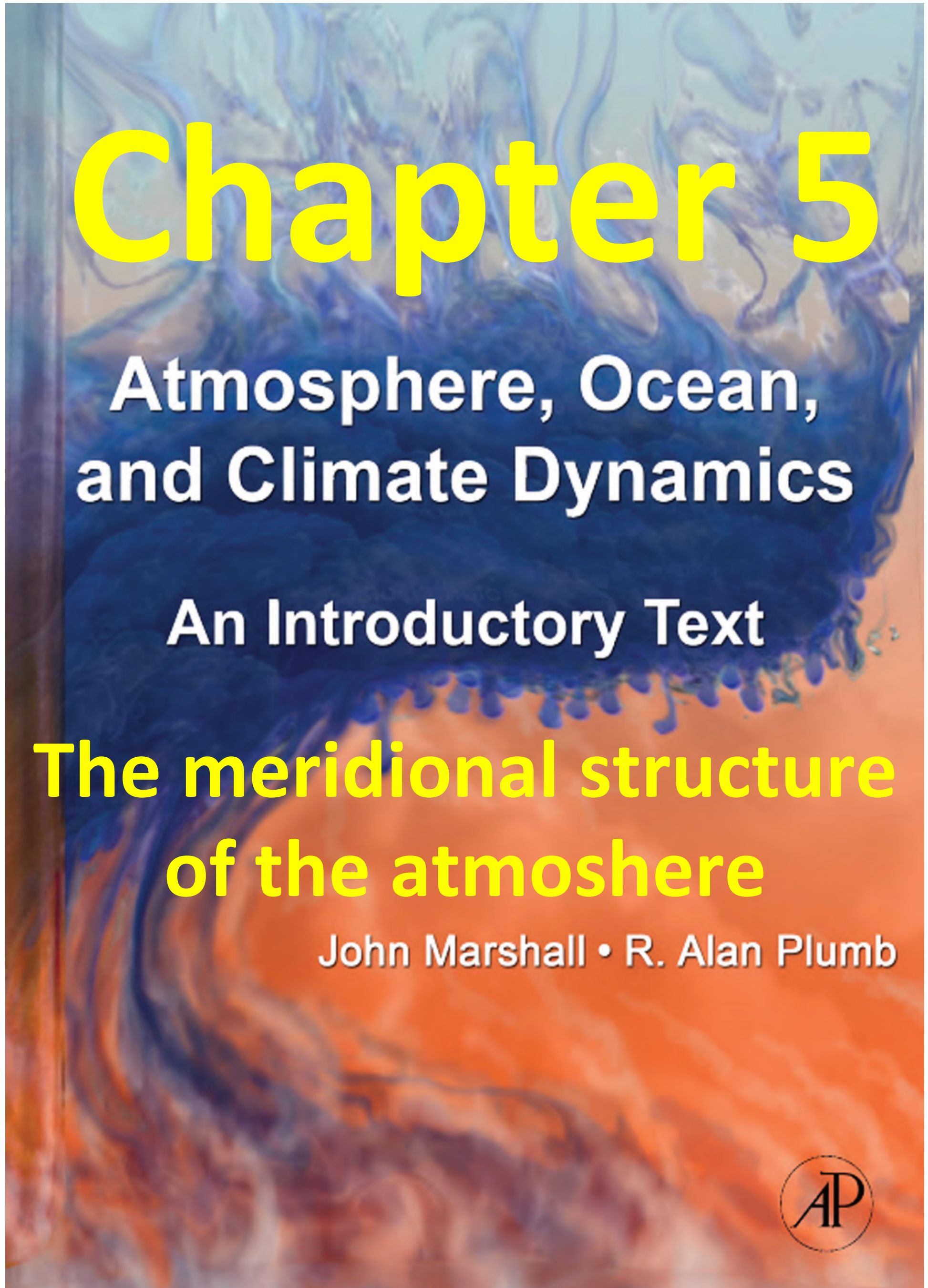

Chapter 5
Atmosphere, Ocean, and Climate Dynamics
An Introductory Text
The meridional structure of the atmoshere
John Marshall • R. Alan Plumb
AP

Chapter 5

Outline

- **Understand the concept of radiation balance between intake and loss of energy by the earth and atmosphere**

- **Observed climatology of atmospheric temperature, pressure, humidity, ad wind.**

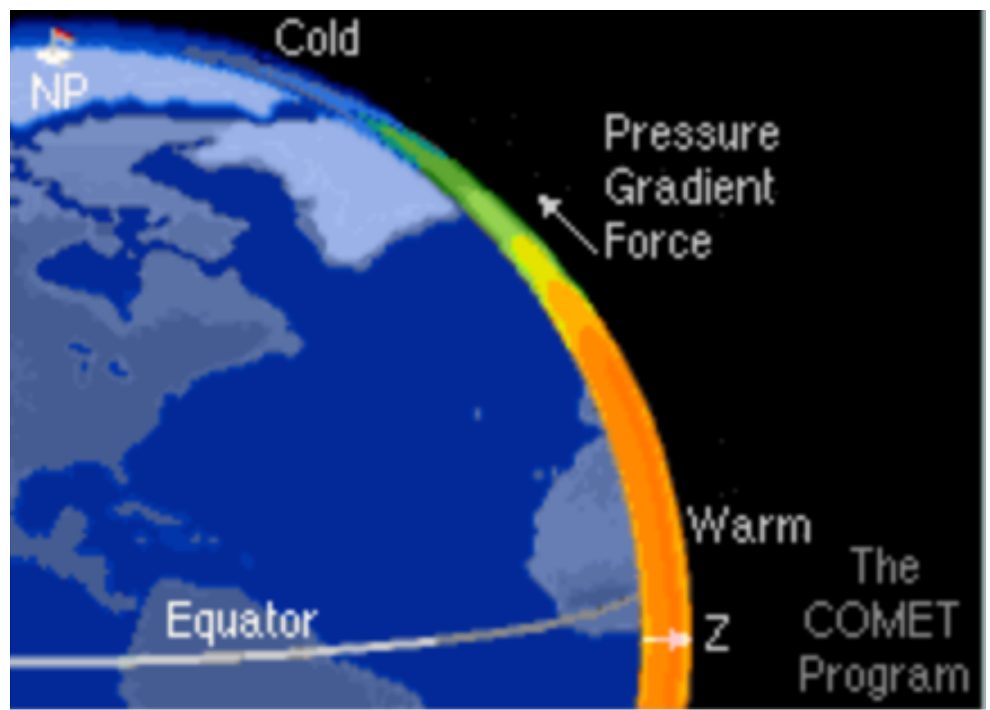
Readings: Chapter 5

Energy is transferred by

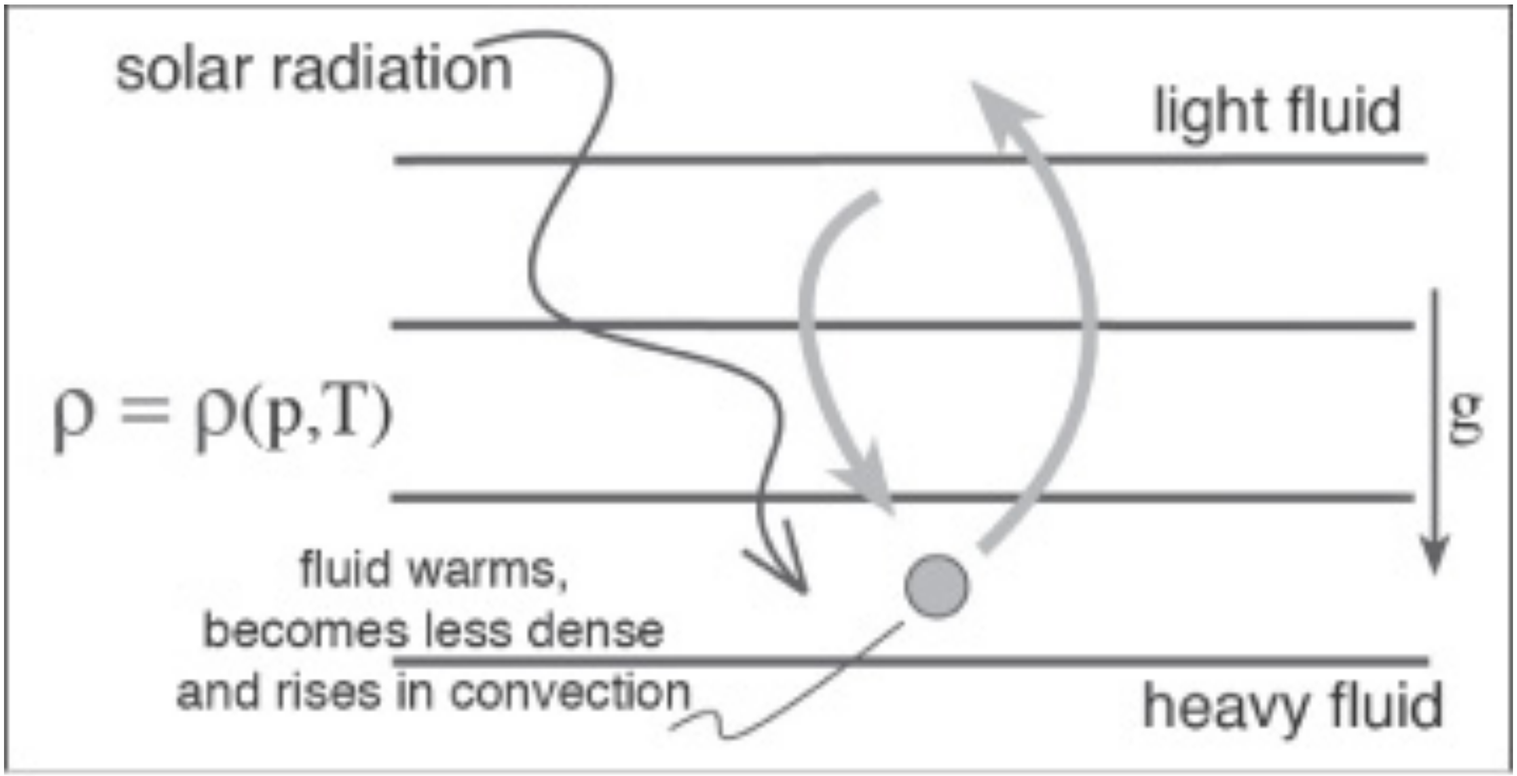
- *radiation* (no mass exchange, no medium required, radiation moves at the speed of light);
- *conduction* (no mass exchanged, heat transferred by vibration and collision among atoms and molecules), and
- *convection* (mass exchanged, fluid parcels with different amounts of energy change places, the net movement of mass is not necessary for energy to be transferred).

