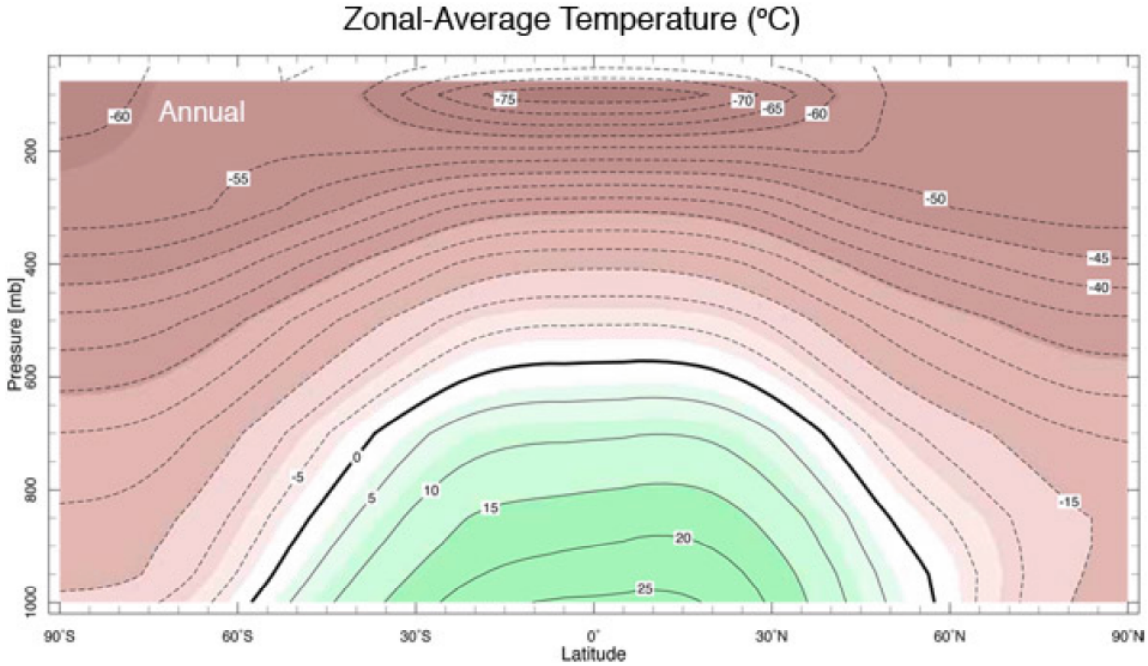
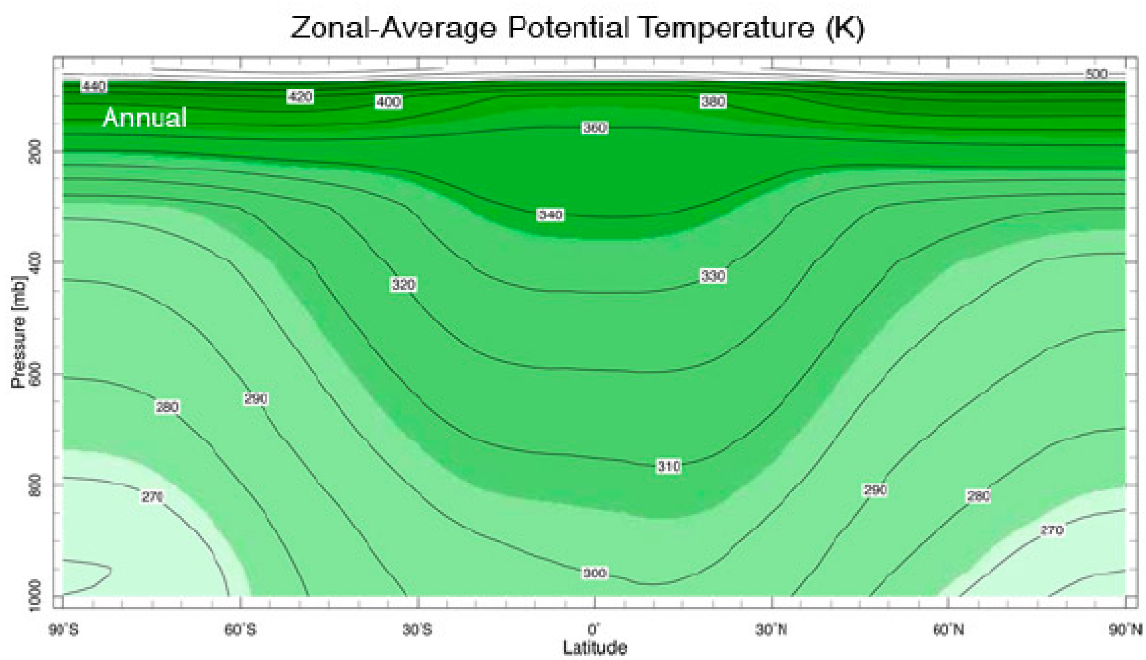
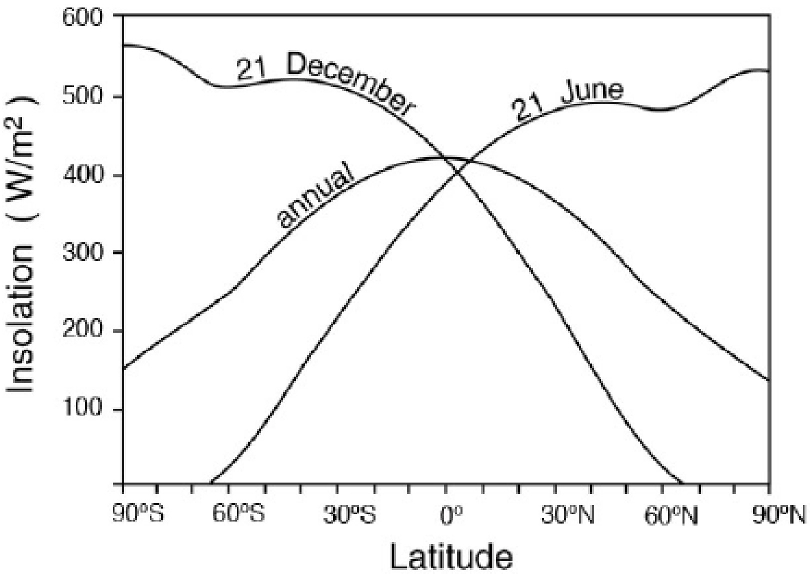
The general circulation of the atmosphere

We will now see how much of the general circulation of the atmosphere that we can understand based what we have learned so far.





The two figures above are easy to understand. We started with the radiative-convective equilibrium and see the tropics are warmer than the poles. This is easy to explain given the insolation. Note that insolation is the highest over the poles during solstices, but high albedo, small greenhouse effect and finite thermal capacity prevent the poles from becoming warmer than the tropics. We also understand why potential temperature increases with height: moist convection tends to adjust the atmosphere toward a moist adiabatic profile, particularly in the tropics.



