Characteristics of Atmospheric Flow

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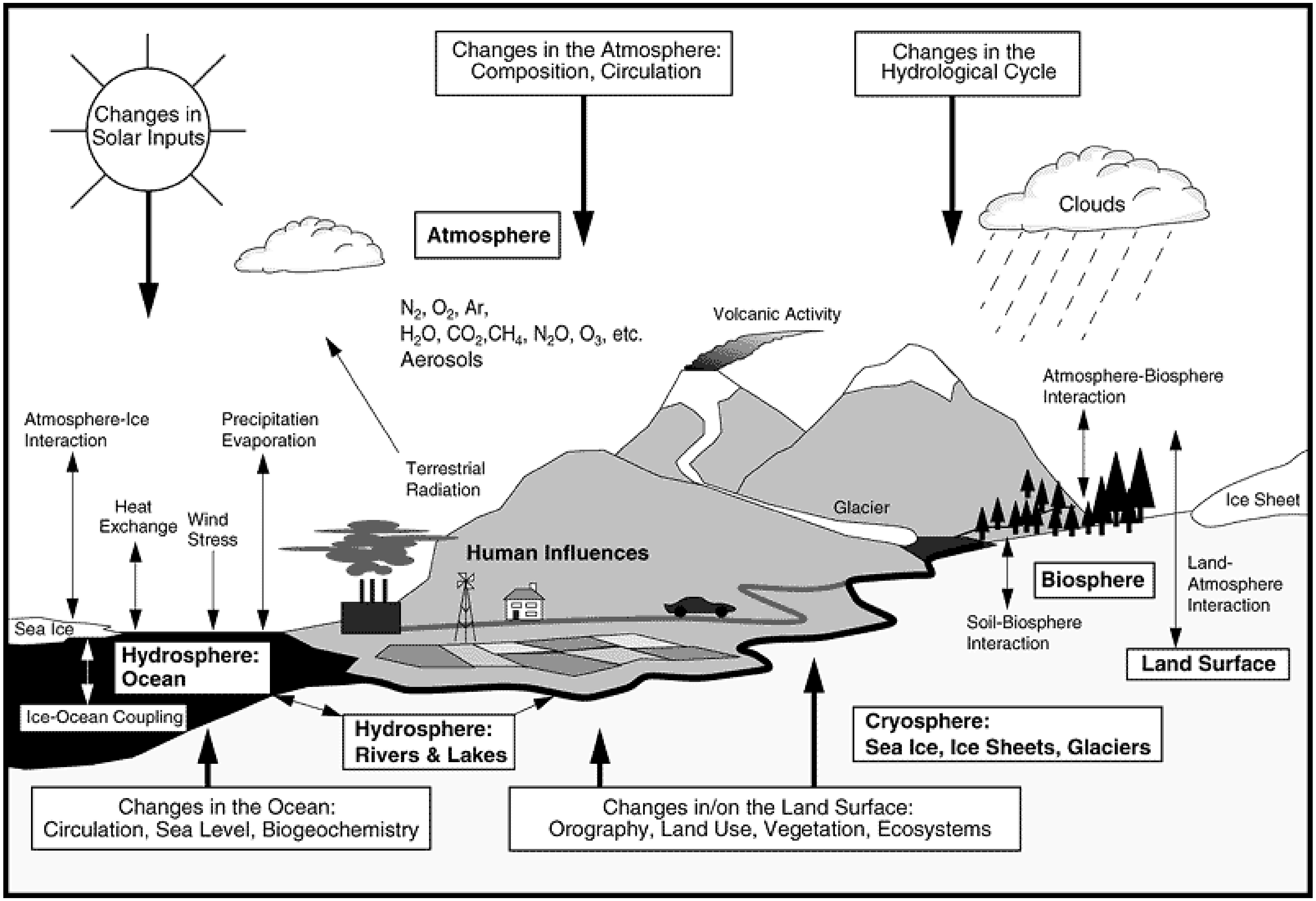
Lecture 1

Acknowledgments: The material in these lectures were taken from different sources and if possible given credit, others from web sites were lectures that did not have any clear identification.

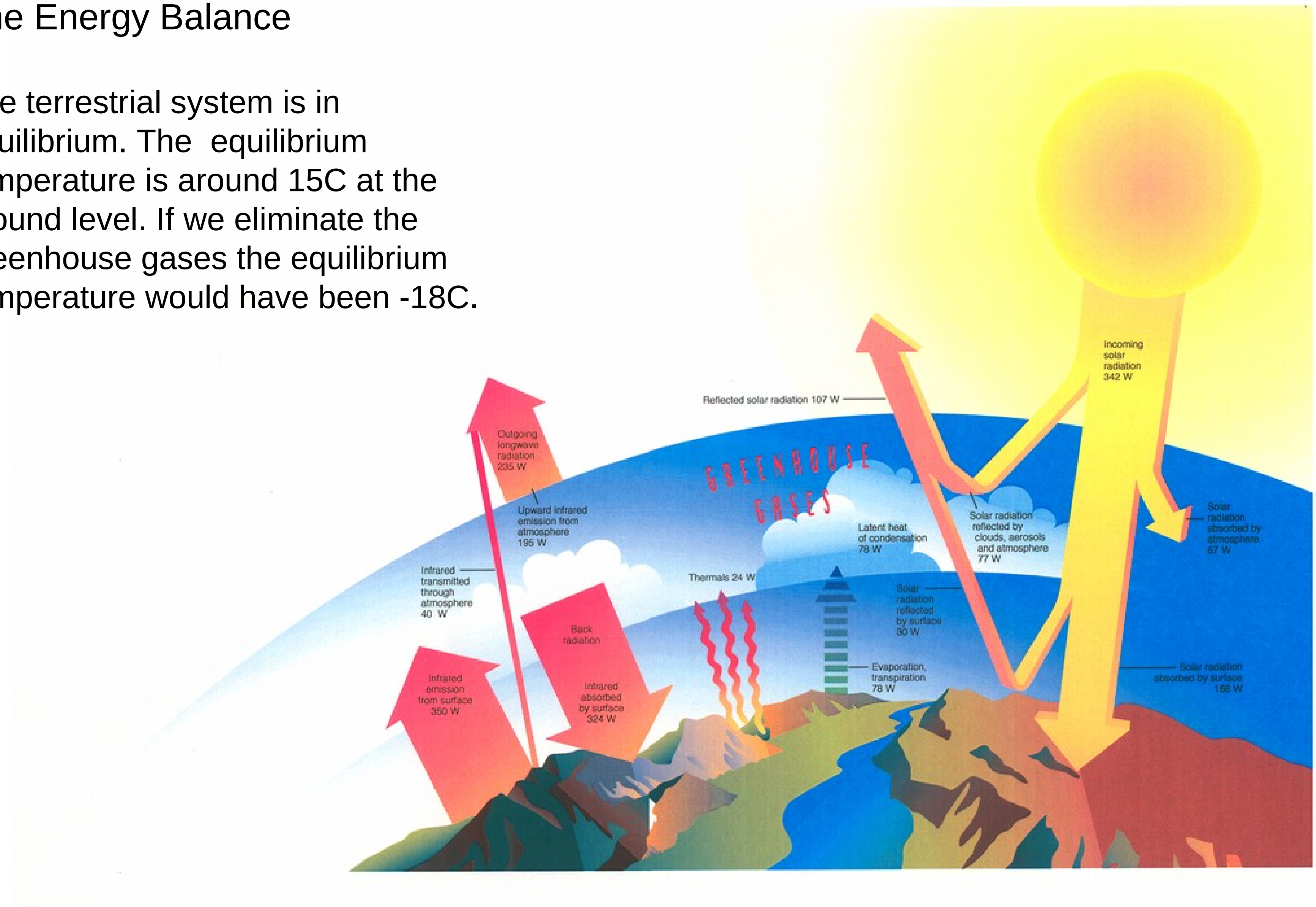
What Drives Climate Change?



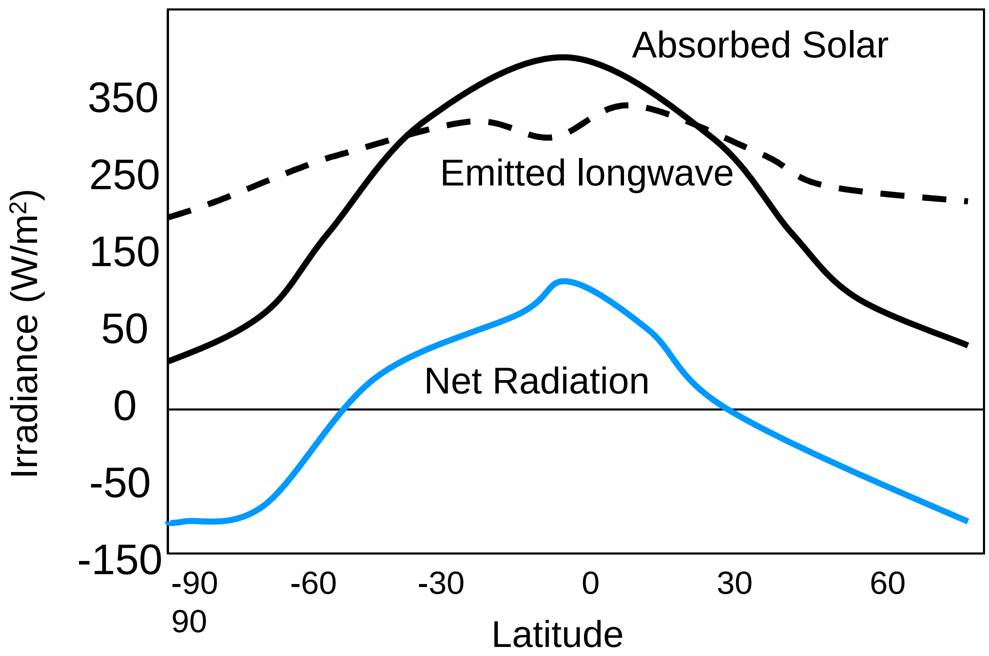
IPCC Third Assessment Report http://www.grida.no/climate/ipcc_tar/wg1/fig1-1.htm

The Energy Balance

The terrestrial system is in equilibrium. The equilibrium temperature is around 15C at the ground level. If we eliminate the greenhouse gases the equilibrium temperature would have been -18C.