Energy density min at NP

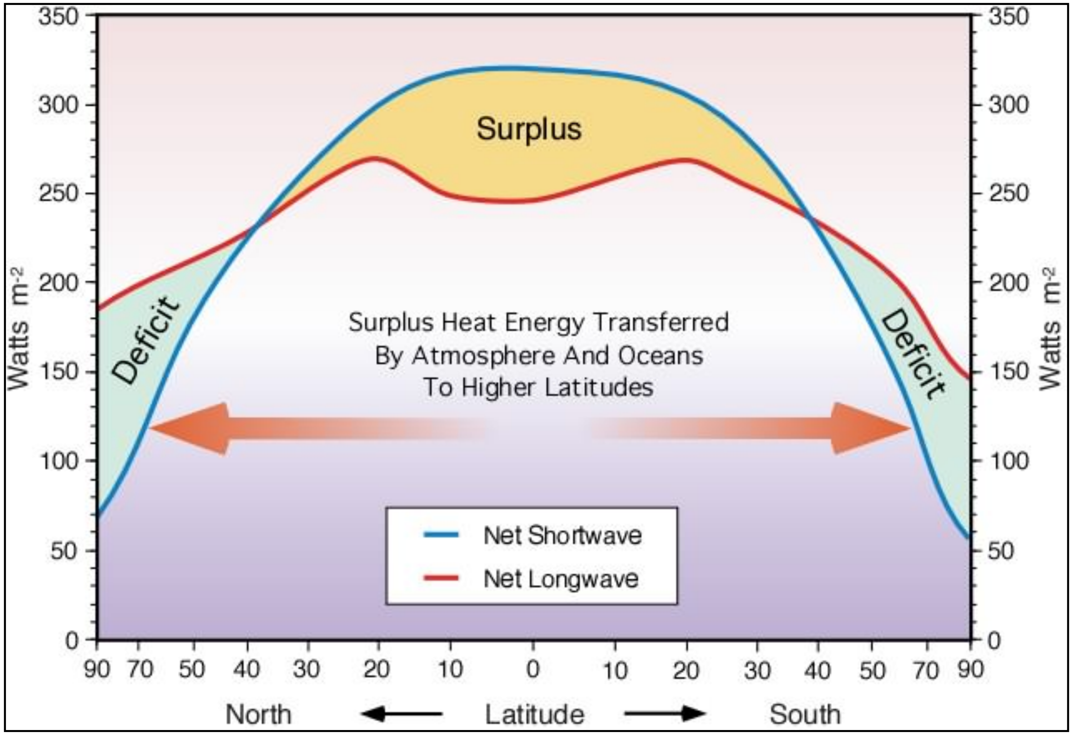
Energy density max at EQ

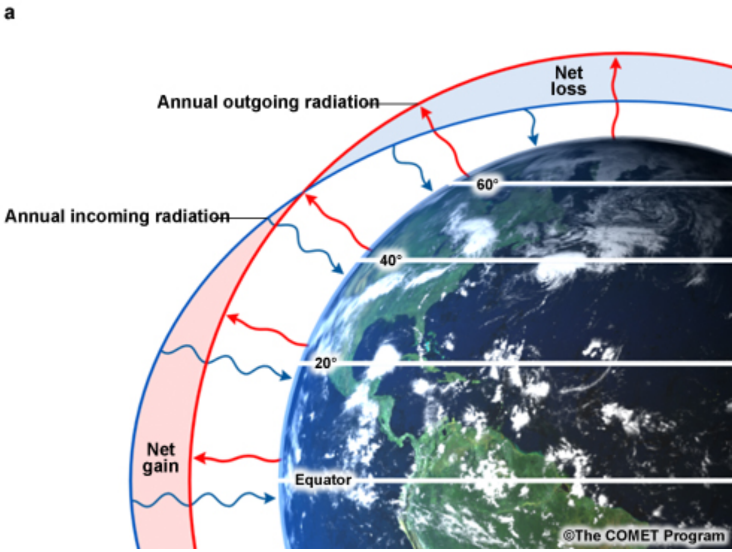


CONVECTION


solar radiation
light fluid
$\rho = \rho(p,T)$
g
fluid warms,
becomes less dense
and rises in convection
heavy fluid

The General Circulation





The role of the general circulation is to redistribute energy from the tropics (surplus) to the poles (deficit)

Meridional transport -> local energy balance -> transport of energy

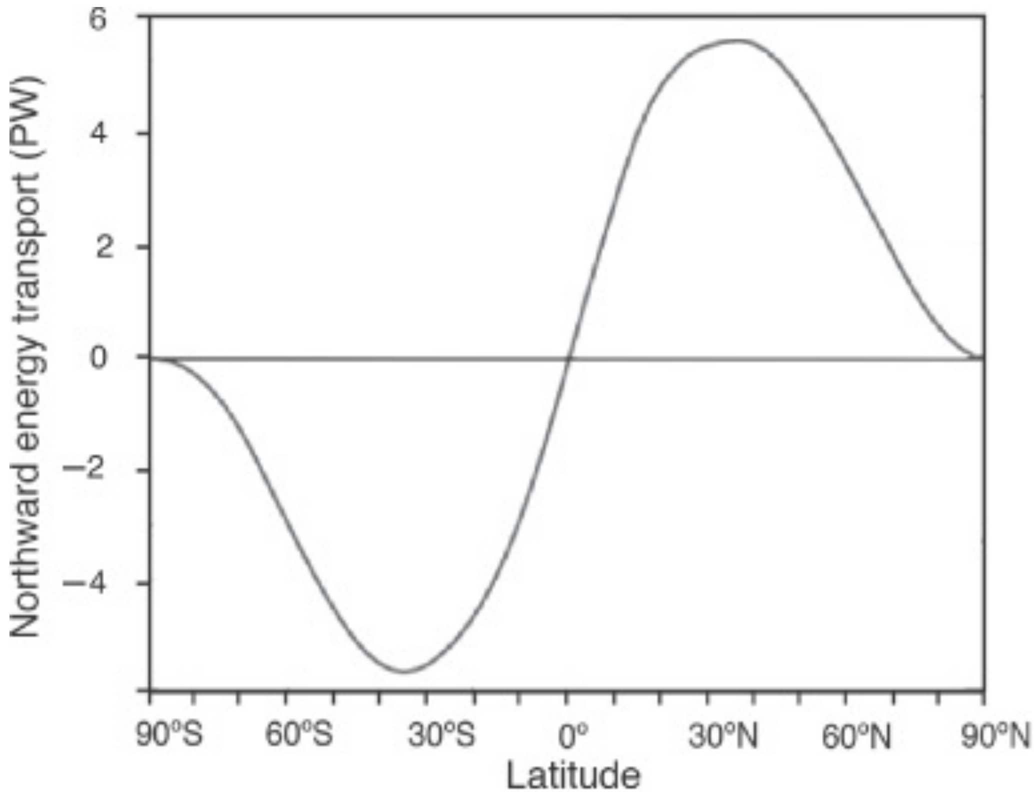
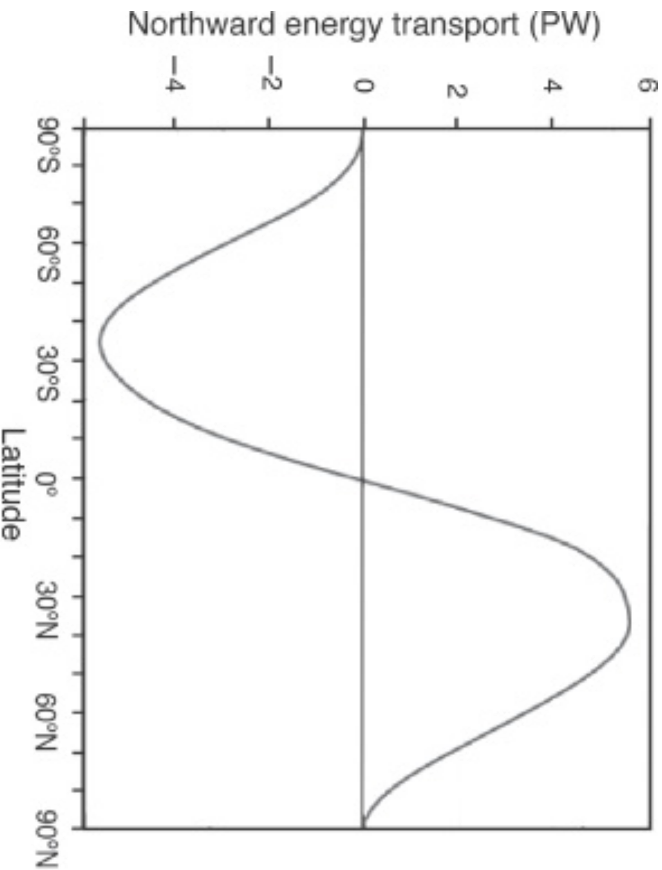


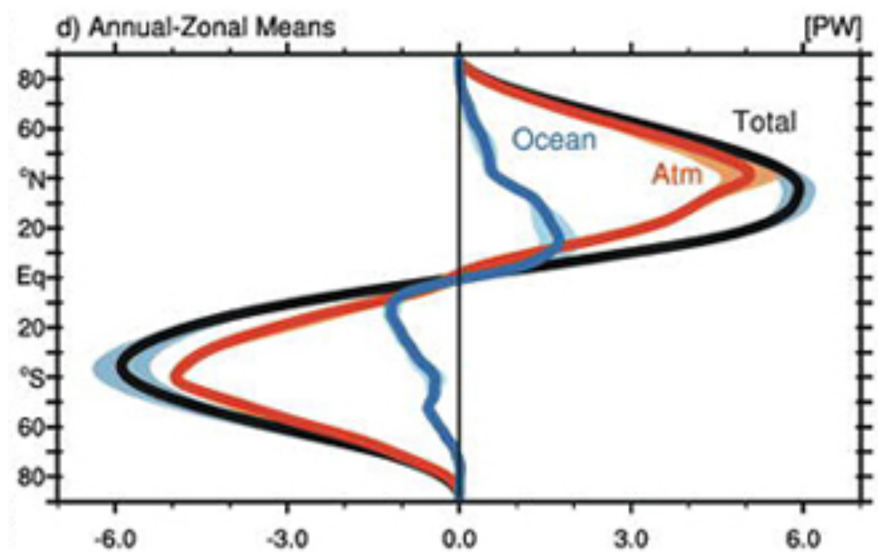
Figure 5.6: The northward energy transport deduced by top of the atmosphere measurements of incoming and outgoing solar and terrestrial radiation from the ERBE satellite. The units are in $PW = 10^{15}$W (see Trenberth and Caron, 2001). This curve is deduced by integrating the "net radiation" plotted in Fig. 5.5 meridionally. See Chapter 11 for a more detailed discussion.