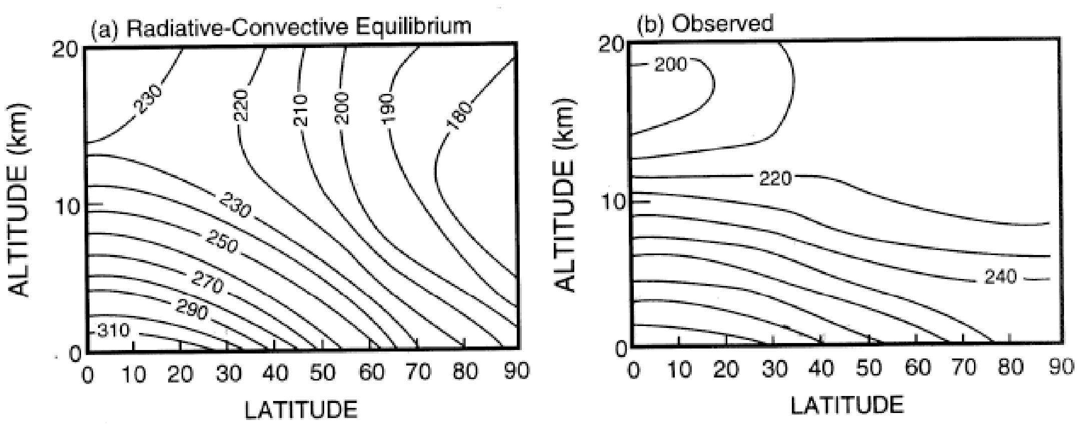


Figure 10.1 Zonal–mean temperature as a function of latitude and height (a) under radiative–convective equilibrium and (b) observed during northern winter. Without horizontal heat transfer, radiative–convective equilibrium establishes a meridional temperature gradient that is much stronger than observed. Sources: Liou (1990) and Fleming *et al.* (1988).

So question is more why the observed equator-to-pole temperature difference is smaller than that of radiative-convective equilibrium. Now if the atmosphere is in hydrostatic equilibrium, there must be horizontal pressure gradient. This will drive a circulation with air converging to the tropics at the bottom and diverging at the top. This gives rise to the Hadley cell, which transports heat poleward.

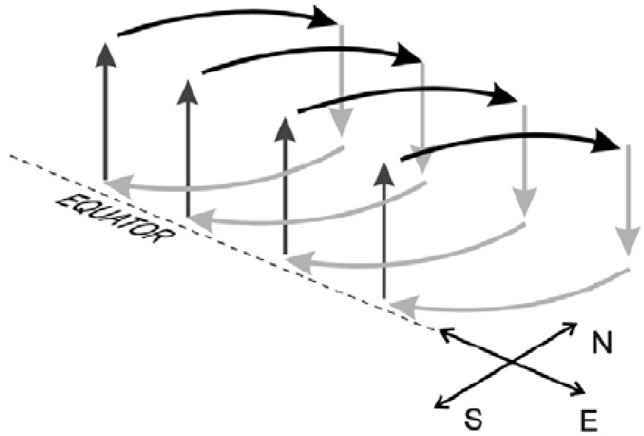
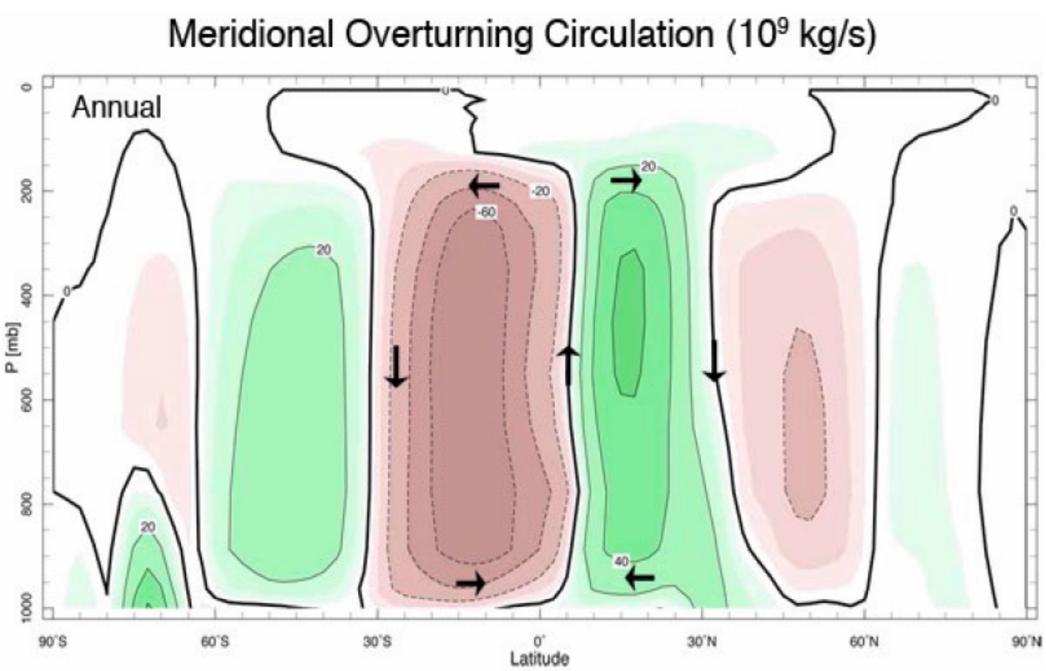


Figure 8.4: Schematic of the Hadley circulation (showing only the Northern Hemispheric part of the circulation; there is a mirror image circulation south of the equator). Upper level poleward flow induces westerlies; low level equatorward flow induces easterlies



Note the circulation cells in the extratropics (The Ferrel cell) go the other way.

Now the question is why the Hadley cells do not go all the way to the pole? And why the Ferrel cells rise at higher latitudes, which are colder and sink at lower latitudes?

At the upper level, as air flows poleward, it turns toward the east under the Coriolis force, while the opposite is true at the surface. In the subtropics, the strong subsidence forms the trade inversion and caps boundary layer turbulence. Below the trade inversion, effects of friction are strong and winds are in general weak. Above the trade inversion, frictional effects are weak, thermal wind balance is satisfied through wave adjustment. This explains the subtropical jets.

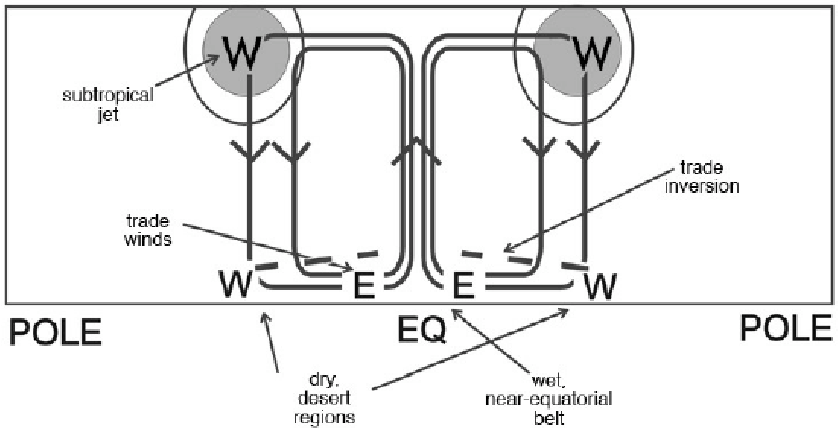
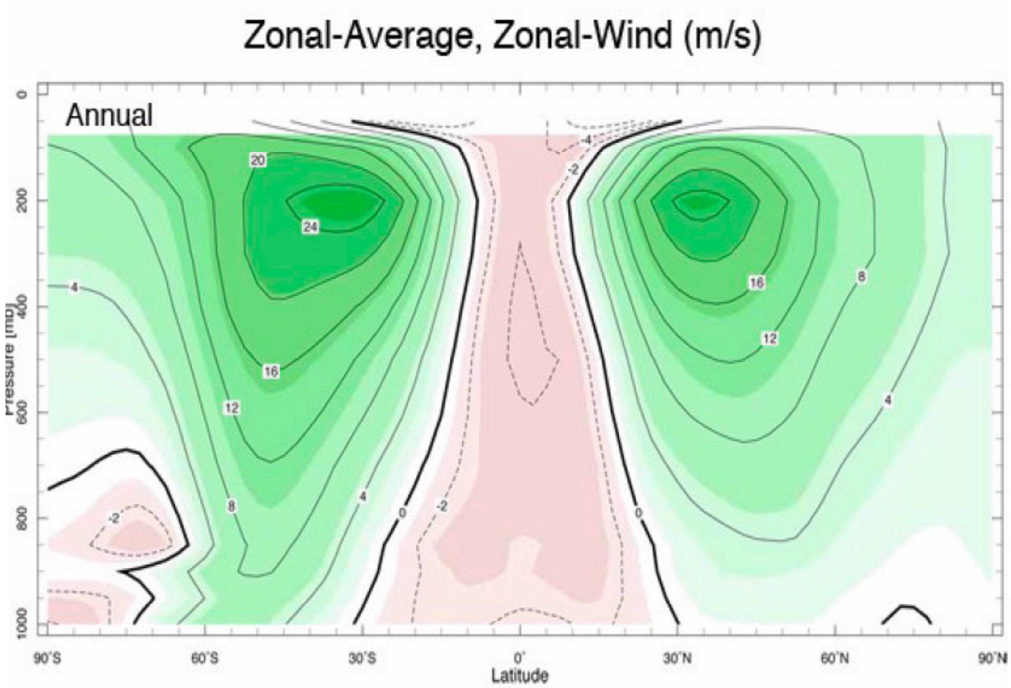