Graph of annual-mean absorbed solar radiation, OLR and net radiation averaged a latitude circles



Energy Balance of the Atmosphere

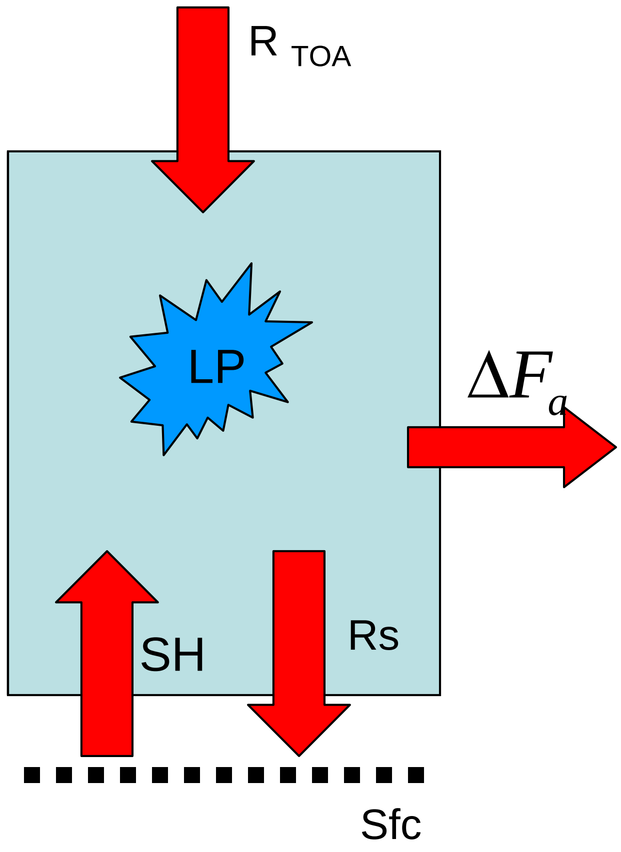
$$\frac{\partial E_a}{\partial t} = R_a + LP + SH - \Delta F_a$$

Where

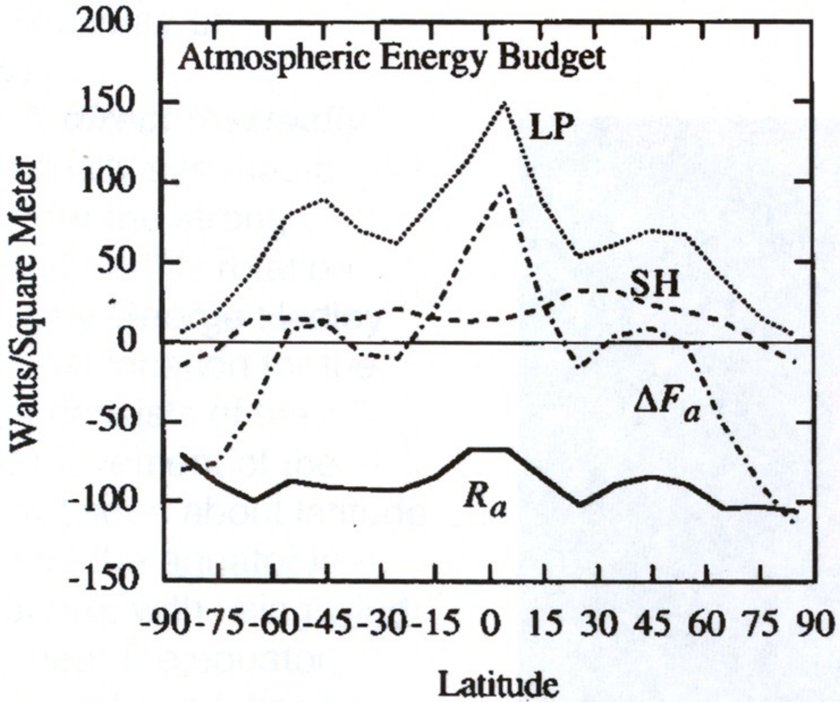
$R_a = R_{TOA} - R_s$

Energy balance

$$R_a + LP + SH \sim \Delta F_a$$



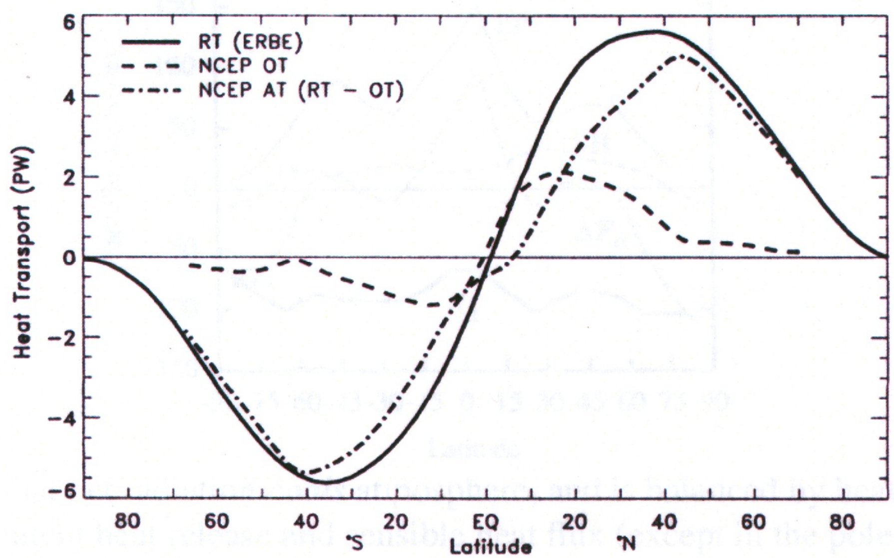
LP is the heating of the atmosphere by latent heat release due to condensation. Ea is a dry Energy- it does not include the energy content in water vapor. SH is sensible heat from the surface and Fa is the transport by the atmospheric circulation.



Note that net radiation ***cools*** atmosphere, and is balanced by heating from latent heat release and sensible heat flux (except in the poles), and atmospheric transports. Note that the equator-to-pole gradient in atmospheric heating is set up primarily by latent heat release;the atmosphere then transports heat from tropics to polar regions

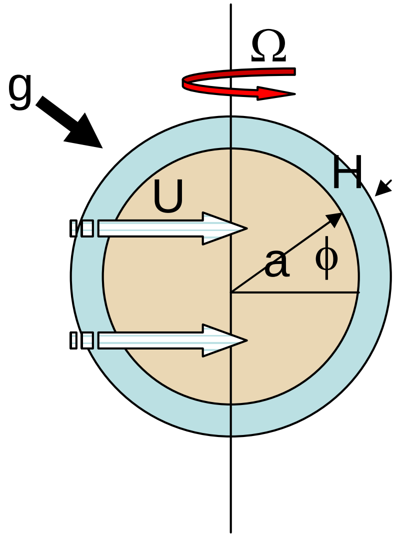
From Atmospheric circulation III, Berkeley University.

Northward heat transport across each latitude (IPW=10^{15} W)



Trenberth and Caron (J Clim v14 p3433 2001)
Solid line - atm+ocn transport; Dashed - ocn only; dash-dot - atm only

Characteristics of Atmospheric and Oceanic Dynamics



Gravitational attraction

Shallow fluid: H << a

Hydrostatic balance:
Pressure force ~ buoyancy force

Rapid rotation and large stratification

$a\Omega \sim (gH)1/2 > (NH)1/2 > U$

465m/s 330m/s 100m/s 20m/s

a= earths radius= 6730 km.
H= atmospheric depth~20km
U= relative velocity ~ 20m/s
g= gravity ~ 9.81m/s^2
Ω= rotation frequency=2π/86400s=7.272e-5s^{-1}
ϕ= latitude
f(ϕ)=coriolis parameter=2* Ω*sin(ϕ)

The Basics of convection and its parameterization

$$N^2 = g/\Theta \, d\Theta/dz$$

N Is the Buoyancy frequency
For the atmosphere and oceans
is about 0.01/sec

if the temperature profile has
$N^2 > 0$ the column is stable, if
$N^2 < 0$ is convectively unstable

Convective Instability

$$\frac{dw}{dt} = g\frac{(\rho_e - \rho_p)}{\rho_p}$$

Vertical acceleration = buoyancy

$$\frac{d\rho}{dt} = \frac{\partial \rho}{\partial t} + u\frac{\partial \rho}{\partial x} + v\frac{\partial \rho}{\partial y} + w\frac{\partial \rho}{\partial z} = 0$$

Mass conservation

$$\frac{\partial \rho'}{\partial t} + w\frac{\partial \bar{\rho}}{\partial z} = 0$$

$$\rho' = \rho_p - \rho_o$$

Assuming small perturbations, the local change of perturbation density=
the vertical advection of density change.

The buoyancy can be written as follow:

$$\frac{(\rho_e - \rho_p)}{\rho_p} = \frac{1}{\rho}\frac{d\rho}{dz}\Delta z$$

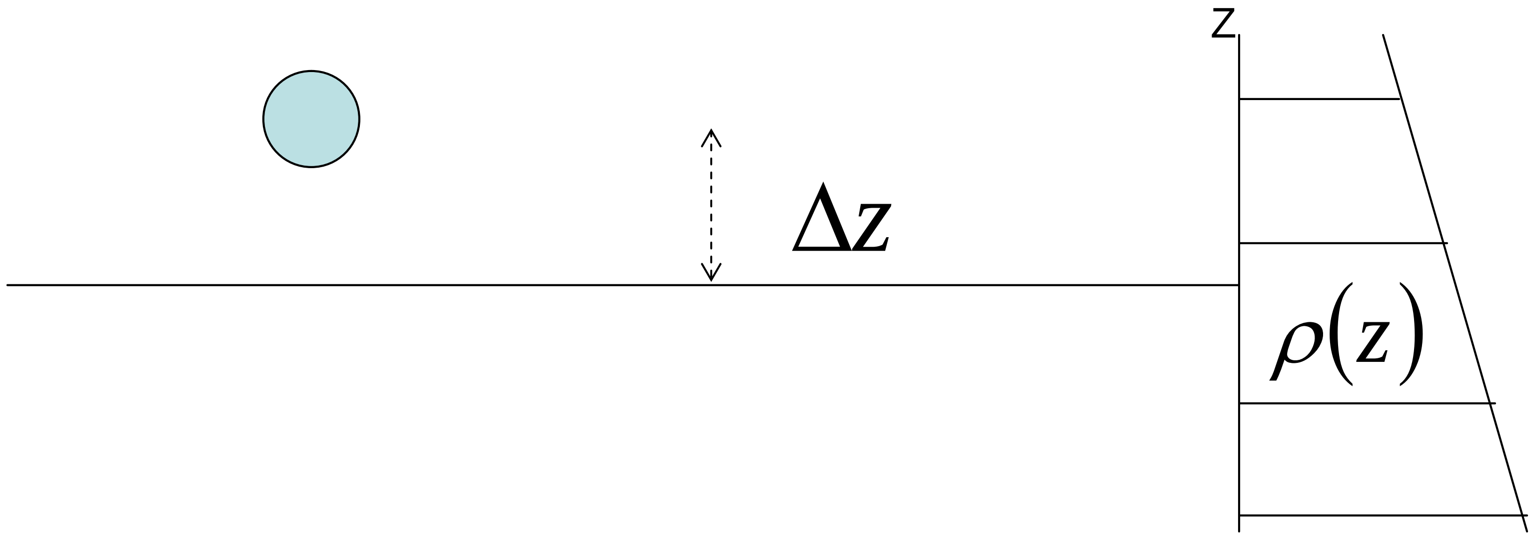
$$g\frac{(\rho_e - \rho_p)}{\rho_p} = \frac{g}{\rho}\frac{d\rho}{dz}\Delta z = -N^2\Delta z$$