Flux of energy in each hemisphere is 6×10^{15}W

Meridional transport -> local energy balance -> transport of energy

Annual Mean Meridional Energy Transport





Fasullo and Trenberth 2008

Meridional winds - Hadley's suggestion

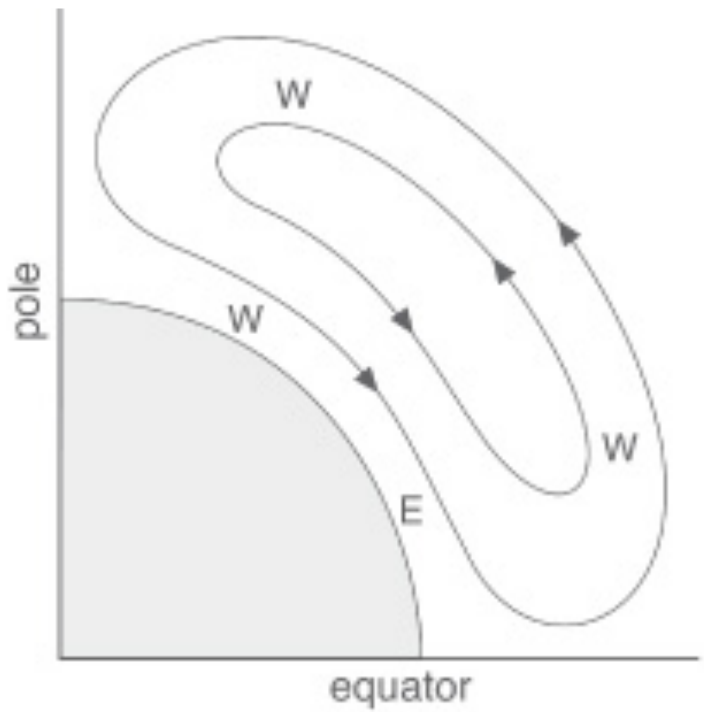
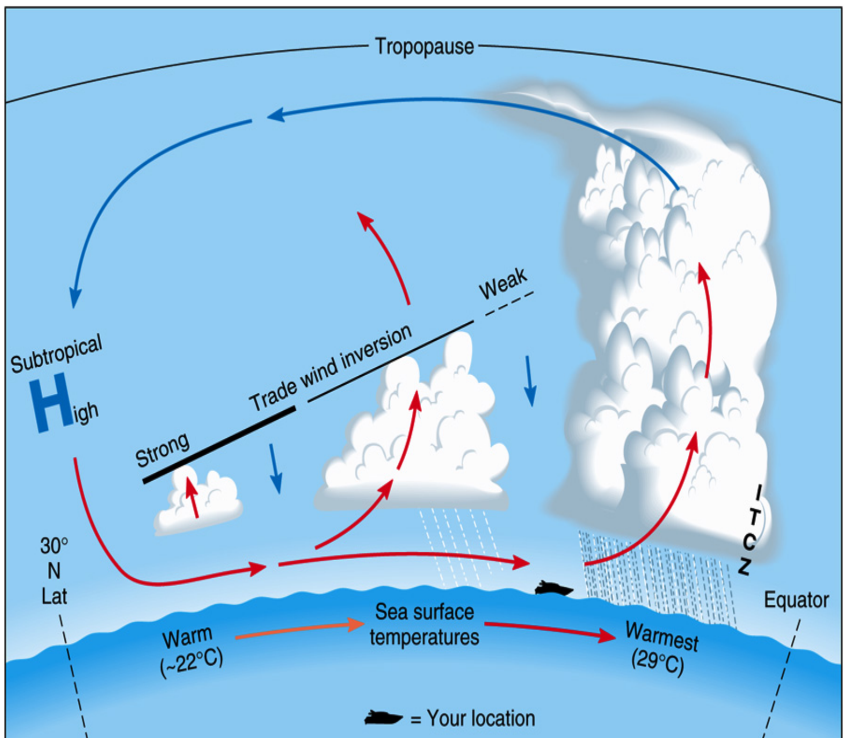
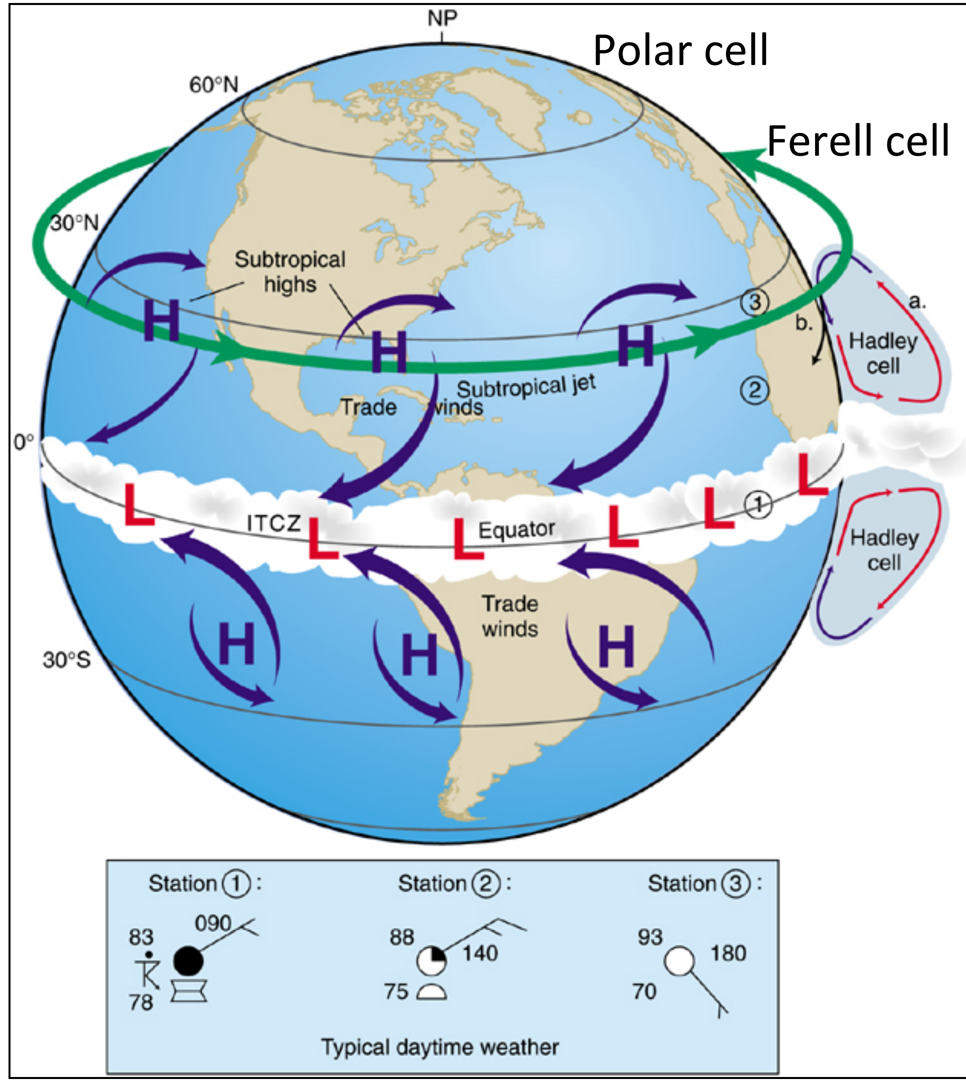


Figure 5.19: The circulation envisaged by Hadley (1735) comprising one giant meridional cell stretching from equator to pole. Regions where Hadley hypothesized westerly (W) and easterly (E) winds are marked.

The General Circulation - Tropics

The Hadley circulation describes a large (almost half the surface of the Earth) thermal circulation



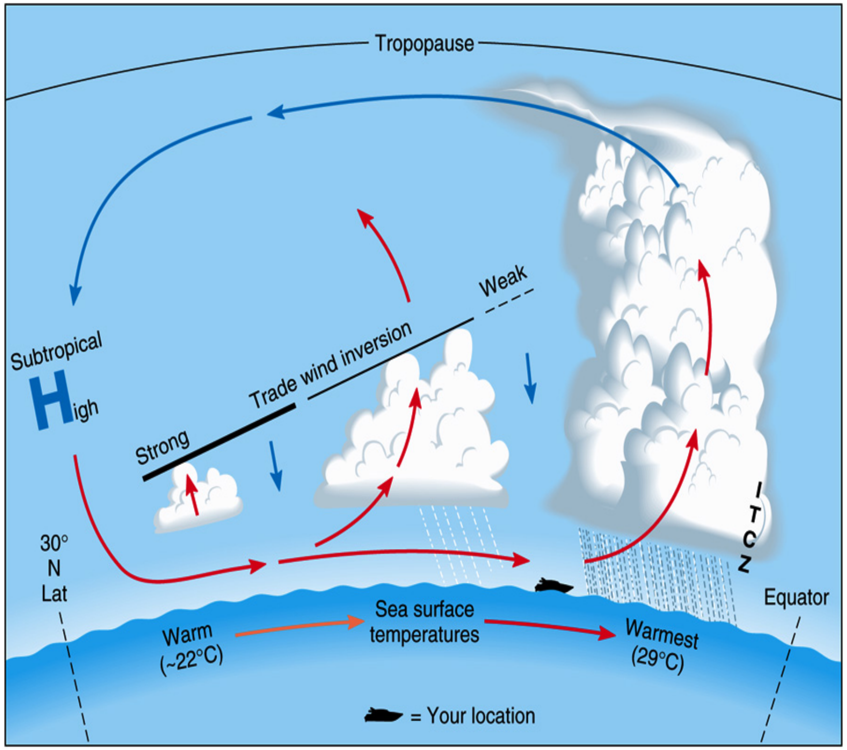


DLA Fig. 10.27

(Fig. 5.18 and 5.19 Marshall and Plumb)

The General Circulation - Tropics

The Hadley circulation describes a large (almost half the surface of the Earth) thermal circulation