Figure 8.5: A schematic diagram of the Hadley circulation and its associated zonal flows and surface circulation.



Surface winds however are a different matter. We saw earlier that given the surface pressure distribution, we can deduce the surface wind distribution based on balance among pressure gradient force, Coriolis force, and friction. But we don't know the surface pressure distribution a priori. Why cannot we have a situation where surface pressure is uniform everywhere so that there are no winds? Indeed, when there are surface westerlies, surface friction imparts westward momentum onto the air column and reduces the westerlies. How are the westerlies maintained?

The humidity distribution is more readily understood. It to a large extent follows the saturation specific humidity distribution, which is mostly determined by temperature; the departure can be understood based on the circulation. In regions of subsidence, the relative humidity tends to be low.

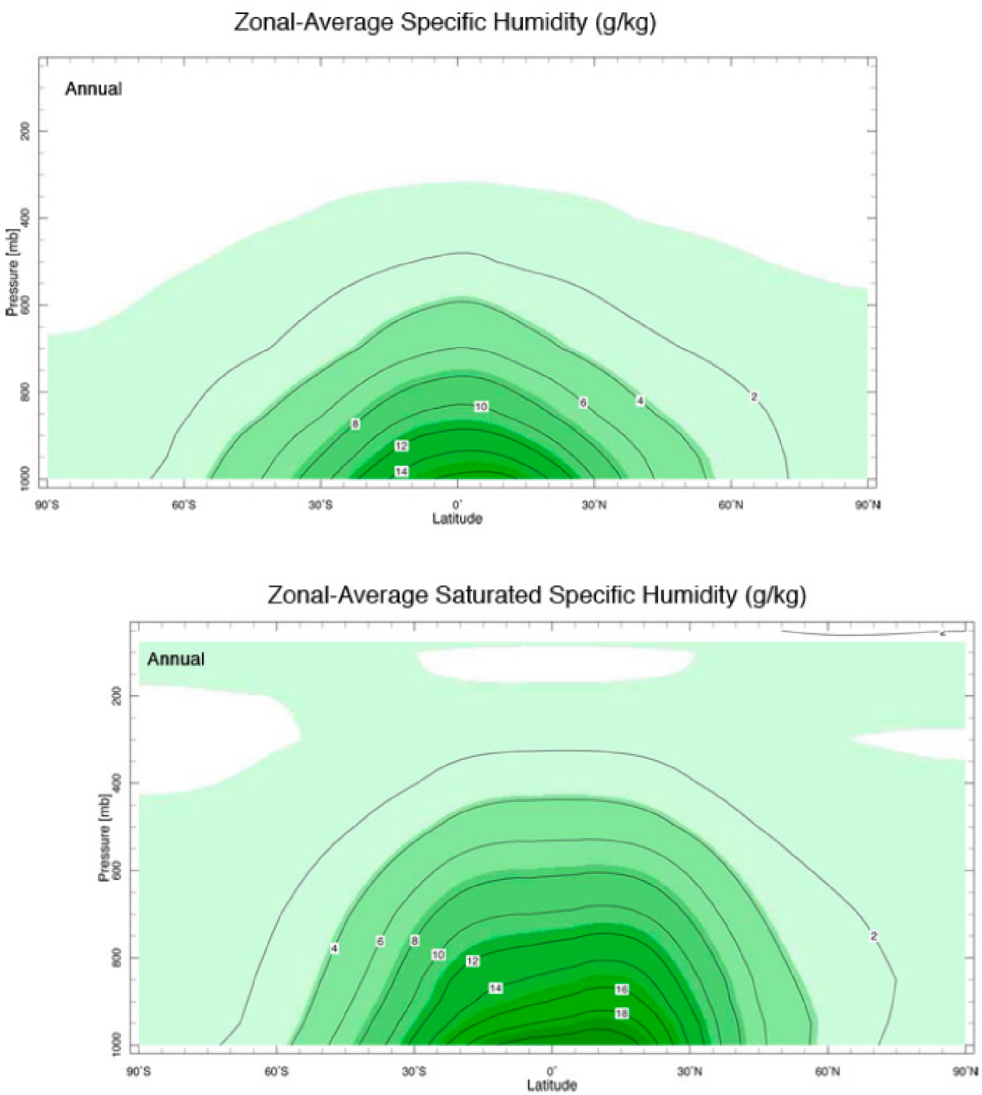


Figure 5.16: Zonally-averaged saturated specific humidity, q_*, in $\mathrm{g\,kg^{-1}}$, for annual-mean conditions.