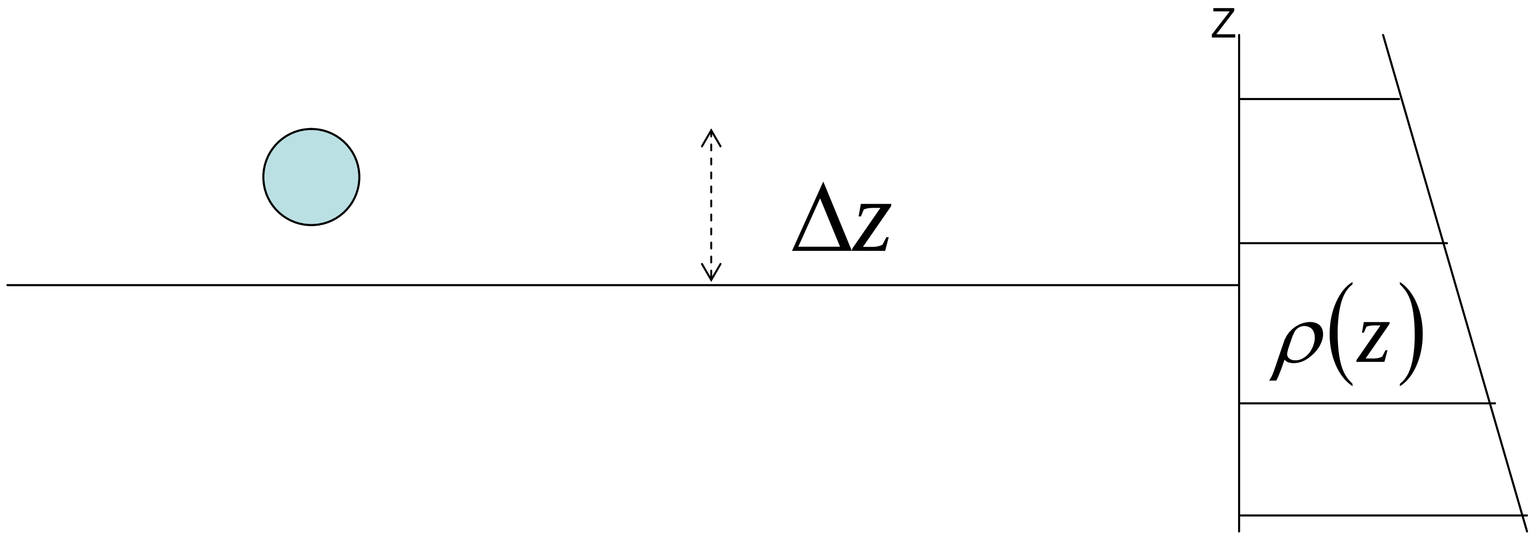
Where N^2 is the square of the Brunt Vaisala frequency;

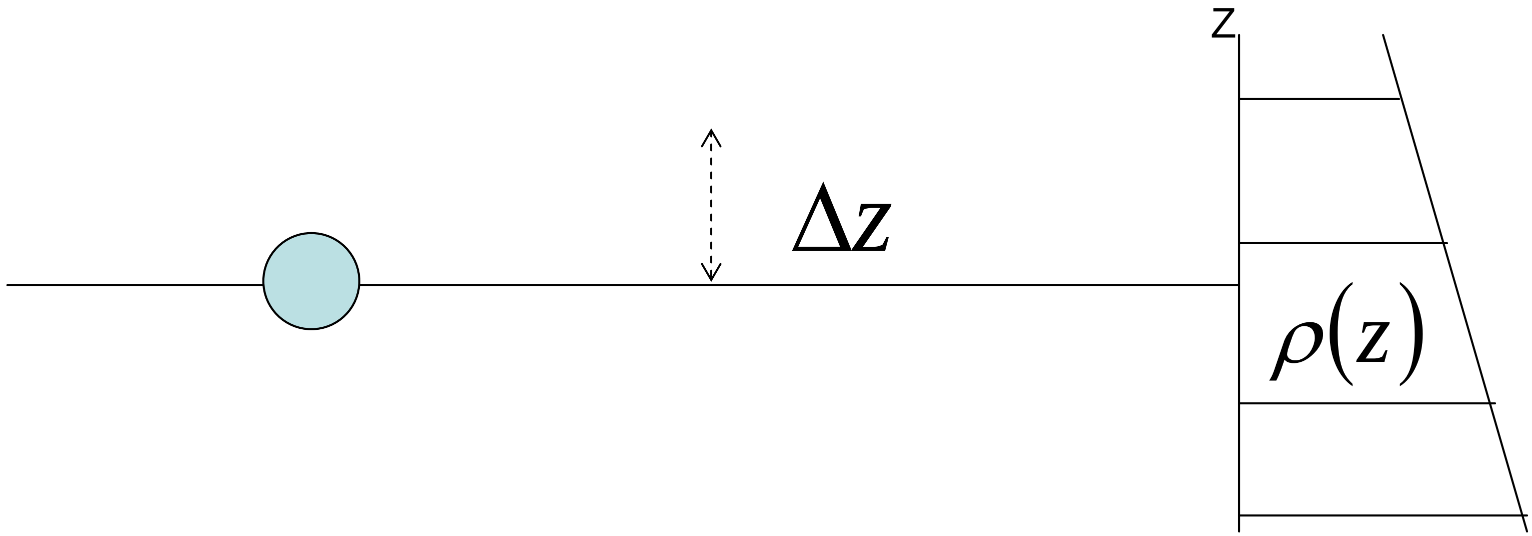
$$\frac{dw}{dt} = g\frac{(\rho_e - \rho_p)}{\rho_p} = -N^2\Delta z = \frac{d^2\Delta z}{dt^2}$$

For N2 > 0 the displacement has an oscillation with frequency +/- N and is called gravity waves. The period of N is:
~5 to 10 min for the atmosphere and oceans.

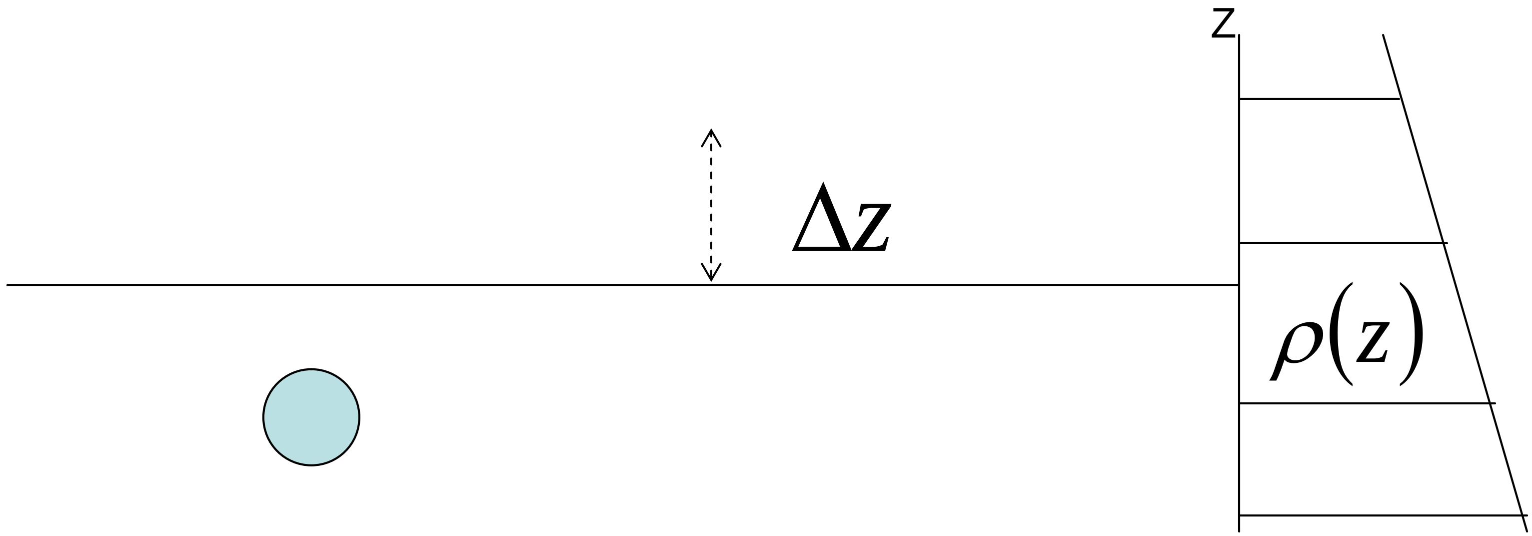
$$\Delta z = \Delta z_0 e^{i\omega t}, \omega = \pm N$$


z
Δz
ρ(z)