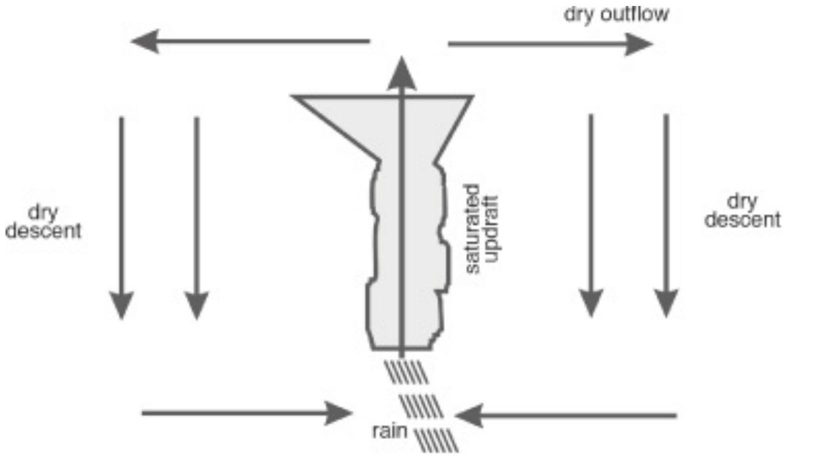
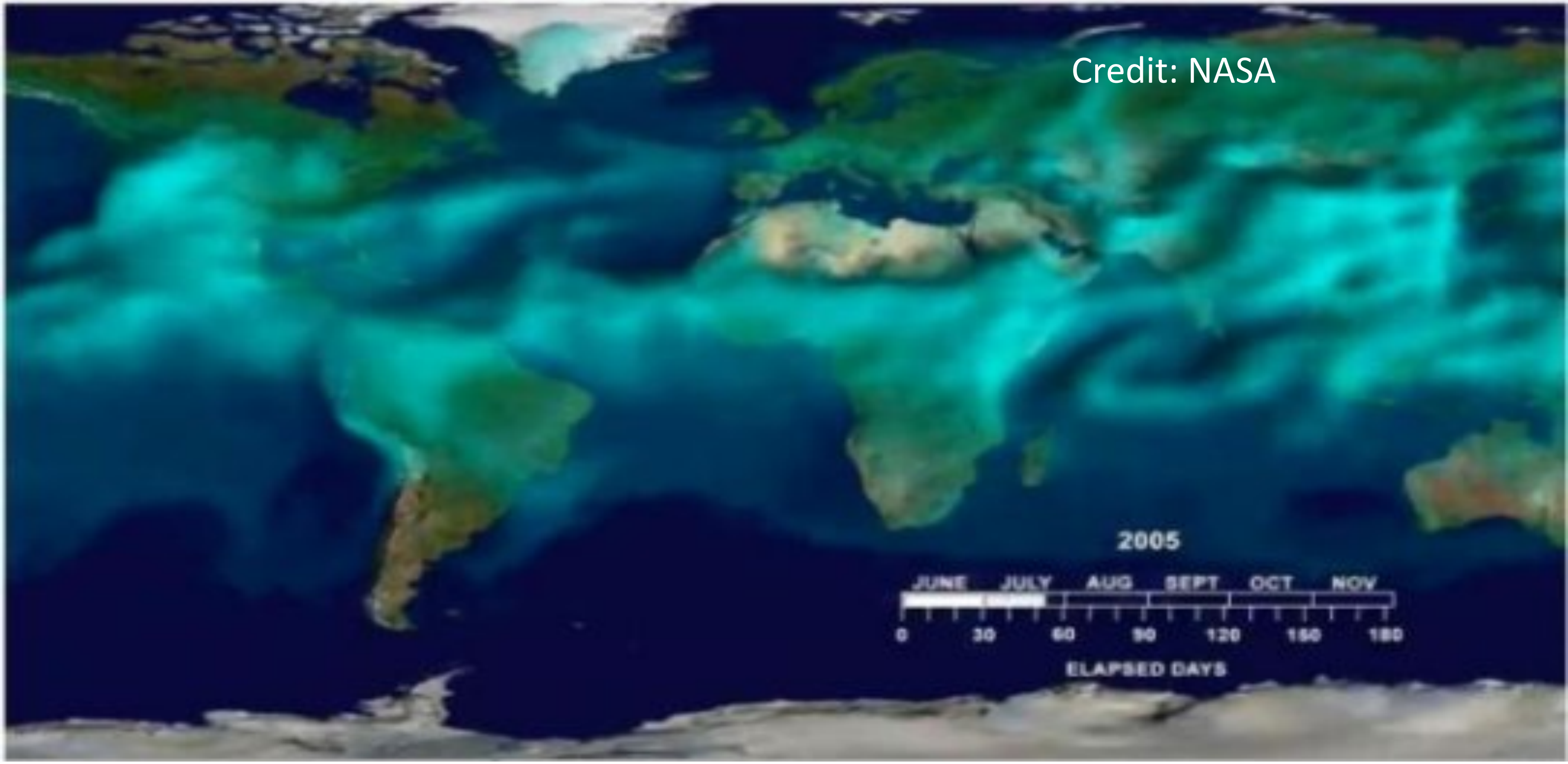


Figure 5.18: Drying due to convection. Within the updraft, air becomes saturated and excess water is rained out. The descending air is very dry. Because the region of ascent is rather narrow and the descent broad, convection acts as a drying agent for the atmosphere as a whole.

- With updraft air became saturated -> producing precipitation
- Top cloud -> lower T than ground -> lost most its water

http://www.nasa.gov/topics/earth/features/vapor_warming.html



The distribution of atmospheric water vapor, a significant greenhouse gas, varies across the globe. During the summer and fall of 2005, this visualization shows that most vapor collects at tropical latitudes, particularly over south Asia.

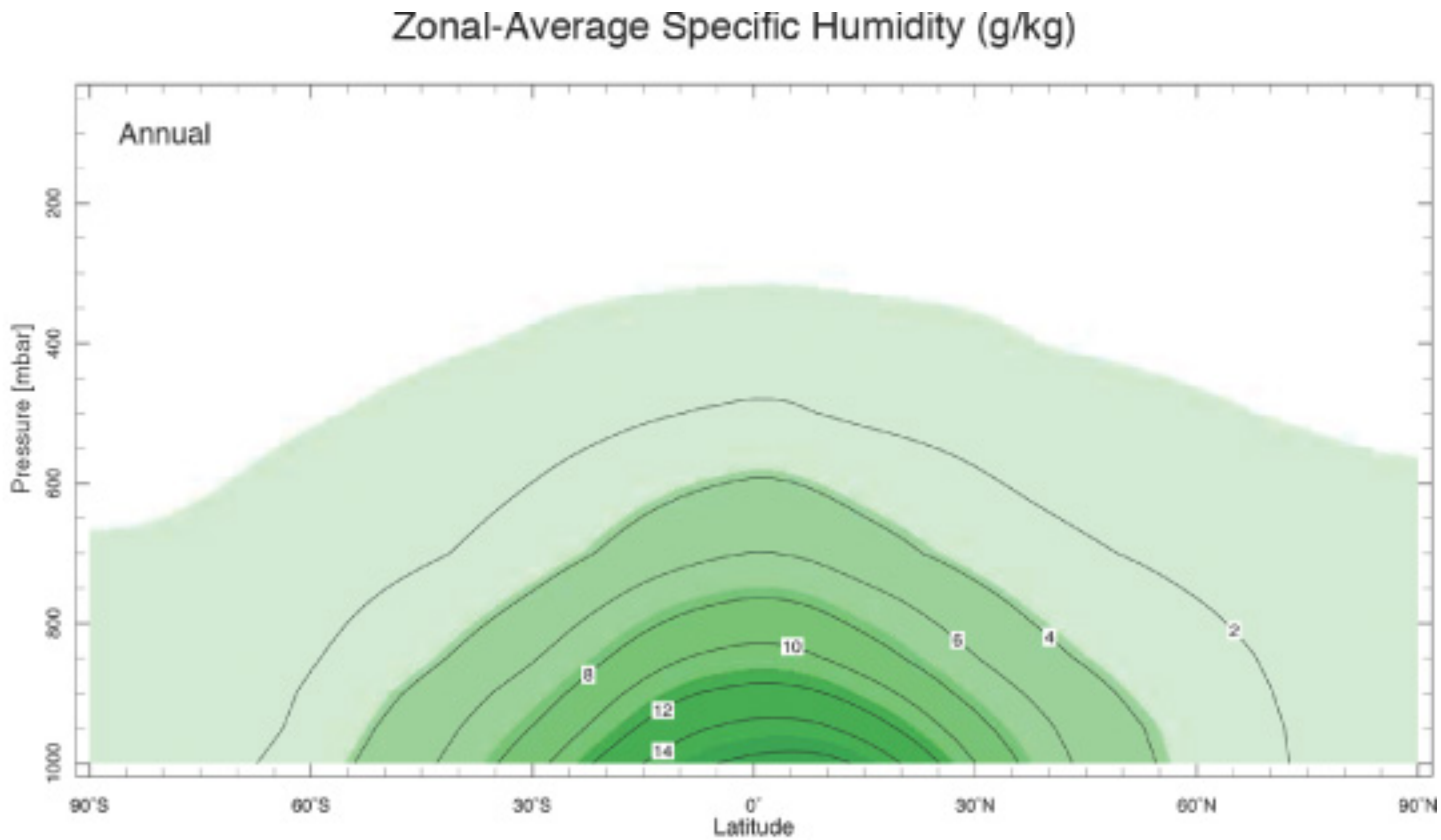


Figure 5.15: Zonally averaged specific humidity *q*, Eq. 4-23, in g kg^{-1} under annual mean conditions. Note that almost all the water vapor in the atmosphere is found where $T > 0°C$ (see Fig. 5.7).

specific humidity mass of water vapor to the mass of air per unit volume defined thus:

ZONAL WINDS

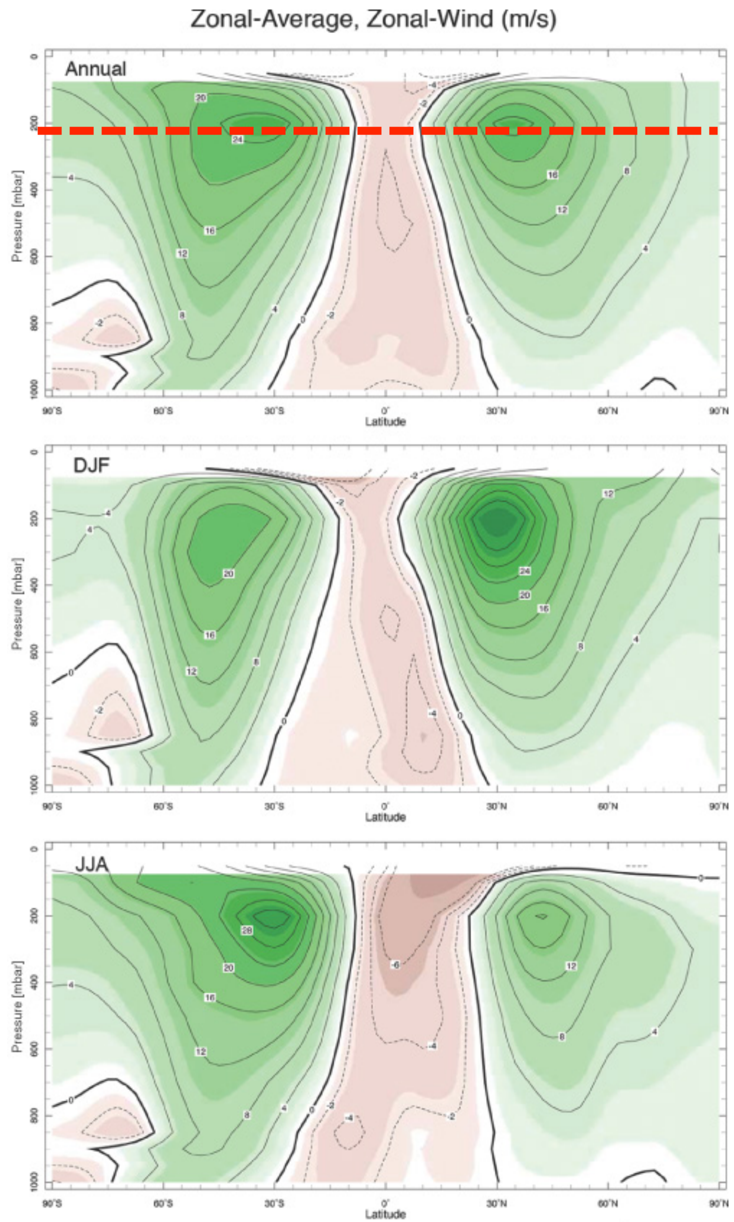
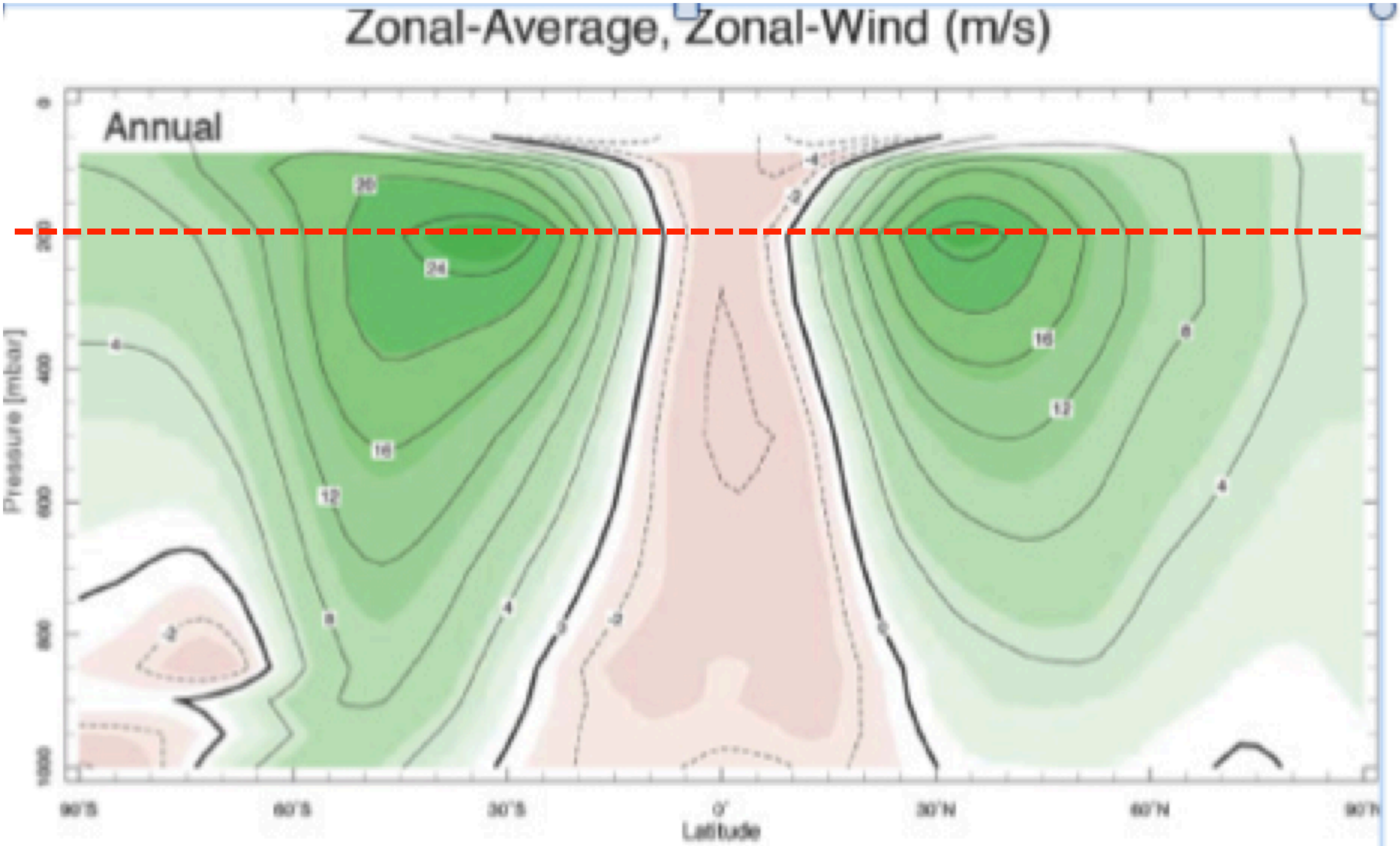


Figure 5.20: Meridional cross-section of zonal-average zonal wind (ms^{-1}) under annual mean conditions (top), DJF (December, January, February) (middle) and JJA (June, July, August) (bottom) conditions.

ZONAL WINDS



Zonal-Average, Zonal-Wind (m/s)
Annual
Pressure [mbar]
Latitude
90°S
60°S
30°S
0°
30°N
60°N
90°N
20
24
16
12
8
4

MERIDIONAL WINDS

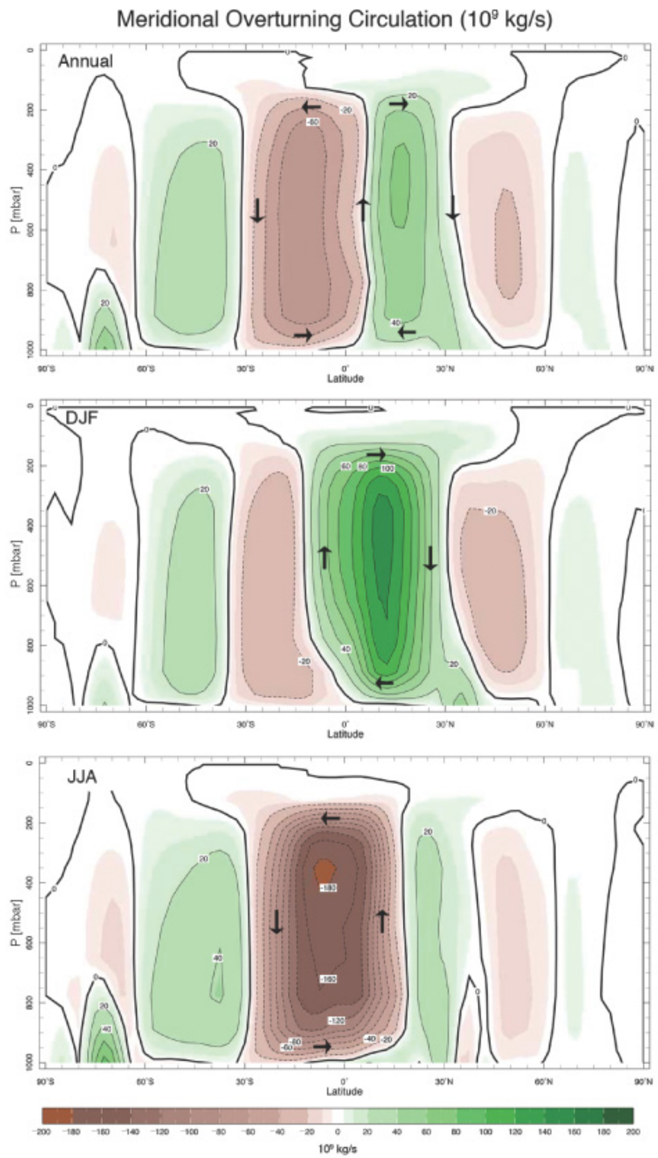


Figure 5.21: The meridional overturning streamfunction ? of the atmosphere in annual mean, DJF, and JJA conditions. [The meridional velocities are related to χ by $v = -(\rho a \cos\varphi)^{-1}\,\partial\chi/\partial z$; $w = (\rho a^2 \cos\varphi)^{-1}\,\partial\chi/\partial\varphi$ Units are in 10^9 kg s^{-1}, or Sverdrups, as discussed in Section 11.5.2. Flow circulates around positive (negative) centers in a clockwise (anticlockwise) sense. Thus in the annual mean, air rises just north of the equator and sinks around $\pm 30°$.