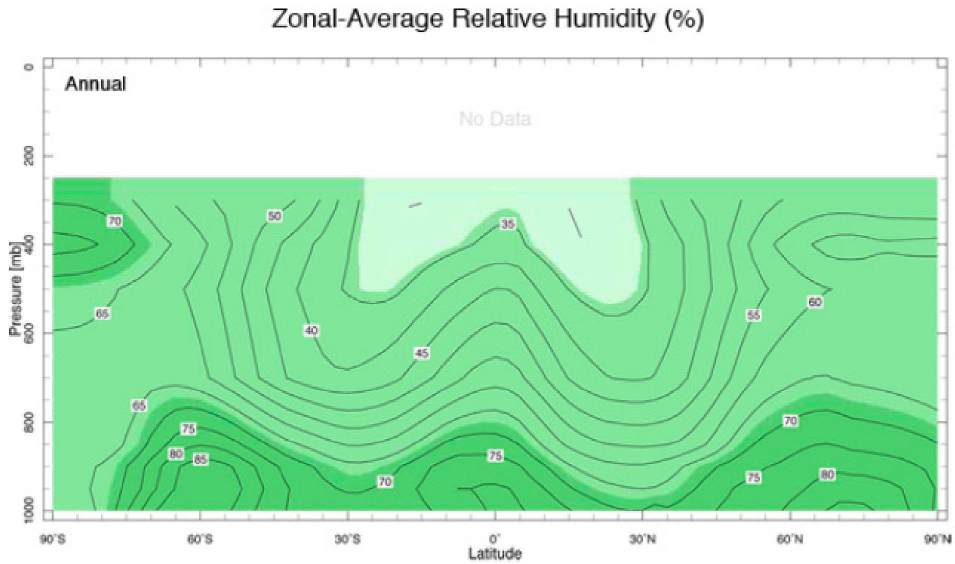
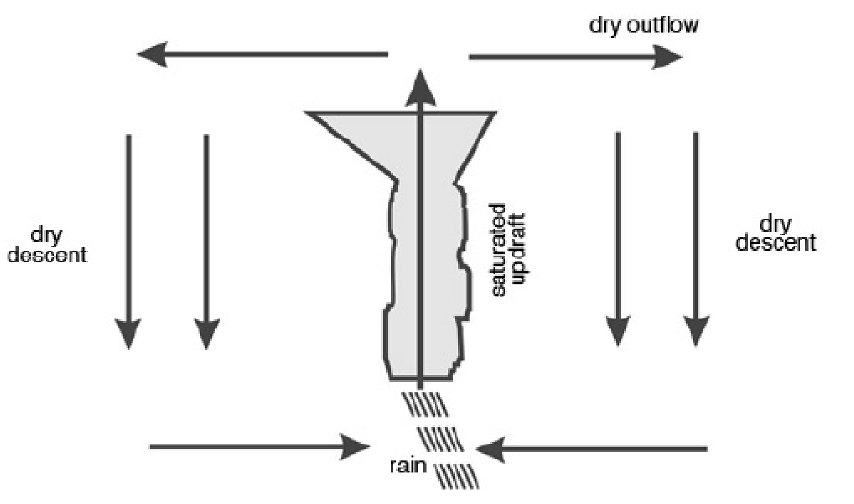


Figure 5.17: Zonal mean relative humidity (%), Eq.(4.25), under annual mean conditions. Note that data is not plotted above 300mb where q is so small that it is difficult to measure accurately by routine measurements.



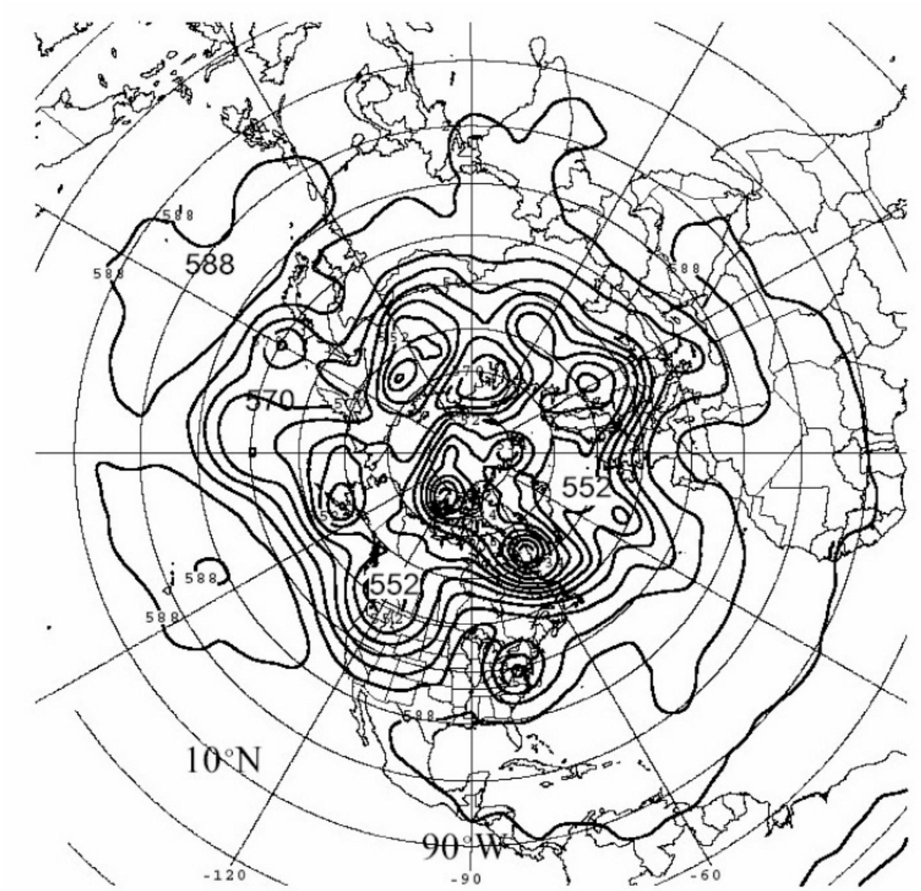


Figure 5.22: Typical 500 mbar height analysis: the height of the 500 mbar surface (in decameters) at 12 GMT on June 21st, 2003. The contour interval is 6 decametres = 60 m. The minimum height is 516 decametres, and occurs in the intense lows over the pole.

The real atmosphere is full of eddies. Where do they come from? We have so far only considered zonally symmetric conditions. Could the cause be the zonally asymmetric distribution of land/ocean and orography etc.? There is some truth to this in terms of long term averages and may be checked by comparing the northern hemisphere and the southern hemisphere.

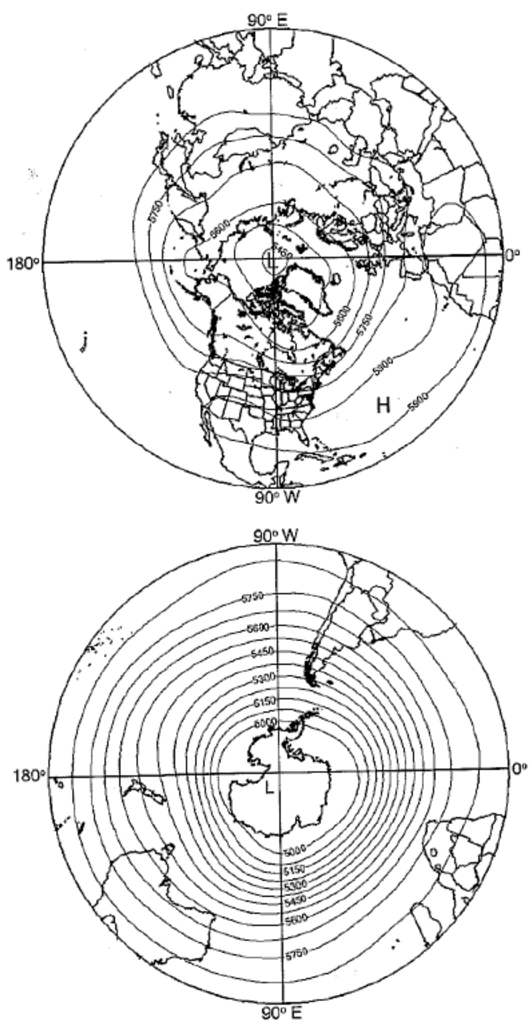


Figure 7.3 The mean contours (gpm) of the 500-mb pressure surface in July for the northern and southern hemispheres, 1970 to 1999.

Source: NCEP/NCAR Reanalysis Data from the NOAA-CIRES Climate Diagnostics Center.