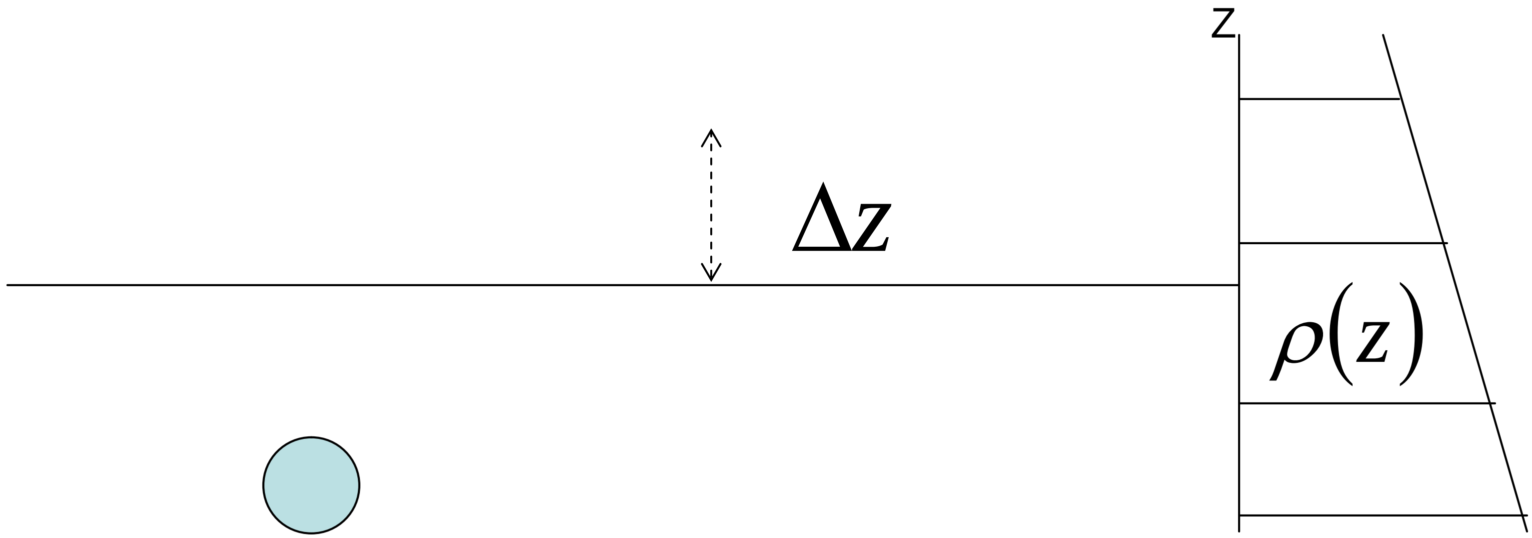

z
Δz
$\rho(z)$



Δz
z
$\rho(z)$



z
Δz
$\rho(z)$



z
Δz
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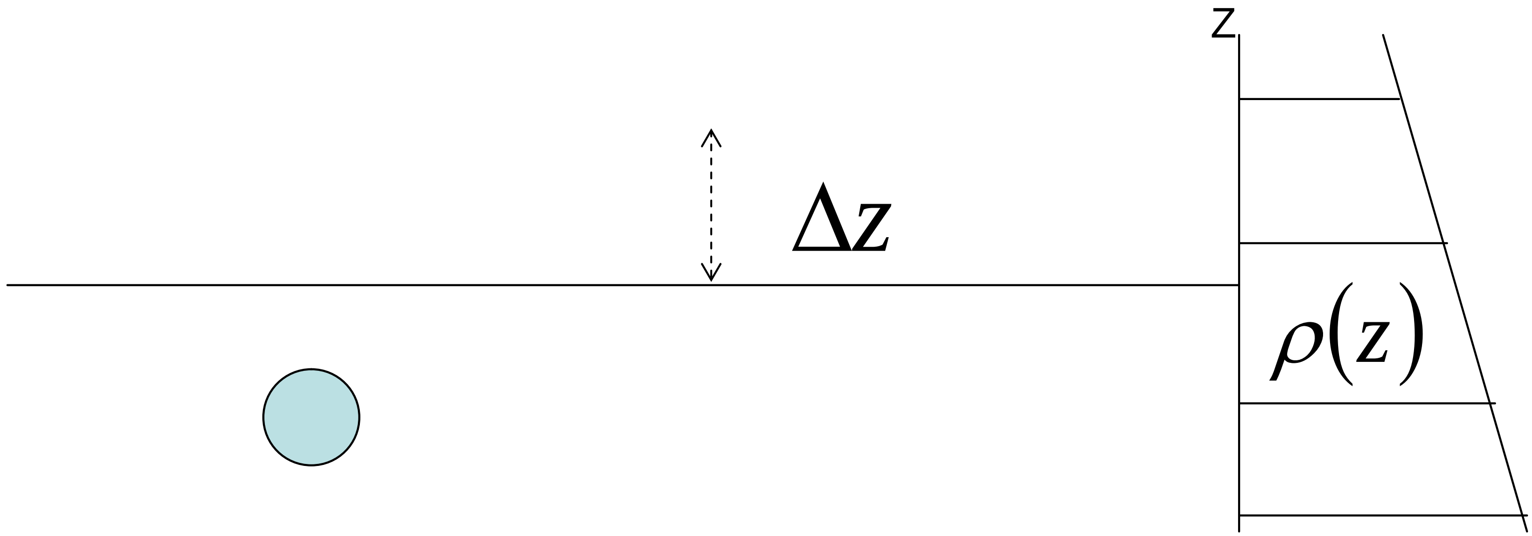
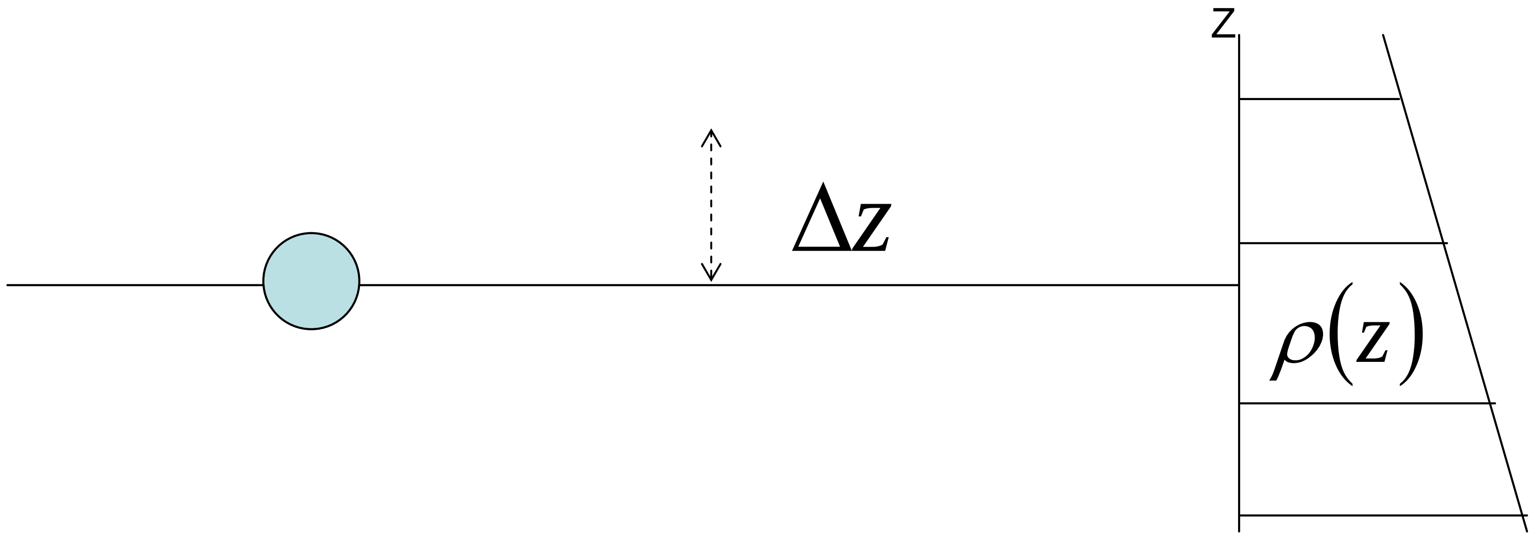
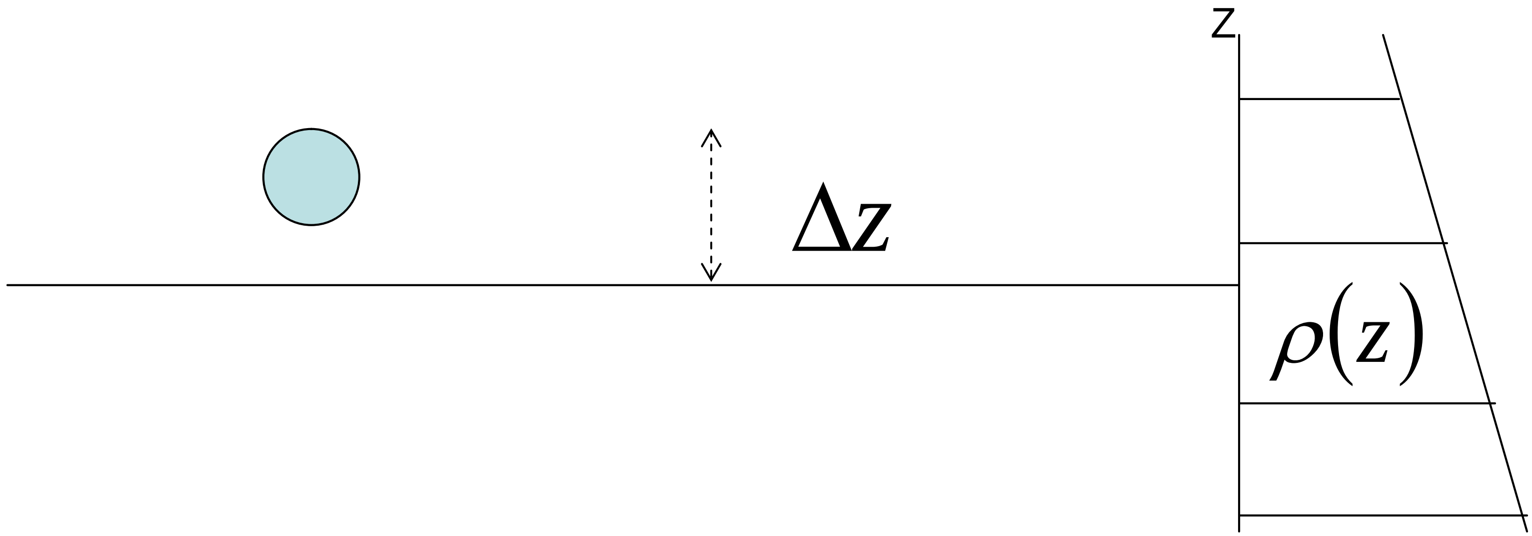