DJF air raises just south of the equator and sinks in the subtropics (30N)

JJA air raises just North of the equator and sinks in the subtropics (30S)

TEMPERATURE IN THE TROPOSPHERE

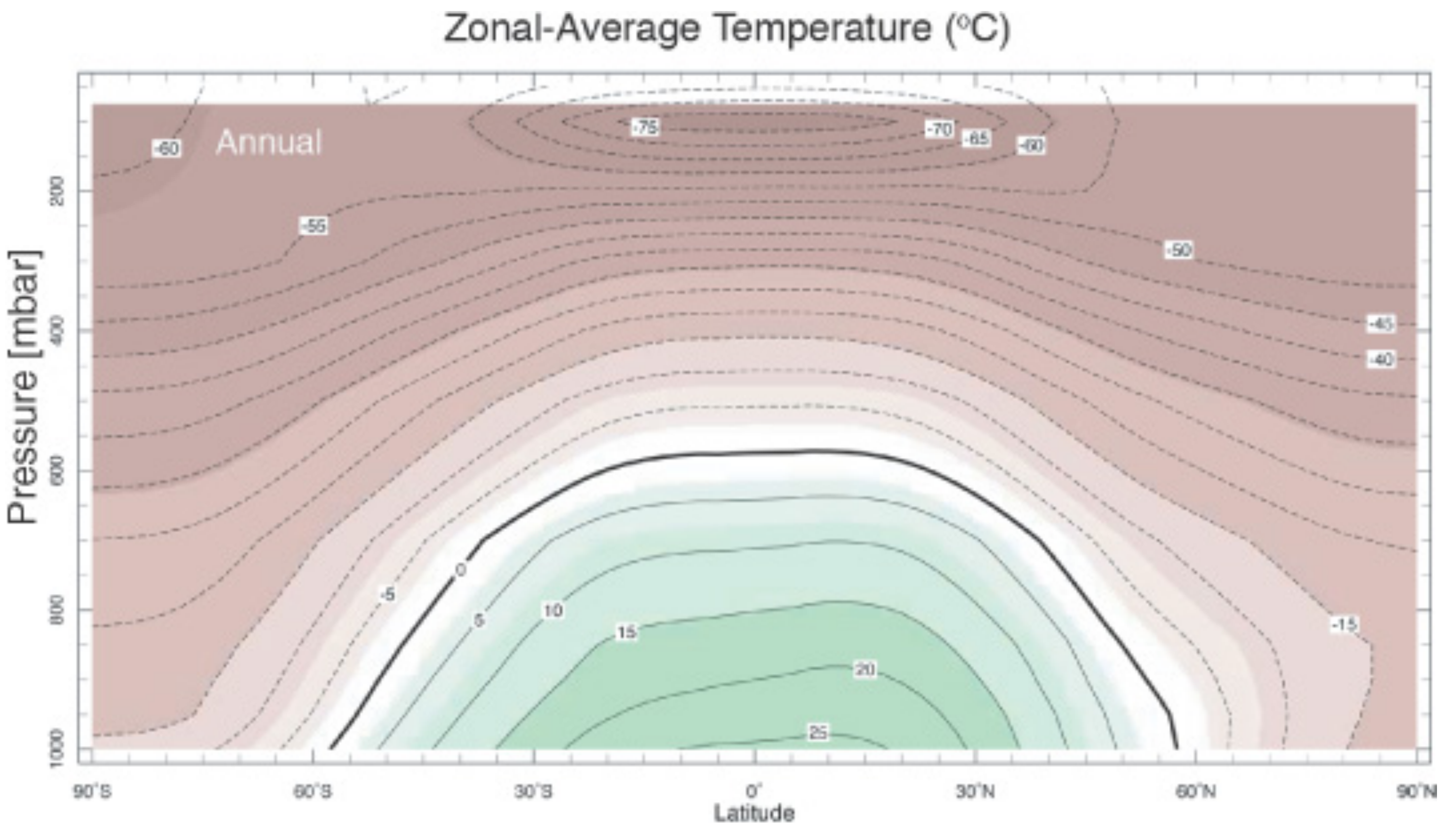


Figure 5.7: The zonally averaged annual-mean temperature in °C.

Annual mean pole-equator temperature difference of 40 degrees Celsius

POTENTIAL TEMPERATURE
DRY OR SUPERSATURATED CONDITIONS

Perfect gas law

$$p = \rho RT$$

First law of thermodynamics

$$\delta Q = c_v dT + p dV$$

Under adiabatic conditions:

$$\kappa = R/c_p = 2/7$$

Cp:specific heat capacity

$$\frac{d\theta}{\theta} = \frac{dT}{T} - \kappa \frac{dp}{p} = 0$$

$$\theta = T \left(\frac{p_0}{p} \right)^{\kappa}$$

The **potential temperature** of a parcel of fluid at pressure is the temperature that the parcel would acquire if adiabatically brought to a standard reference pressure, usually 1000 millibars.

Stability

$$\left.\begin{array}{c}\text{unstable}\\ \text{neutral}\\ \text{stable}\end{array}\right\}\text{ if }\left(\frac{d\theta}{dz}\right)_E\left\{\begin{array}{l}<0\\ =0\\ >0\end{array}\right. . \qquad (4\text{-}18)$$

Attitude to convective systems: Potential temperature is a useful measure of the stability of the unsaturated atmosphere

If the potential temperature decreases with height, the atmosphere is unstable to vertical motions.

Zonal-Average Potential Temperature (K)

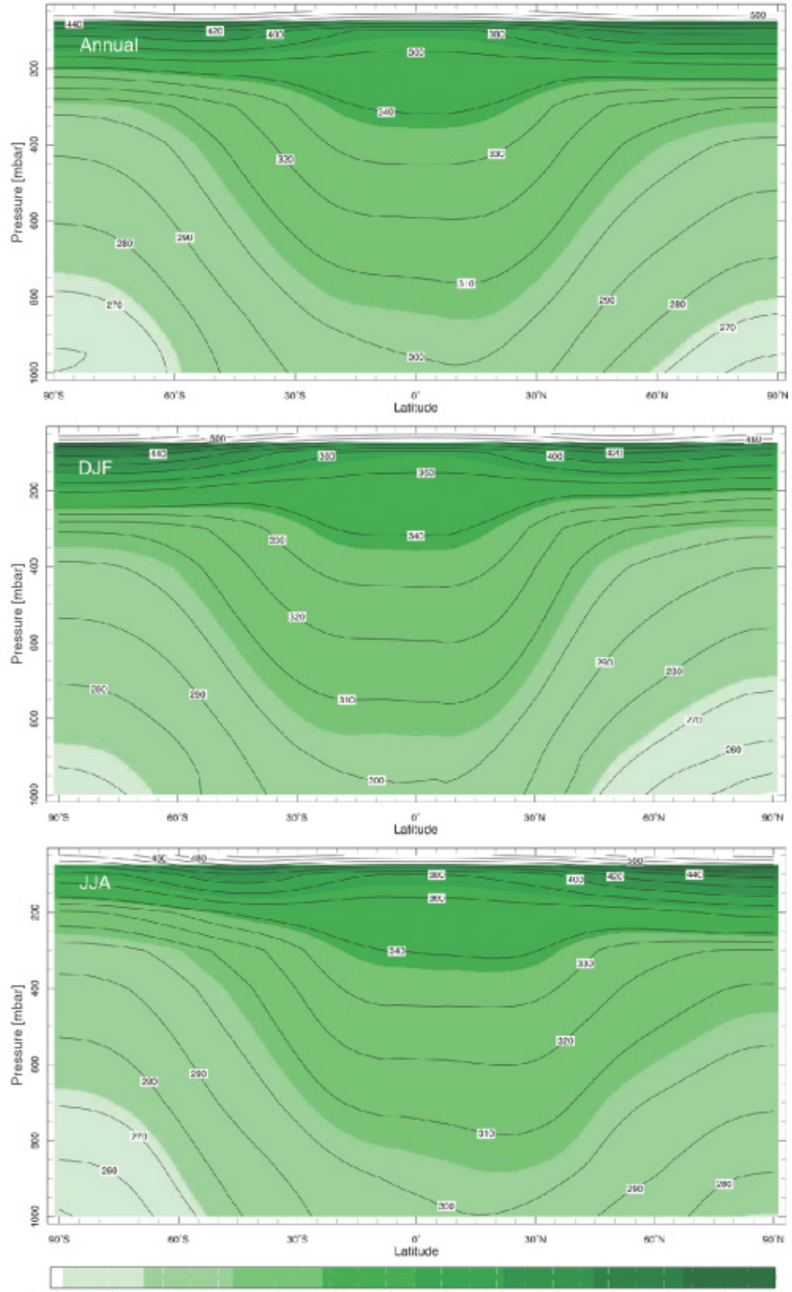


Figure 5.8: The zonally averaged potential temperature in (top) the annual mean, averaged over (middle) December, January, and February (DJF), and (bottom) June, July, and August (JJA).

EQUIVALENT POTENTIAL TEMPERATURE
SATURATED CONDITIONS

Under adiabatic conditions:

$$\theta_e = \theta \exp\left(\frac{Lq}{c_p T}\right), \qquad (4\text{-}30)$$

Equivalent potential temperature θ_e is conserved.

Temperature of a parcel of air that would reach if all the water vapor in the parcel were to condense, releasing its latent heat, and the parcel was brought adiabatically to stantdard reference pressure 1000mbar

$$\theta_e = \theta \exp\left(\frac{Lq}{c_p T}\right), \qquad (4\text{-}30)$$

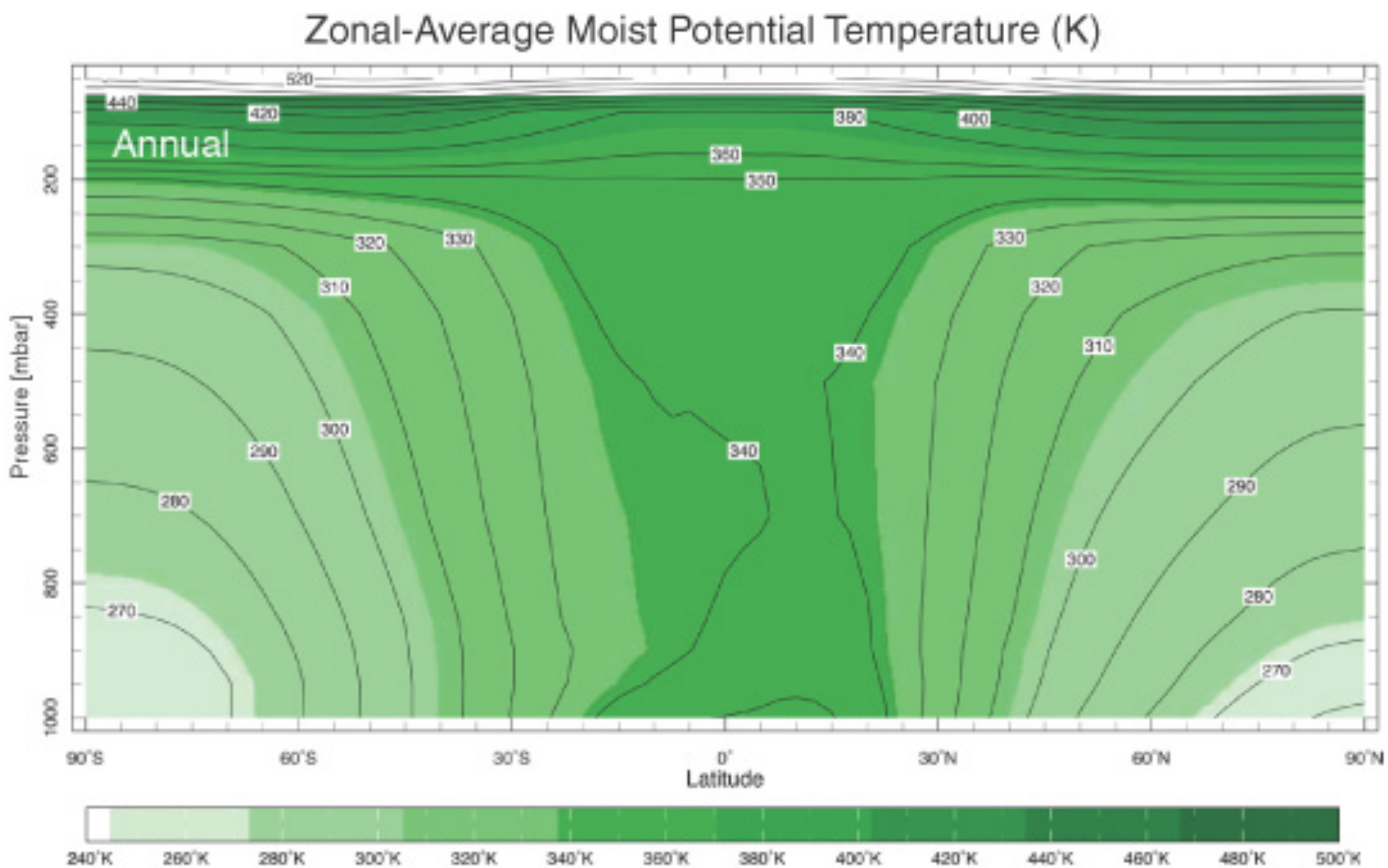


Figure 5.9: The zonal average, annual mean equivalent potential temperature, θ_e, Eq. 4-30.