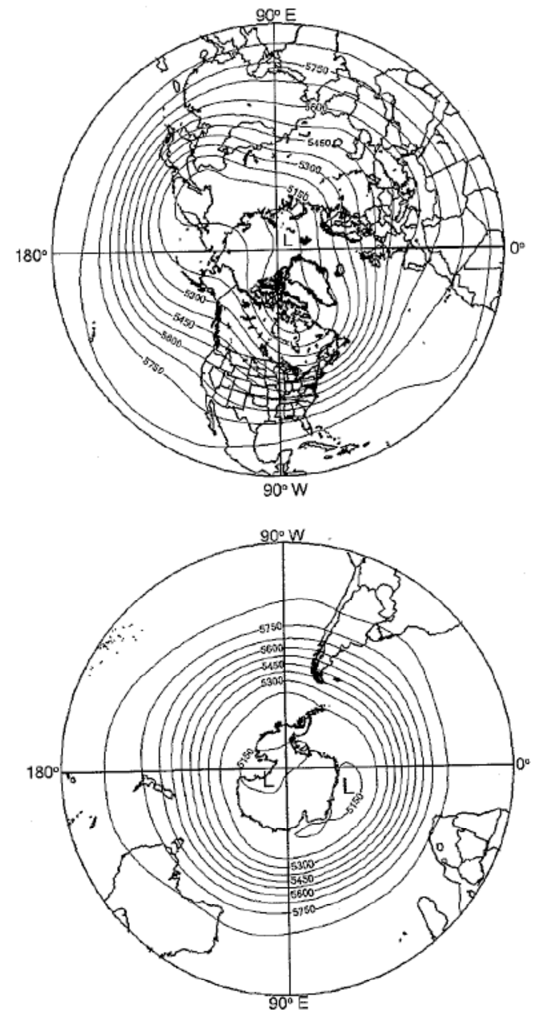


Figure 7.4 The mean contours (gpm) of the 500-mb pressure surface in January for the northern and southern hemispheres, 1970 to 1999.

Source: NCEP/NCAR Reanalysis Data from the NOAA-CIRES Climate Diagnostics Center.

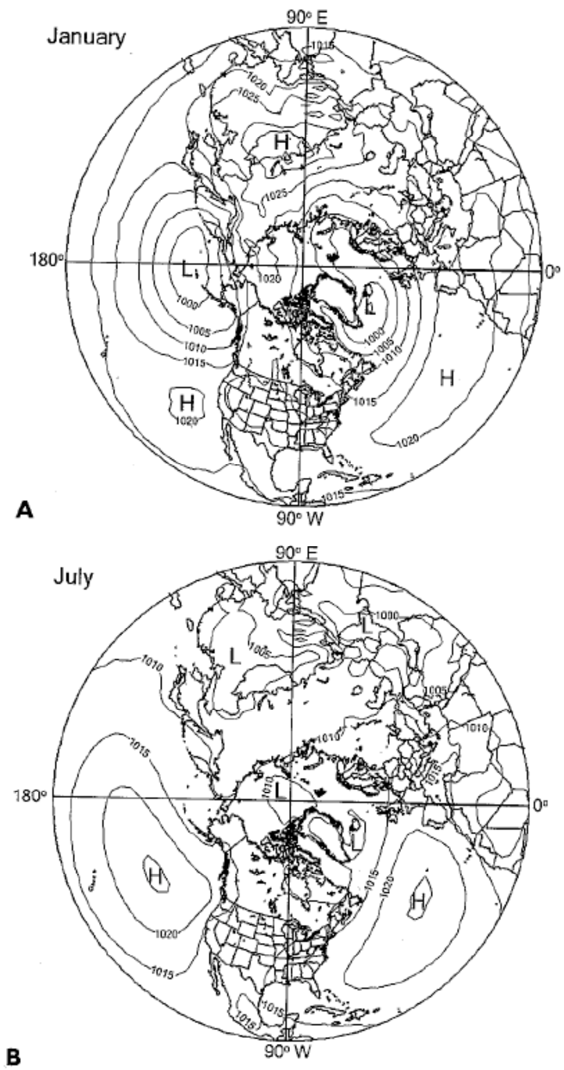


Figure 7.9 The mean sea-level pressure distribution (mb) in January and July for the northern hemisphere, 1970 to 1999.

Source: NCEP/NCAR Reanalysis Data from the NOAA-CIRES Climate Diagnostics Center.

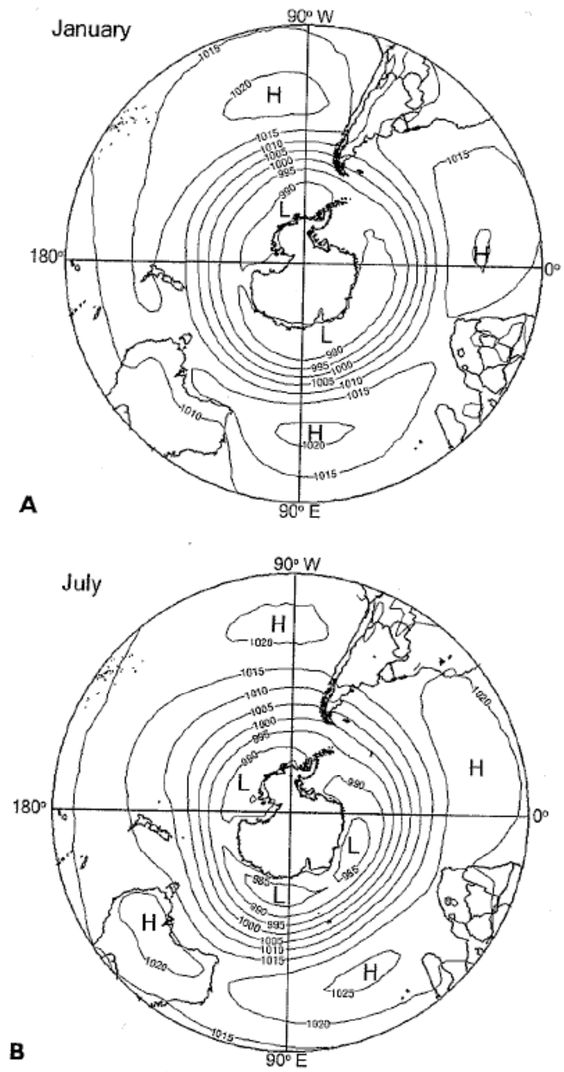


Figure 7.10 The mean sea-level pressure distribution (mb) in January and July for the southern hemisphere, 1970 to 1999. Isobars not plotted over the Antarctic ice sheet.

Source: NCEP/NCAR Reanalysis Data from the NOAA-CIRES Climate Diagnostics Center.

So in the mean, the zonal asymmetry in the NH is because of topography and associated zonal asymmetric in heating (to be explained later when we discuss Rossby waves) and the Southern Hemisphere is a lot more zonally symmetric. But this is not to say there is no weather in the southern hemisphere!

Figure 8.7: (Top) Baroclinic eddies in the 'eddy' regime viewed from the side. (Bottom) View from above. Eddies draw fluid from the periphery in toward the centre at point A and vice-versa at point B. The eddies are created by the instability of the thermal wind induced by the radial temperature gradient due to the presence of the ice bucket at the centre of the tank. The diameter of the ice bucket is 15 cm.