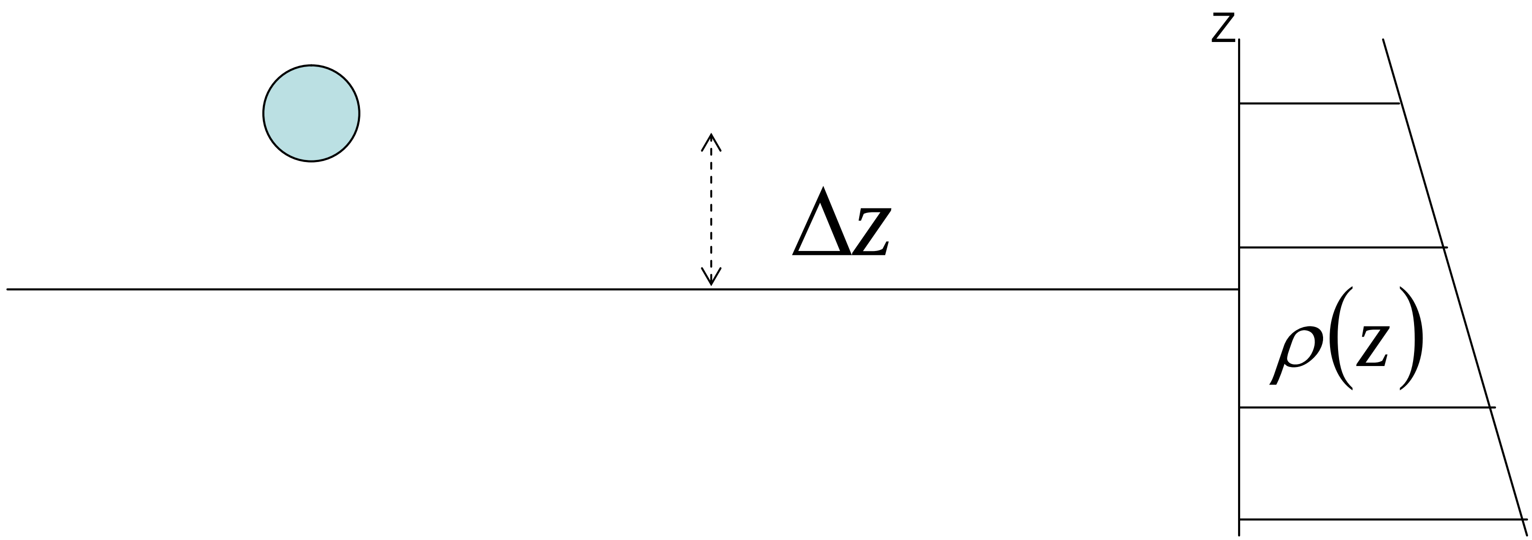
z
Δz
$\rho(z)$



Z
Δz
$\rho(z)$



z
Δz
ρ(z)



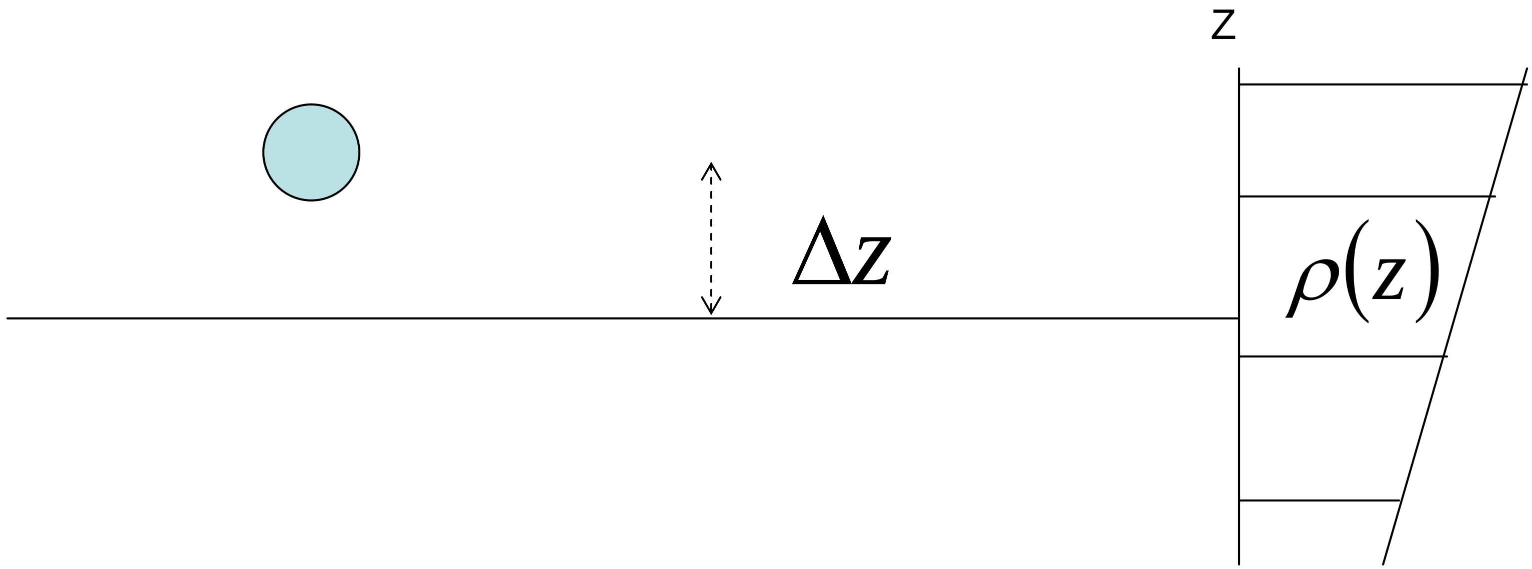
z
Δz
ρ(z)