TEMPERATURE IN THE TROPOSPHERE AND STRATOSPHERE

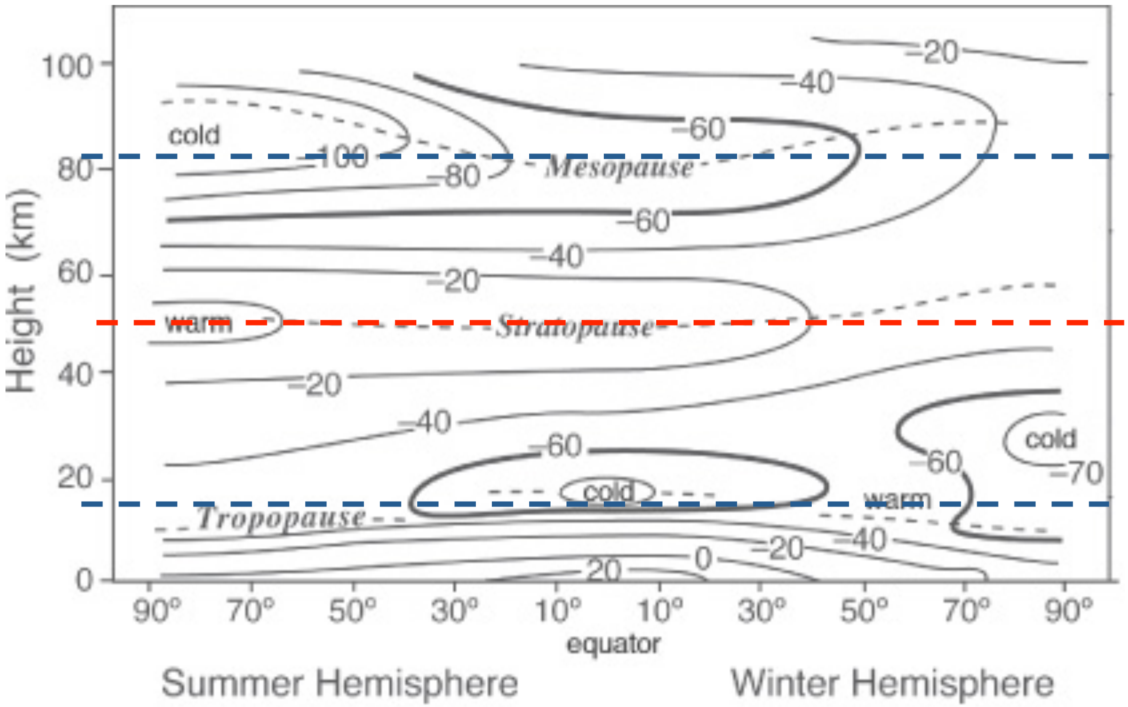


Figure 5.10: The observed, longitudinally averaged temperature distribution (*T*) at northern summer solstice from the surface to a height of 100 km (after Houghton, 1986). Altitudes at which the vertical *T* gradient vanishes are marked by the dotted lines and correspond to the demarcations shown on the *T*(*z*) profile in Fig. 3.1. The −60°C isopleth is thick. Note the vertical scale is in km compared to Fig. 5.7, which is in pressure. To convert between them, use Eq. 3-8.

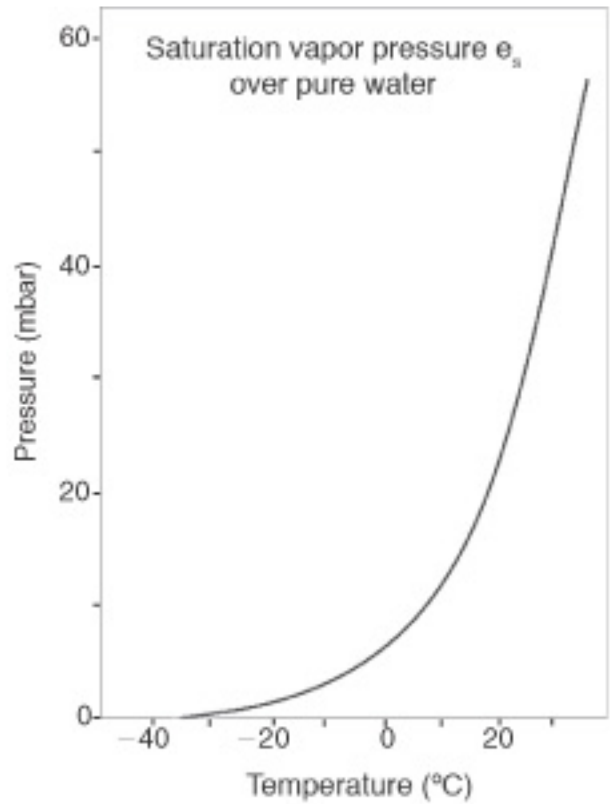


Figure 1.5: Saturation vapor pressure e_s (in mbar) as a function of T in °C (solid curve).

(From Wallace & Hobbs, (2006).)

PRESSURE COORDINATE

$$\frac{\partial p}{\partial z} = -\frac{gp}{RT} \,. \qquad (3\text{-}5)$$

$$\frac{\partial z}{\partial p} = -\frac{RT}{gp}, \qquad (5\text{-}1)$$

or, noting that $p\frac{\partial}{\partial p} = \frac{\partial}{\partial \ln p}$,

$$\frac{\partial z}{\partial \ln p} = -\frac{RT}{g} = -H,$$

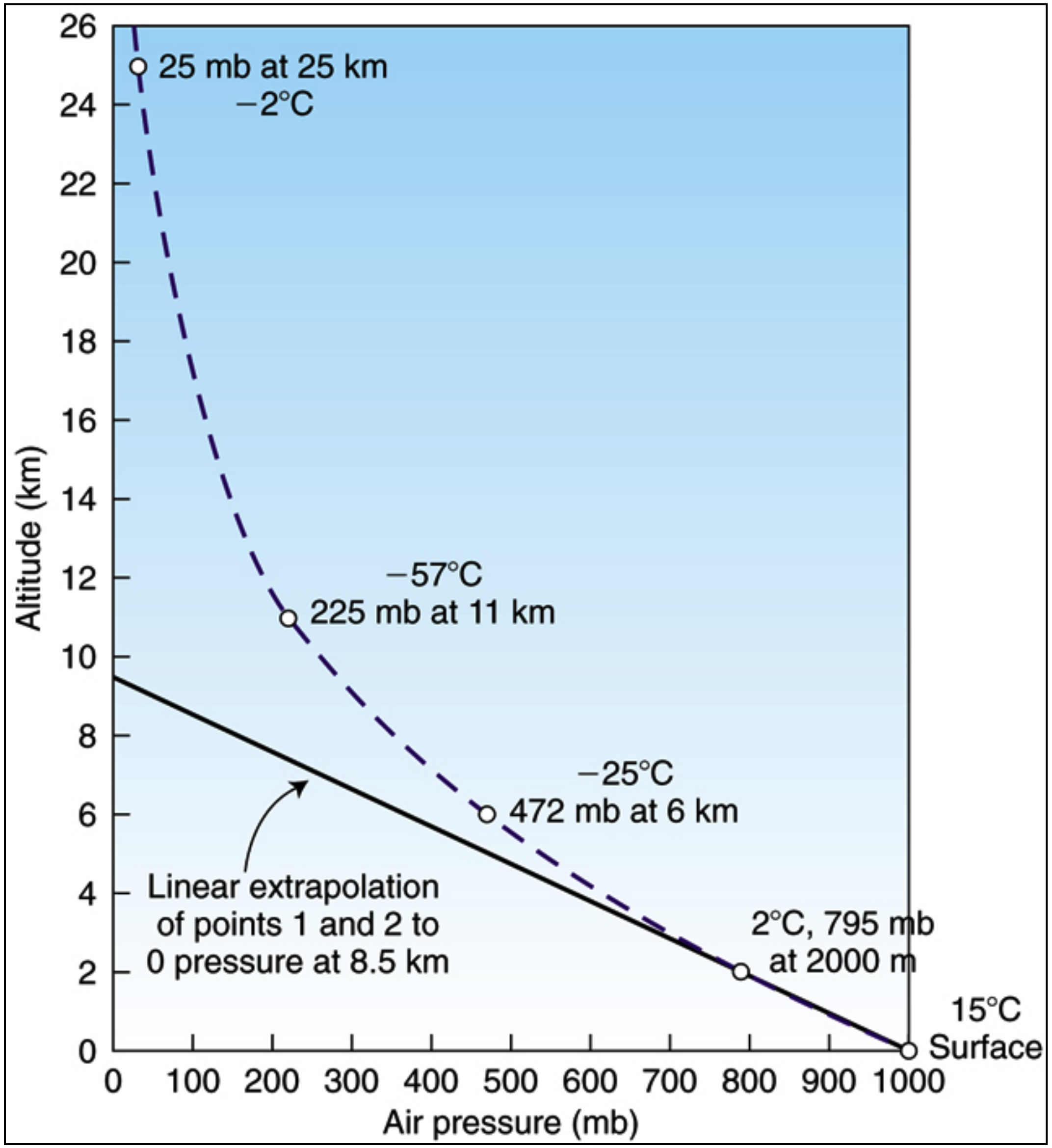
Isothermal atmosphere:

T&*H* const with *z* and *p*

z varies as ln*p*

p varies exponentially with *z*

Vertical Structure – Pressure / Height



DLA Fig. 2.16

nonlinear relationship between pressure and geometric height

GEOPOTENTIAL HEIGHT

$$z(p) = R\int_{p}^{p_s} \frac{T}{g}\frac{dp}{p}, \qquad (5\text{-}2)$$

where we have set $z(p_s) = 0$

Geopotential height high in warm conditions

PRESSURE SURFACE

Monthly mean

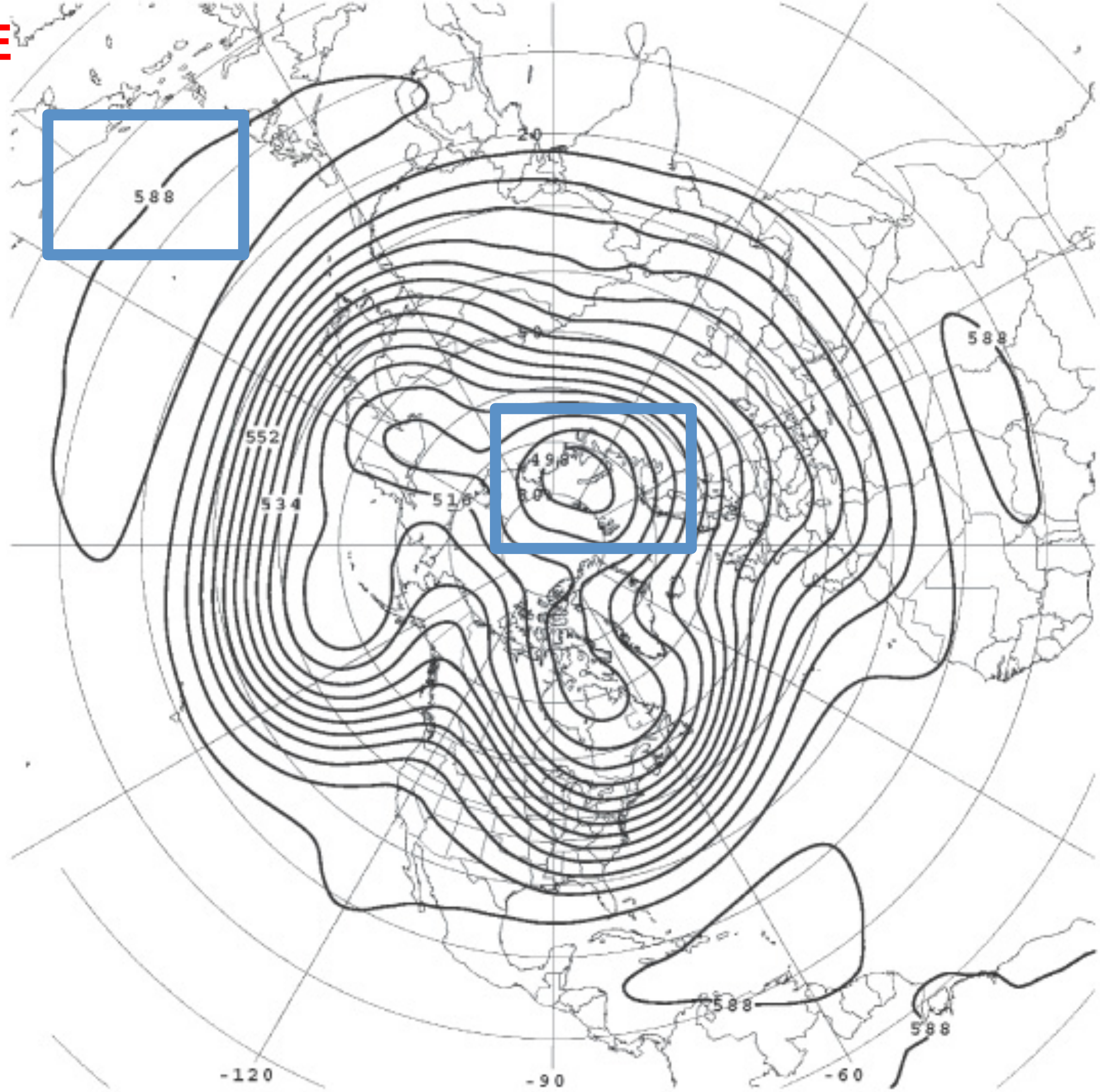


Figure 5.12: The mean height of the 500 mbar surface in January, 2003 (monthly mean). The contour interval is 6 decameters ≡ 60 m. The surface is 5.88 km high in the tropics and 4.98 km high over the pole. Latitude circles are marked every 10°, longitude every 30°.

THICKNESS OF PRESSURE LAYERS

$$z_2 - z_1 = R \int_{p_2}^{p_1} \frac{T}{g} \frac{dp}{p}, \qquad (5\text{-}4)$$

TILT OF PRESSURE SURFACE

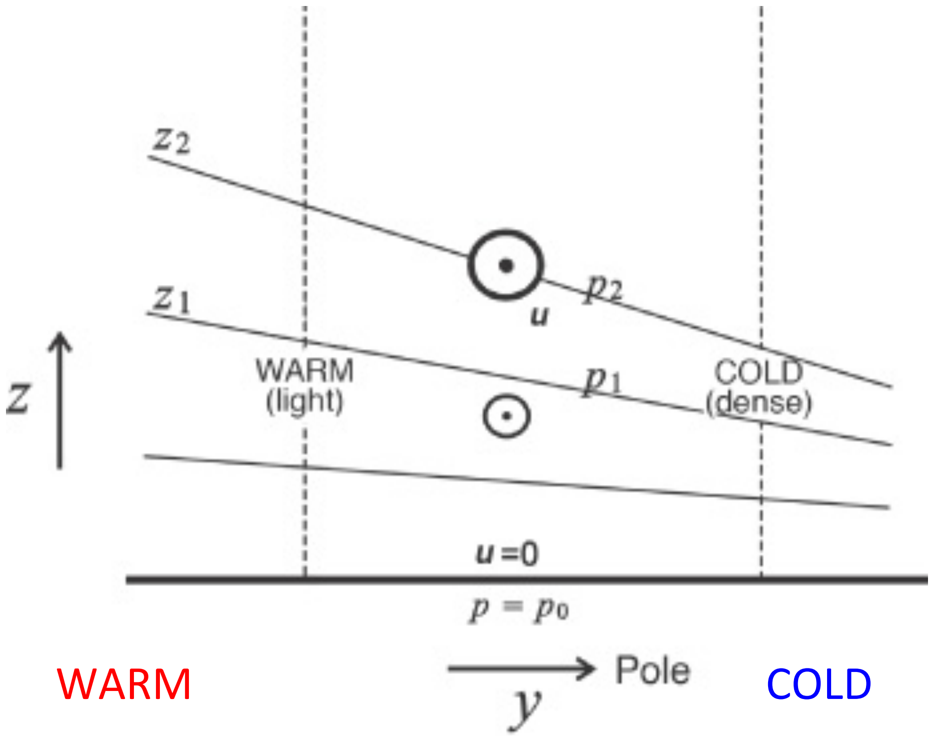
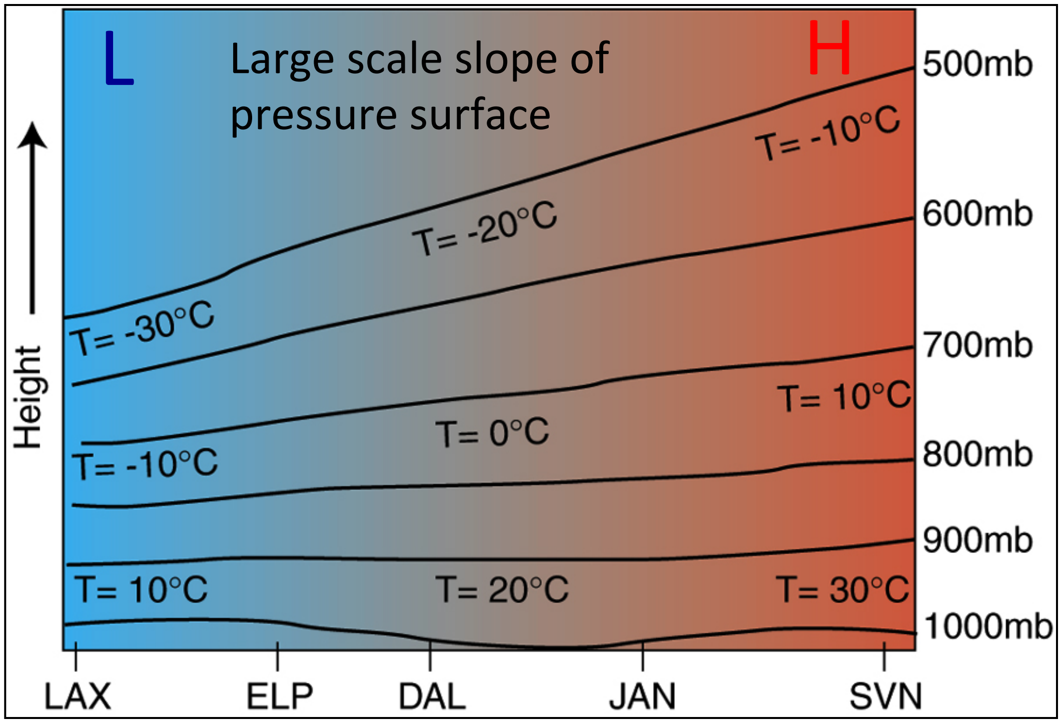


Figure 5.14: Warm columns of air expand, cold columns contract, leading to a tilt of pressure surfaces, a tilt which typically increases with height in the troposphere. In Section 7.3, we will see that the corresponding winds are out of the paper, as marked by *e* in the figure.

Each surface tilting steeper

Layer Thickness, Temperature & Wind(defined by slope)

North **DLA Fig. 7.22** South



thin

thick

COLD → WARM

more dense → less dense

Ideal Gas Law

$P = \rho R T$

$$z_2 - z_1 = R \int_{p_2}^{p_1} \frac{T}{g} \frac{dp}{p}, \quad (5\text{-}4)$$

thickness between 2 P levels varies with the mean T

westerly wind, increasing with height

Summary:

We saw how warming the tropical atmosphere and cooling over the poles leads to large scale slope of pressure surface (EQ-Pole).