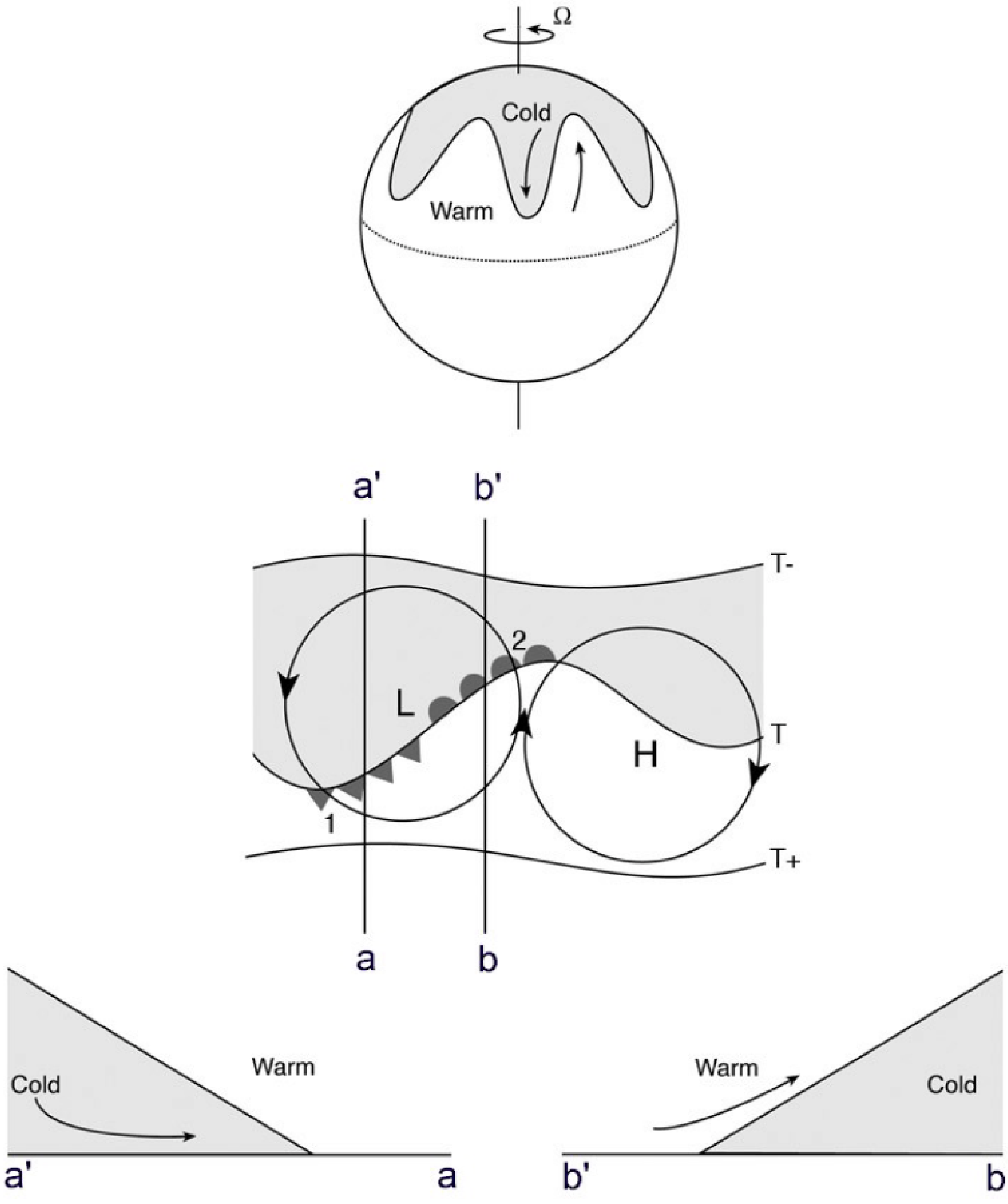


Figure 8.8: (top) In middle latitudes eddies transport warm air poleward and upward and cold air equatorward and downward. Thus, the eddies tend to "stir" the atmosphere laterally reducing the equator-to-pole temperature contrast. (middle) To the west of the 'L' cold air is carried in to the tropics. To the east, warm air is carried toward the pole. The resulting cold fronts (marked by triangles) and warm fronts (marked by semi-circles) are indicated. (bottom) Sections through the cold front, $a \longrightarrow a'$, and the warm front, $b \longrightarrow b'$, respectively.

The eddies arise because of baroclinic instability. Consider the following case with constant Coriolis parameter f,

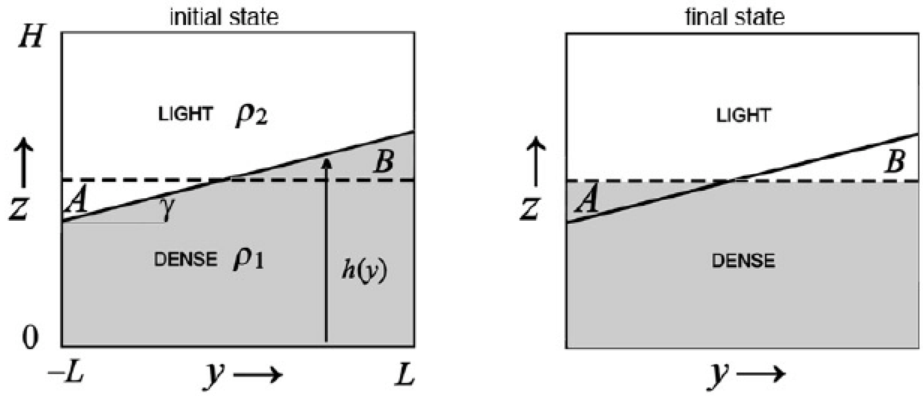


Figure 8.9: Reduction in the available potential energy of a two-layer fluid moving from an initial sate in which the interface is tilted (left) to the final state in which the interface is horizontal (right). Dense fluid is shaded. The net effect of the rearrangement is to exchange heavy fluid downward and light fluid upward such that, in the wedge B, heavy fluid is replaced by light fluid, while the opposite occurs in the wedge A.

In the present case, there is vertical shear and also potential energy that may be released. So it has elements of the Kelvin-Helmhotz instability and the thermal convection but is constrained by thermal wind balance. The amount of energy that can be released by leveling the interface includes both the available potential energy (APE):

$$APE = P - P_{\min} = \frac{1}{3}\rho_1 g' \gamma^2 L^3 \ .$$

where g' is the reduced gravity and the original interface is $H/2+\gamma y$, and the kinetic energy in the mean flow:

$$KE = \int_h^H \int_{-L}^{L} \frac{1}{2}\rho_2 u_2^2 \, dy \, dz = \frac{1}{2}\rho_2 \frac{g'^2\gamma^2}{(2\Omega)^2} HL$$

We have assumed the bottom layer is at rest and calculated the velocity in the upper layer using the thermal wind balance: $u_2=g'\gamma/2\Omega$. The ratio of APE to KE is

$$\frac{APE}{KE} = \frac{2}{3}\left(\frac{L}{L_\rho}\right)^2$$

where L_ρ is the deformation radius sqrt(g'H)/f. In the atmosphere, L is the scale for meridional temperature change and this ratio is ~50. In the ocean, it's even larger. Therefore, there is more energy stored in APE than in KE. In this sense, the instability is different from the Kelvin-Helmholtz instability but more similar to thermal convection.

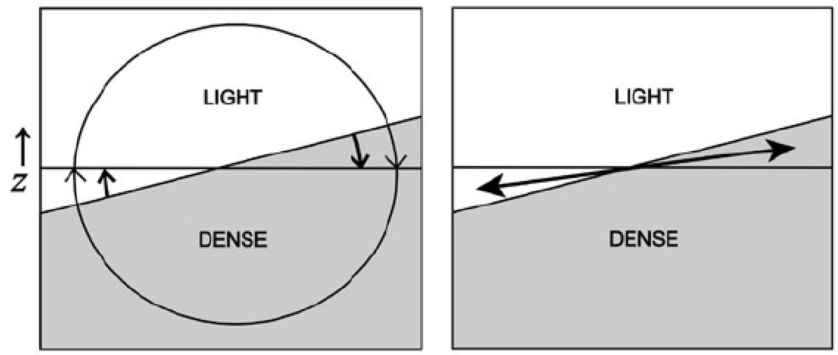


Figure 8.10: Release of available potential energy in a two-component fluid. (left) Non-rotating (or very slowly rotating) case: azimuthally uniform overturning. (right) Rapidly rotating case: sloping exchange in the wedge of instability by baroclinic eddies.