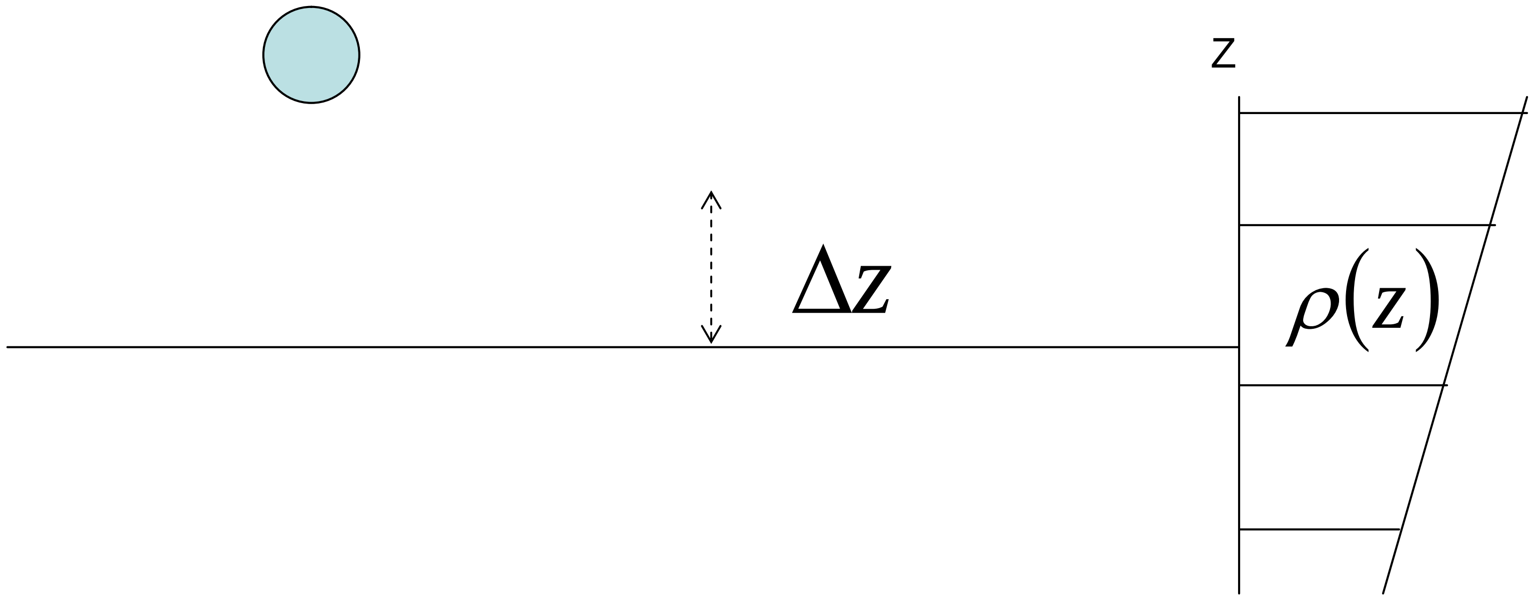
Gravitational or convective Instability

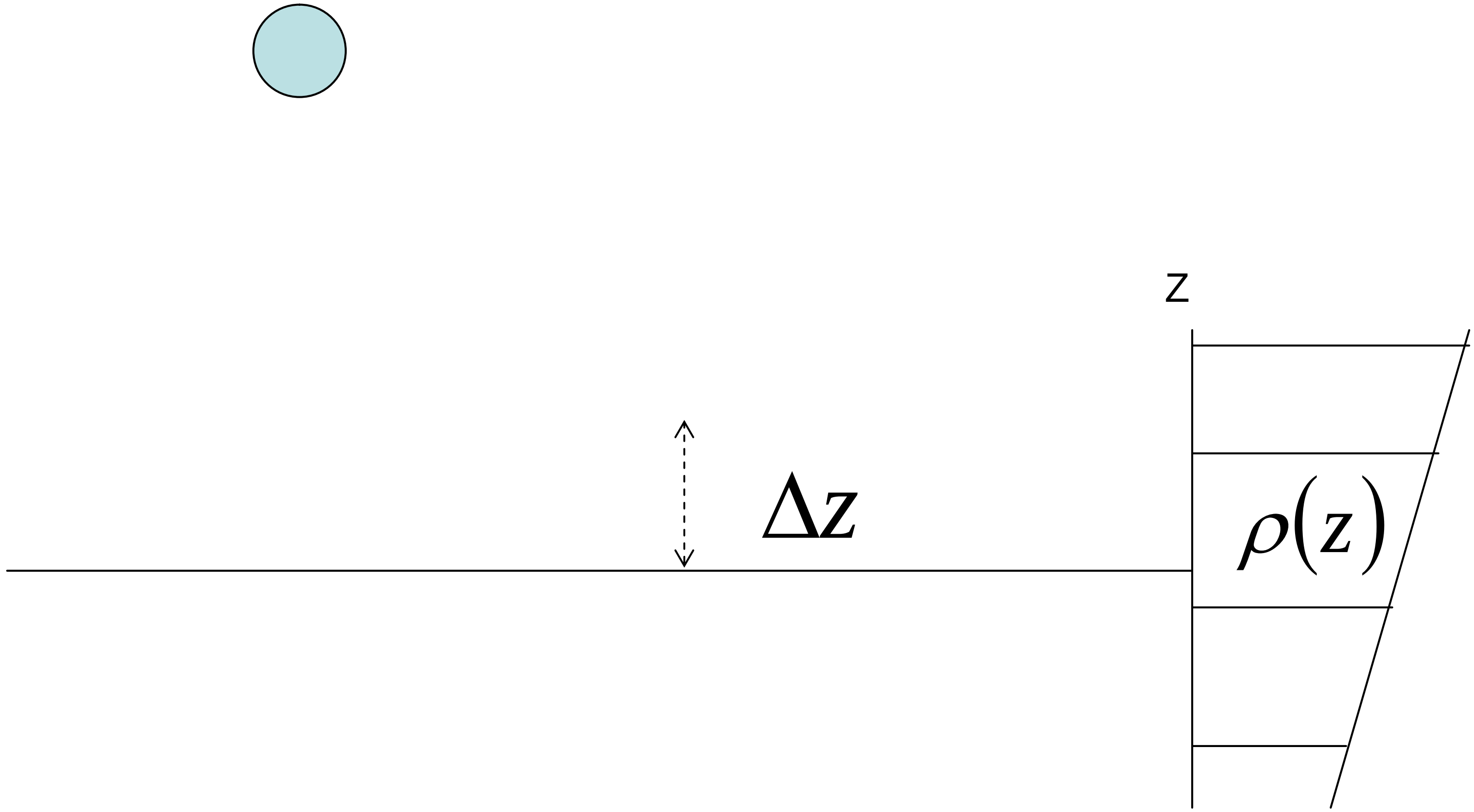
$$\frac{d\rho}{dz} > 0, N^2 < 0$$

$$\Delta z = \Delta z_0 e^{\pm \alpha t}$$

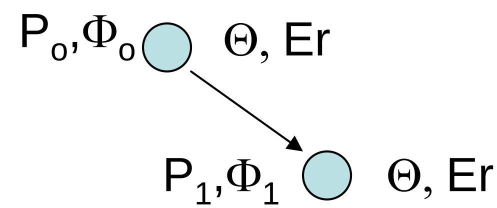
Where α is real and the solution are exponential no oscillations.
This is the basic of most of the atmospheric and ocean convection.


Δz
z
$\rho(z)$


Δz
z
$\rho(z)$


z
Δz
$\rho(z)$

Lagrangian conservation properties for adiabatic frictionless fluid in a rotating sphere:



Entropy or Potential temperature Θ

$\Theta = T\,(Po/P)^{\kappa}$, $\kappa = R/cp = 287/1004 = 0.286$

Potential Vorticity $Er = \rho^{-1}(2\Omega + \text{curl } u).\text{grad}\Theta$

Conservation of mass
$d\rho/dt = -\rho * \text{Div}(u)$

Incompressible fluid

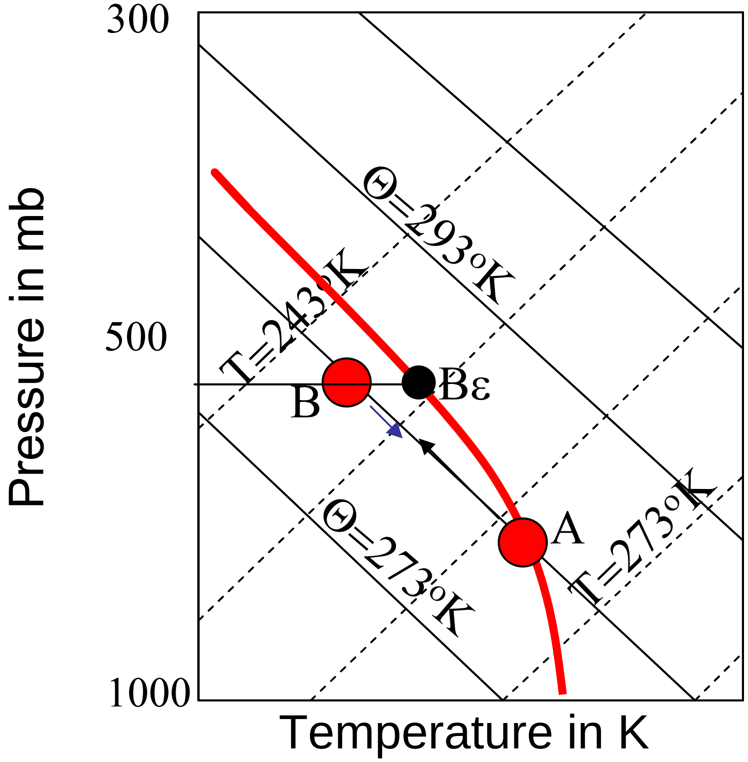
Div(u) = 0

$d\rho/dt = 0$

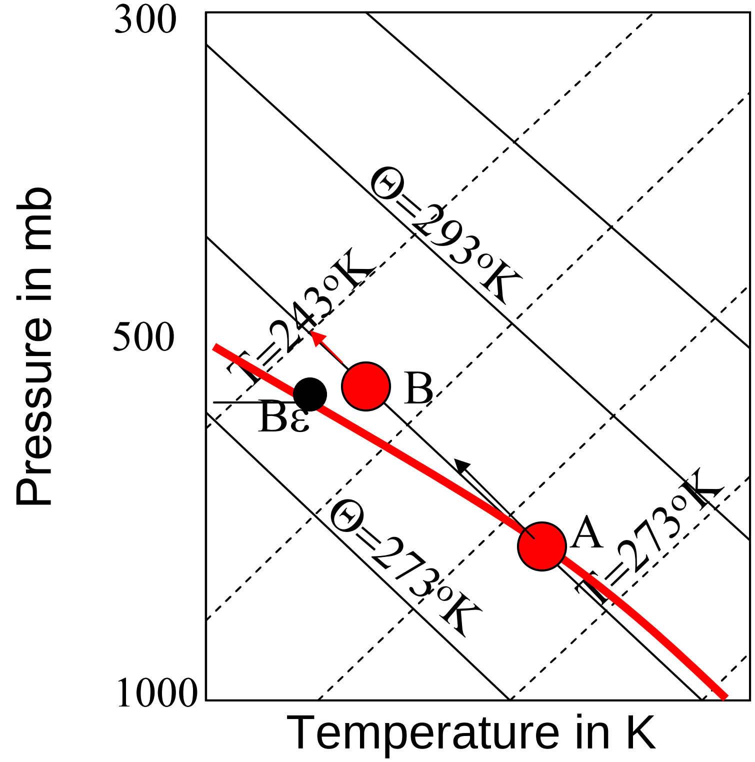
Compressible fluid

$d\Theta/dt = 0$

Statically stable



Statically unstable



$$c_p T \frac{d\theta}{\theta} = c_p dT + g dz$$

$$\frac{1}{\theta}\frac{d\theta}{dz} = \frac{1}{T}\left(\frac{dT}{dz} + \frac{g}{c_p}\right)$$

$$\frac{1}{\theta}\frac{d\theta}{dz} = \frac{1}{T}(\Gamma_d - \Gamma)$$

Something about moisture

If we let q_s denote the mass of water vapor per unit mass of dry air in a saturated parcel (q_s is called the saturation mixing ratio). Then the rate of diabatic heating per unit mass is:

$$J = -L_c \frac{Dq_s}{Dt}$$

Where Lc is the latent heat of condensation, From the first law of thermodynamics:

$$c_p \frac{D\vartheta}{Dt} = -\frac{L_c}{T}\frac{Dq_s}{Dt}$$

For a saturated parcel undergoing pseudoadiabatic ascent the rate of change of qs following the motion is much larger than the rate of change of Lc or T then:

$$d(\ln \vartheta) \approx -d\left(\frac{L_c q_s}{c_p T}\right)$$

Integrating this equation from the initial state (θ,q_s,T) to a state where q_s~0

$$\ln\left(\frac{\vartheta}{\theta_e}\right) \approx -\frac{L_c q_s}{c_p T}$$

Then the Equivalent Potential Temperature Θ_e for a saturated parcel is given:

$$\theta_e \approx \vartheta e^{\left(\frac{L_c q_s}{c_p T}\right)}$$

The Q_e can be related to the moist static energy $h=s+L_c q$, where s is the dry static energy $s=c_p T+gz$

$$c_p T d\left(\ln \theta_e\right) = d\left(h\right)$$

The moist static energy is approximately conserved when Θ_e is conserved.

The Psedoadiabatic lapse rate

Using the definition of Q we can derive

$$\frac{d\ln\vartheta}{dz} - \frac{R}{c_p}\frac{d\ln p}{dz} = -\frac{L_c}{c_p T}\frac{dq_s}{dz}$$

And since $q_s(T,p)$ and using the hydrostatic equation and equation of state

$$\frac{dT}{dz} + \frac{g}{c_p} = -\frac{L_c}{c_p}\left[\left(\frac{\partial q_s}{\partial T}\right)_p \frac{dT}{dz} - \left(\frac{\partial q_s}{\partial p}\right)_T \rho g\right]$$

One can derive the ascending saturated lapse rate

$$\Gamma_s = -\frac{dT}{dz} = \Gamma_d \frac{\left[1 + L_c q_s/(RT)\right]}{\left[1 + \varepsilon L_c^2 q_s/\left(c_p R T^2\right)\right]}$$

Where e=0.622 is the ratio of molecular weight of water to that of dry air. Γ_d=g/cp. Observed range of Γ_s values range from ~4Km^{-1} in warm humid air masses in the lower atmosphere to ~6-7Km^{-1} in the midtroposphere.

Conditional Instability

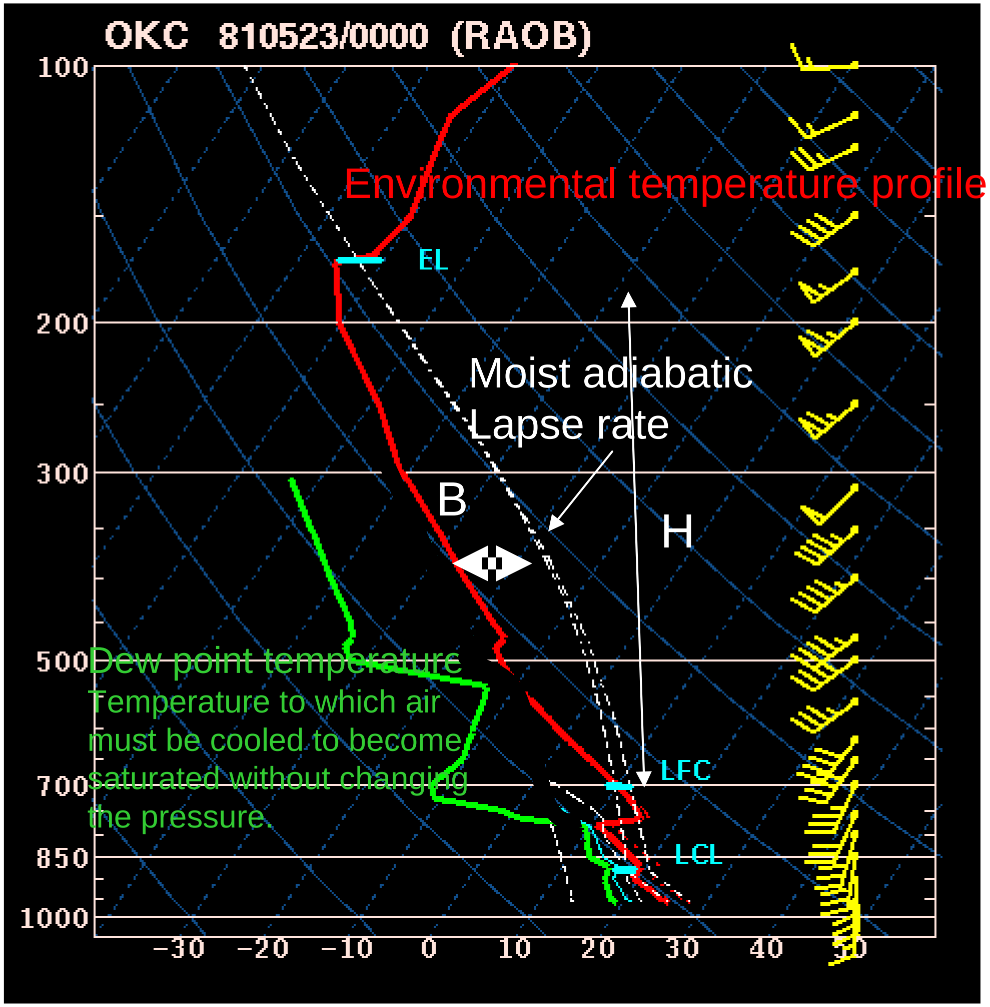
$$\Gamma_s < \Gamma_d \leq \Gamma$$ Absolutely unstable

$$\Gamma_s < \Gamma \leq \Gamma_d$$ Conditionally unstable

$$\Gamma < \Gamma_s < \Gamma_d$$ Absolutely stable

These condition can be expressed as function of the vertical gradient of Θ_e

$$\frac{\partial \theta_e}{\partial z} \left[\begin{array}{l} < 0 \text{ Conditional unstable} \\ = 0 \text{ Saturated neutral} \\ > 0 \text{ conditional stable} \end{array} \right.$$



$B=g*(T-T_m)/T_m$, buoyancy

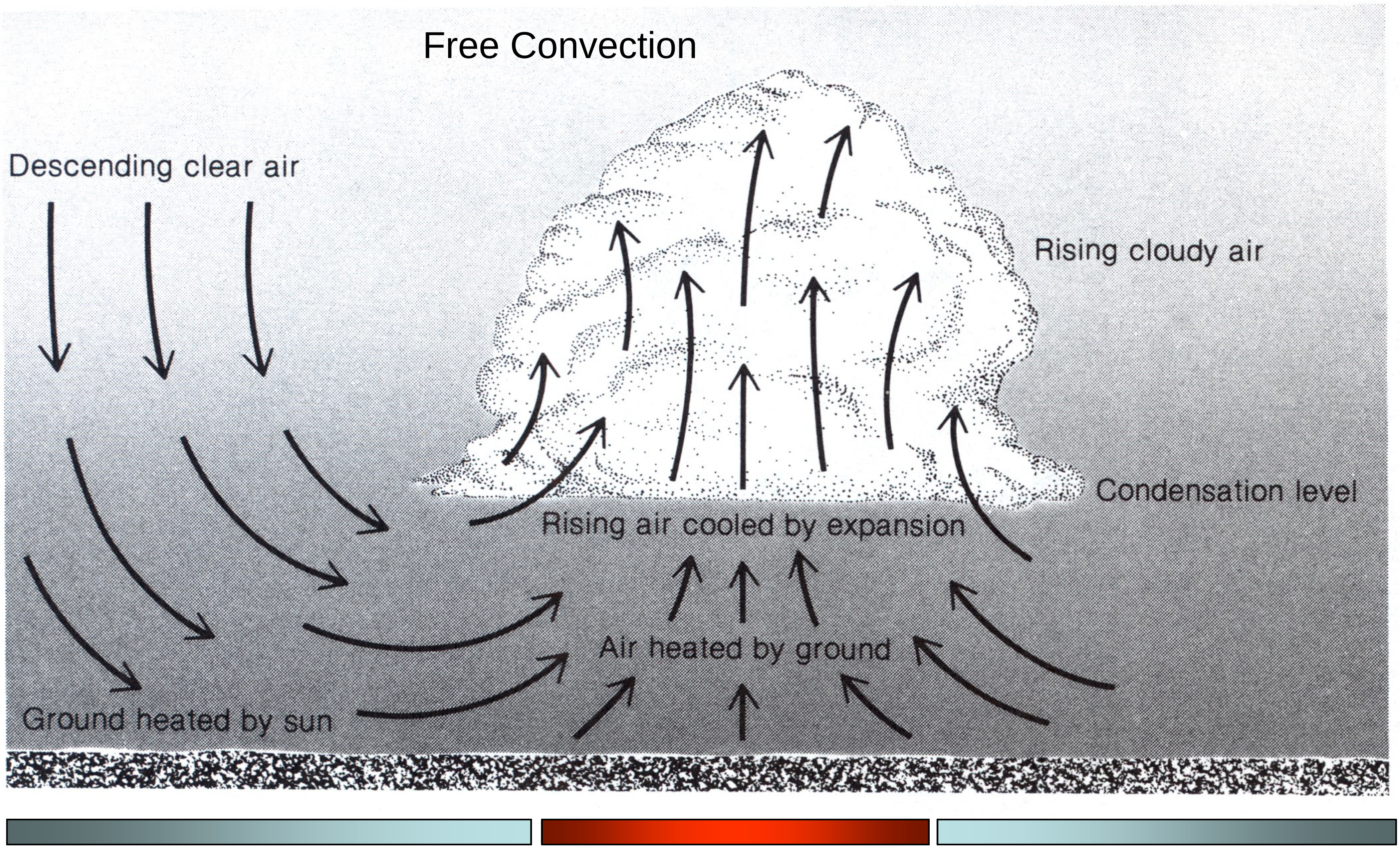
CAPE=B*H

Cape and its relation to vertical velocity.

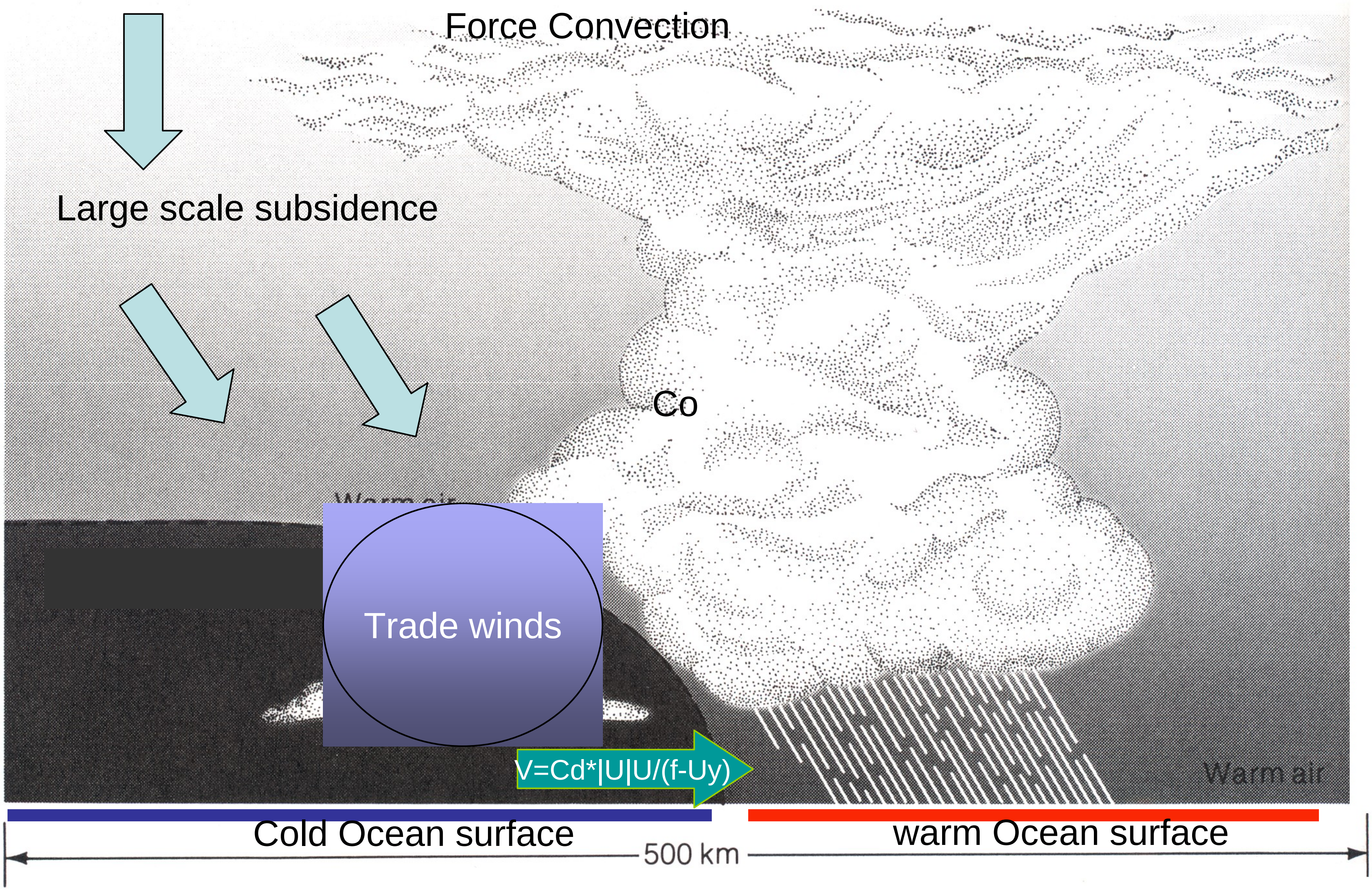
H/L < 1 $w \sim H/L(BH)^{0.5}$

H/L >1 $w \sim (BH)^{0.5}$

Free Convection



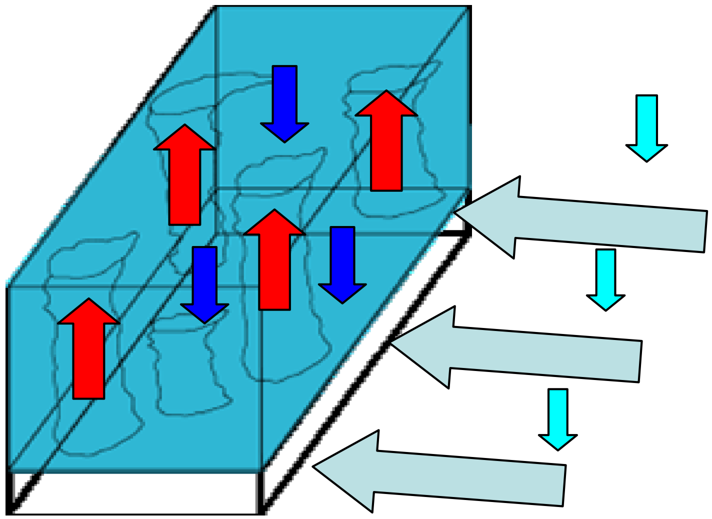
Surface temperature



Force Convection
Large scale subsidence
Co
Trade winds
V=Cd*|U|U/(f-Uy)
Warm air
Cold Ocean surface
warm Ocean surface
500 km