Now how can this APE be released? It cannot be released by direct overturning as in a nonrotating frame. Instead, it occurs through sloping exchange in the wedge of instability: light (warm) fluid moves upwards and radially inward at one azimuth while heavy (cold) fluid moves downward and radially outward at another. A simple model that illustrates this instability is due to Eady (1949), where he showed the unstable waves have typical lateral scale of the deformation radius and have growth rates $\frac{f}{N}\frac{d\bar{u}}{dz} \sim \frac{U}{L_\rho}$.

The same argument can be made for the atmosphere with density replaced by potential temperature:

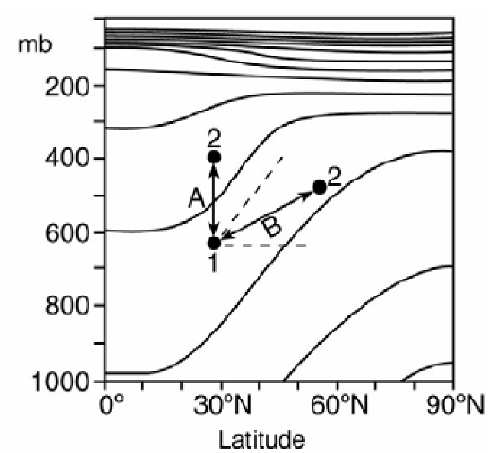


Figure 8.11: Air parcels '1' and '2' are exchanged along paths marked A and B conserving potential temperature θ. The continuous lines are observed θ surfaces — see Fig. 5.8. The tilted dotted line is parallel to the local θ.

As warm fluid moves equatorward and cold fluid moves poleward, these eddies transport heat to high latitudes. This explains why the observed meridional temperature gradient is lower than that from radiative-convective equilibrium. This also explains why there is net radiative heating in the tropics and cooling in the extratropics.

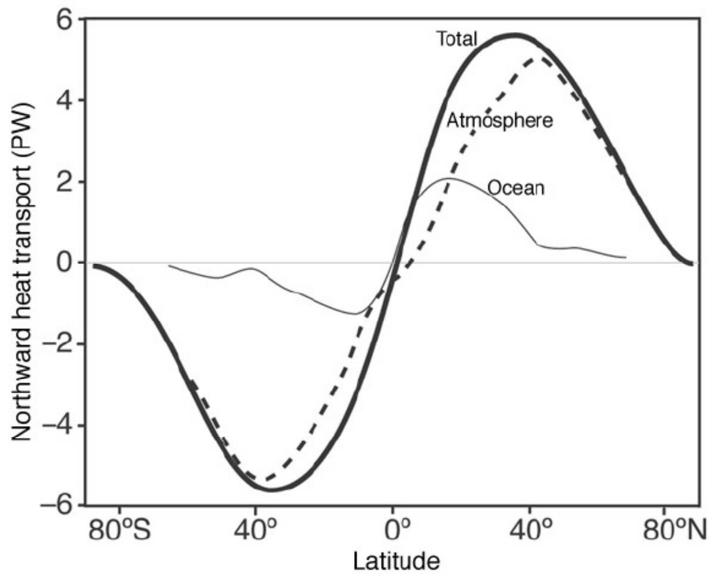


Figure 8.13: The Ocean (thin) and Atmospheric (dotted) contributions to the total northwards heat flux (thick) based on the NCEP reanalysis (in PW$= 10^{15}W$) by (i) estimating the net surface heat flux over the ocean (ii) the associated oceanic contribution, correcting for heat storage associated with global warming and constraining the ocean heat transport to be -0.1 PW at 68°S (iii) deducing the atmospheric contribution as a residual. The total meridional heat flux, as in Fig.5.6, is also plotted (thick). From Trenberth and Caron (2001).

(In comparison, the world power consumption is ~20TW)

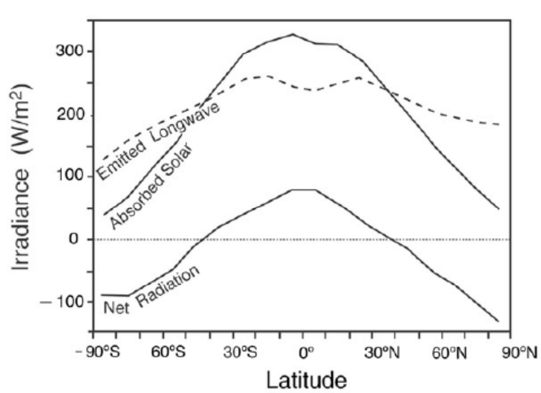


Figure 5.5: Annual mean absorbed solar radiation, emitted longwave radiation and net radiation, the sum of the two. The slight dip in emitted longwave radiation at the equator is due to radiation from the (cold) tops of deep convecting clouds, as can be seen in Fig.4.26.

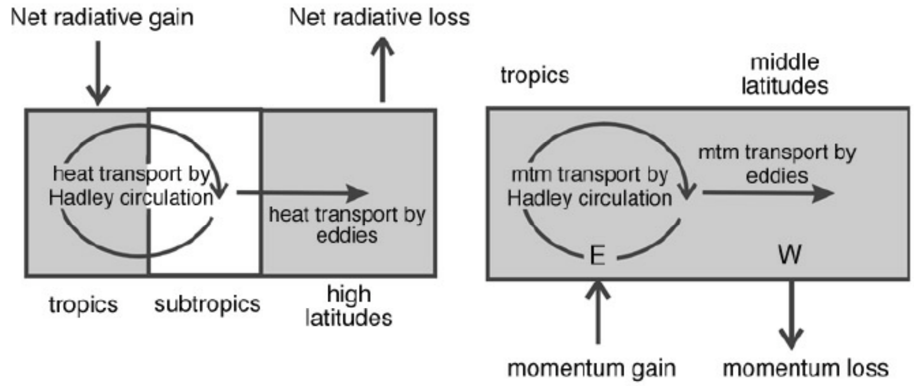


Figure 8.12: Schematic of the transport of (left) energy and (right) momentum by the atmospheric general circulation. Transport occurs through the agency of the Hadley circulation in the tropics, and baroclinic eddies in middle latitudes — see also Fig.8.1.

As eddies propagate meridionally as Rossby waves (to be discussed later), they also transport momentum in such a way that it converges angular momentum into the midlatitude. It happens preferentially in the upper troposphere where the effect of friction on the waves is small. This is ultimately what balances the surface drag on the midlatitude westerly region:

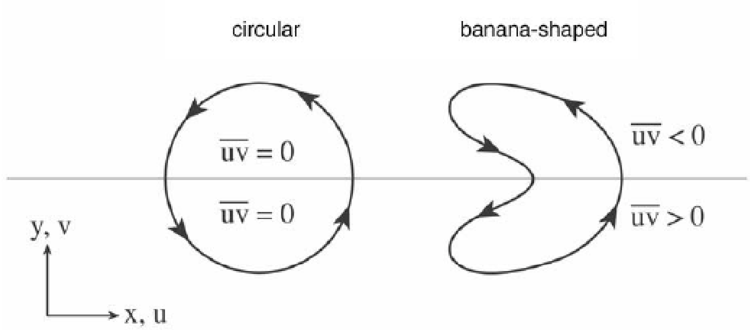


Figure 8.14: Circular eddies on the left are unable to affect a meridional transfer of momentum. The eddies on the right, however, by virtue of their 'banana shape', transfer westerly momentum northwards south of, and southwards north of, the mid-line. Weather systems transport westerly momentum from the tropics ($\overline{uv} > 0$) towards higher latitudes, as required in Fig.8.12(right), by 'trailing' their troughs down in to the tropics (see, e.g., the H_2O distribution shown in Fig.3), as in the southern half of the 'banana-shaped' eddy sketched on the right.

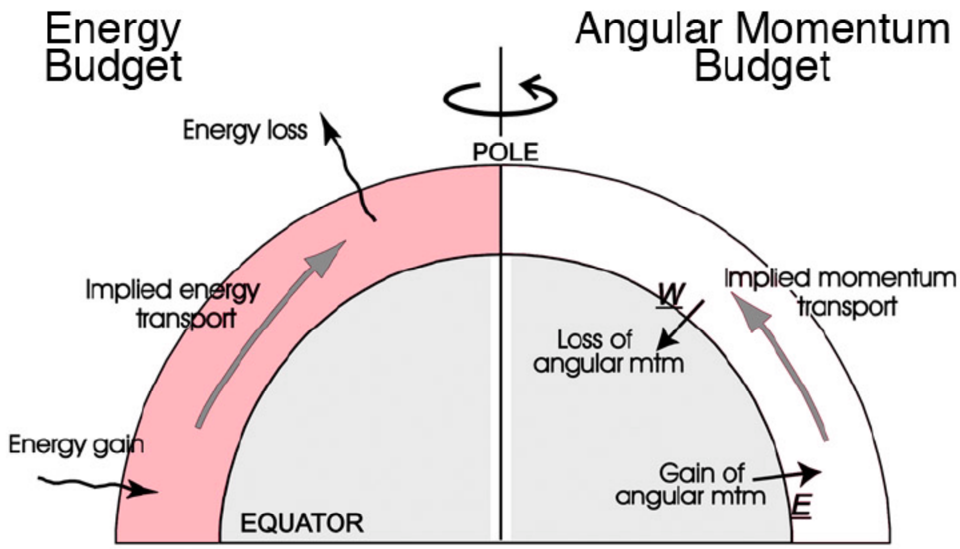
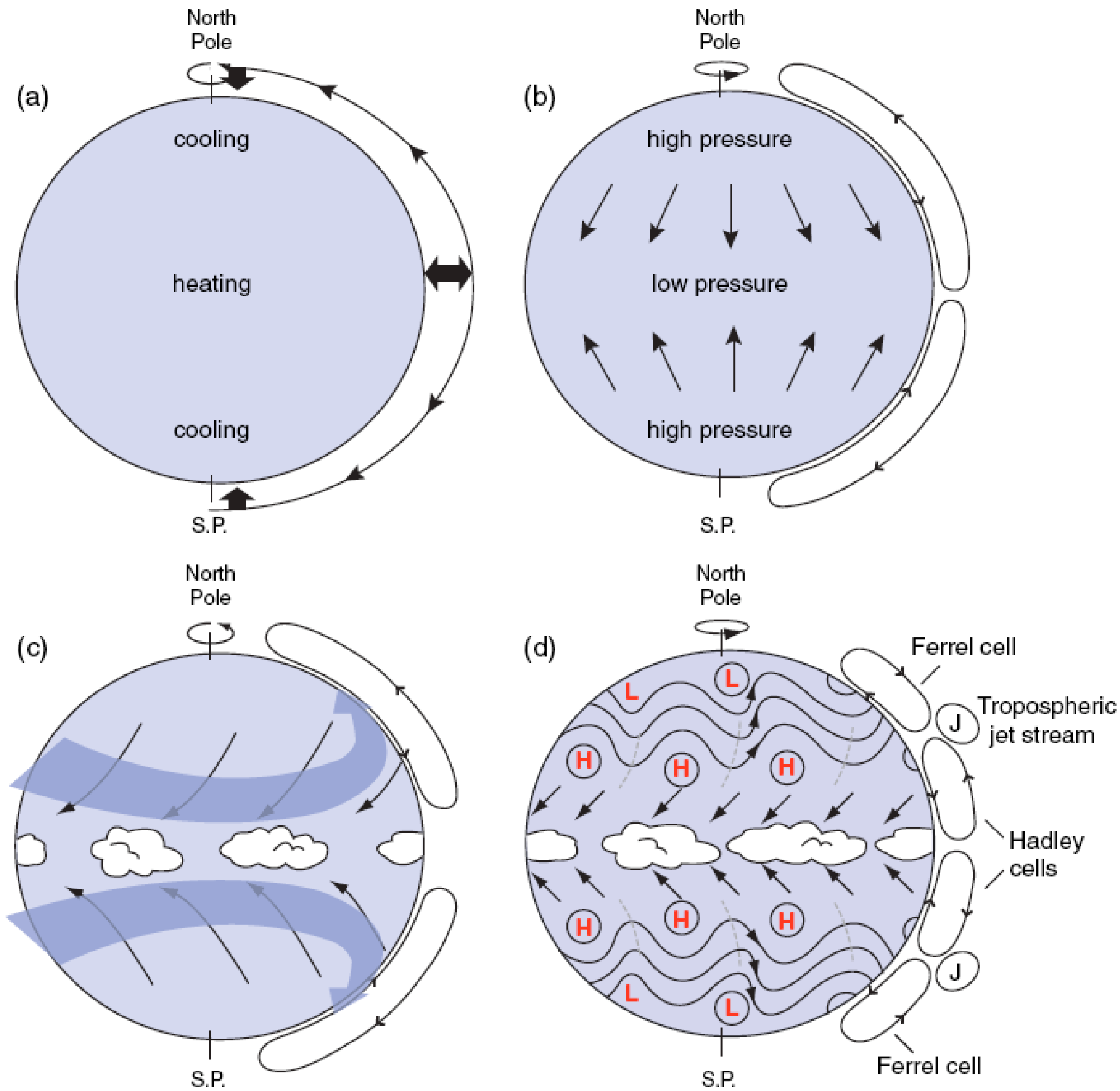


Figure 8.1: Latitudinal transport of (left) heat and (right) angular momentum implied by the observed state of the atmosphere. In the energy budget there is a net radiative gain in the tropics and a net loss at high latitudes; in order to balance the energy budget at each latitude, a poleward heat flux is implied. In the angular momentum budget the atmosphere gains angular momentum in low latitudes (where the surface winds are easterly) and loses it in middle latitudes (where the surface winds are westerly). A poleward atmospheric flux of angular momentum is thus implied.

Now that we have seen that eddies transport heat poleward, thus reducing the meridional temperature gradient, and transport momentum into the upper troposphere at the midlatitudes (where the eddies are generated), we have created a problem. What is it?

The flow can no longer be in thermal wind balance (reduced dT/dy but increased du/dz). This causes a secondary circulation, with air rising at higher latitude and descending in the subtropics. This secondary circulation re-establishes the thermal wind balance by sharpening the temperature gradient and reducing the vertical wind shear. This is the Ferrel Cell, and is driven by the eddy motion.

Putting things together, we have the following picture:


North Pole
(a)
cooling
heating
cooling
S.P.
North Pole
(b)
high pressure
low pressure
high pressure
S.P.
North Pole
(c)
S.P.
North Pole
(d)
Ferrel cell
Tropospheric jet stream
J
L
H
Hadley cells
J
Ferrel cell
S.P.