Explicit and Parameterized Convection

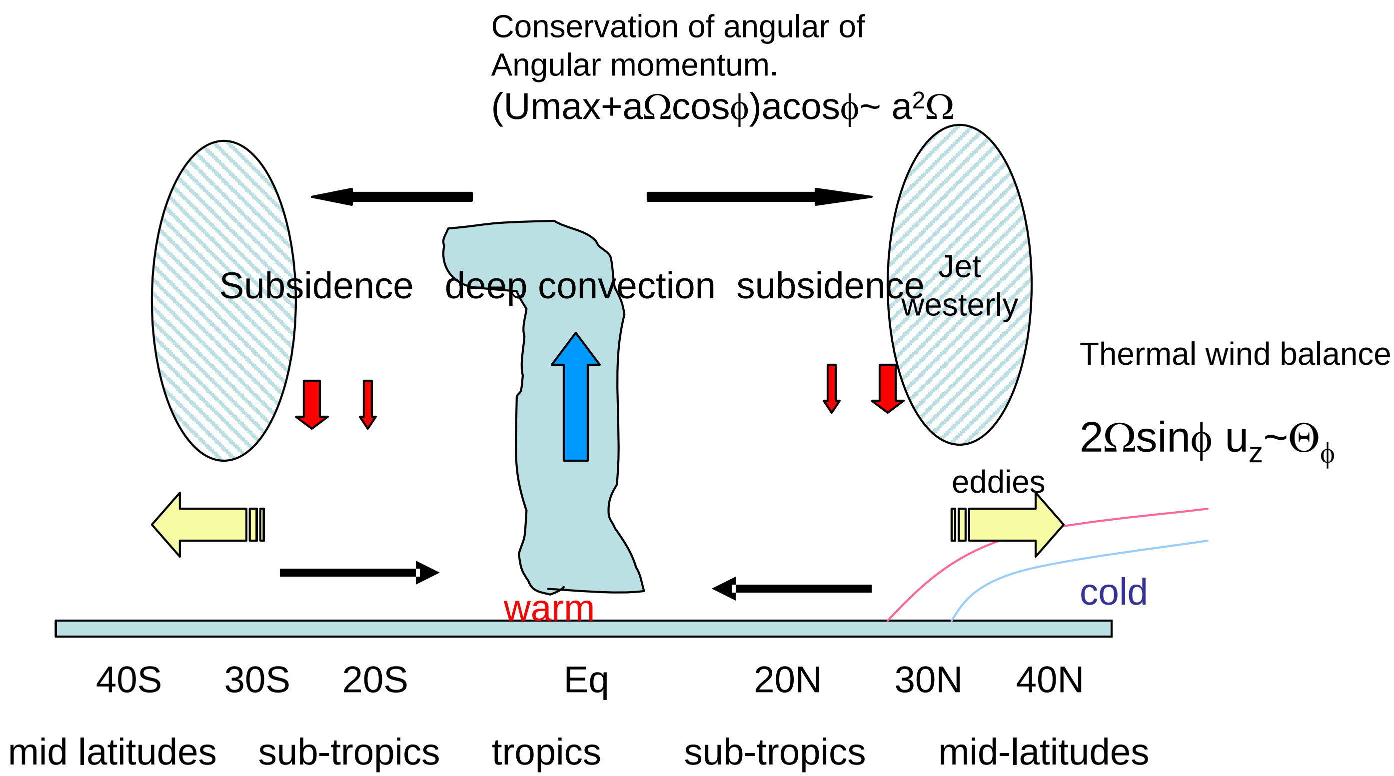
Explicit Convection



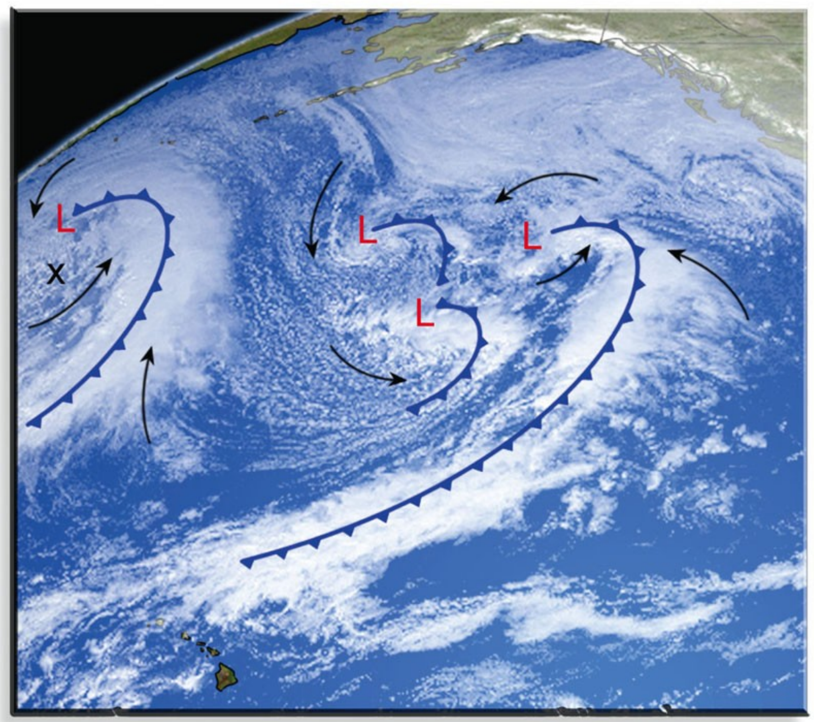
In each cloud, advection of heat and moisture from the surface is transported to the interior of the atmosphere. If condenses produce latent heat. This heat is compensated in large amount by the adiabatic cooling of the parcel that it is ascending. The descending air is produced usually in a larger region outside of the cloud.

f

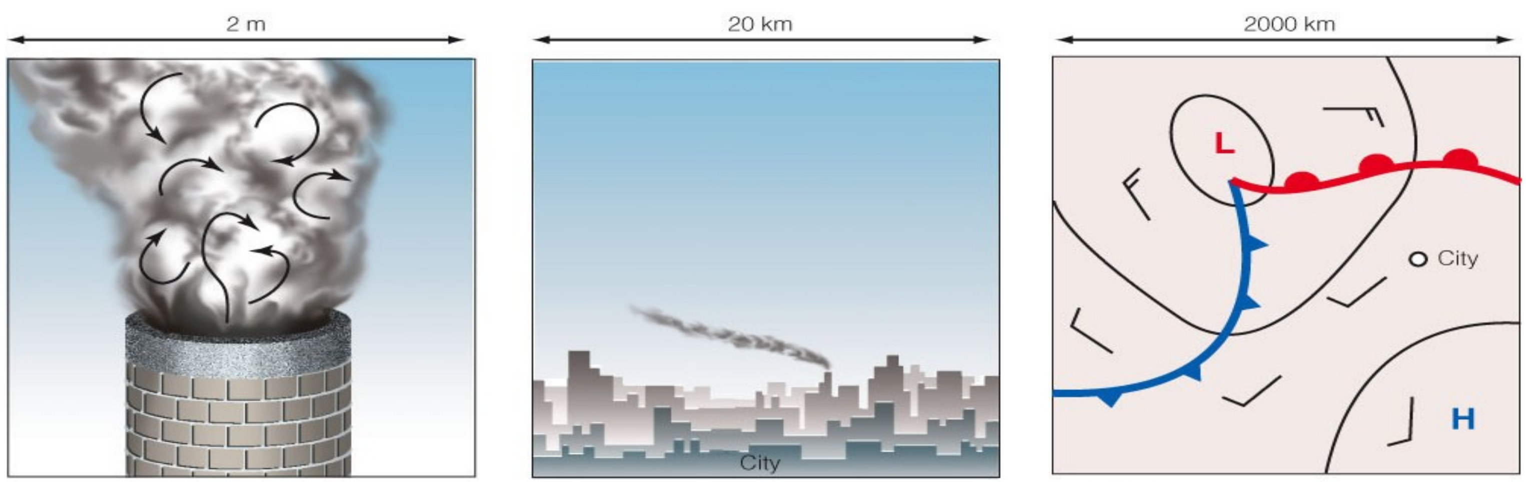
Hadley Cells and tropical temp gradient



– **General Circulation of the Atmosphere**
– **Jet Streams**
– **Atmosphere-Ocean Interactions**



- Scales of Motion
  - **How large of an area does a weather system affect?**
    - **Synoptic scale motions cover the US, 100’s to 1000’s of km.**
    - **Mesoscale motions are the size of thunderstorms, 10’s of km.**
    - **Microscale motions are 10’s to 100’s of meters.**



(a) Microscale (b) Mesoscale (c) Synoptic scale

- General Circulation of the Atmosphere
  - **General circulation of the atmosphere only represents the average (daily or monthly) air flow around the world and actual winds may vary significantly at any one place and time from this average.**
  - **It is difficult to “see” the large scale motions when all the “noise” of the smaller scale motions are super-imposed on the large scale motions.**
  - **Averages over several days or months are a way to remove the “noise” so the large scale motions are visible.**

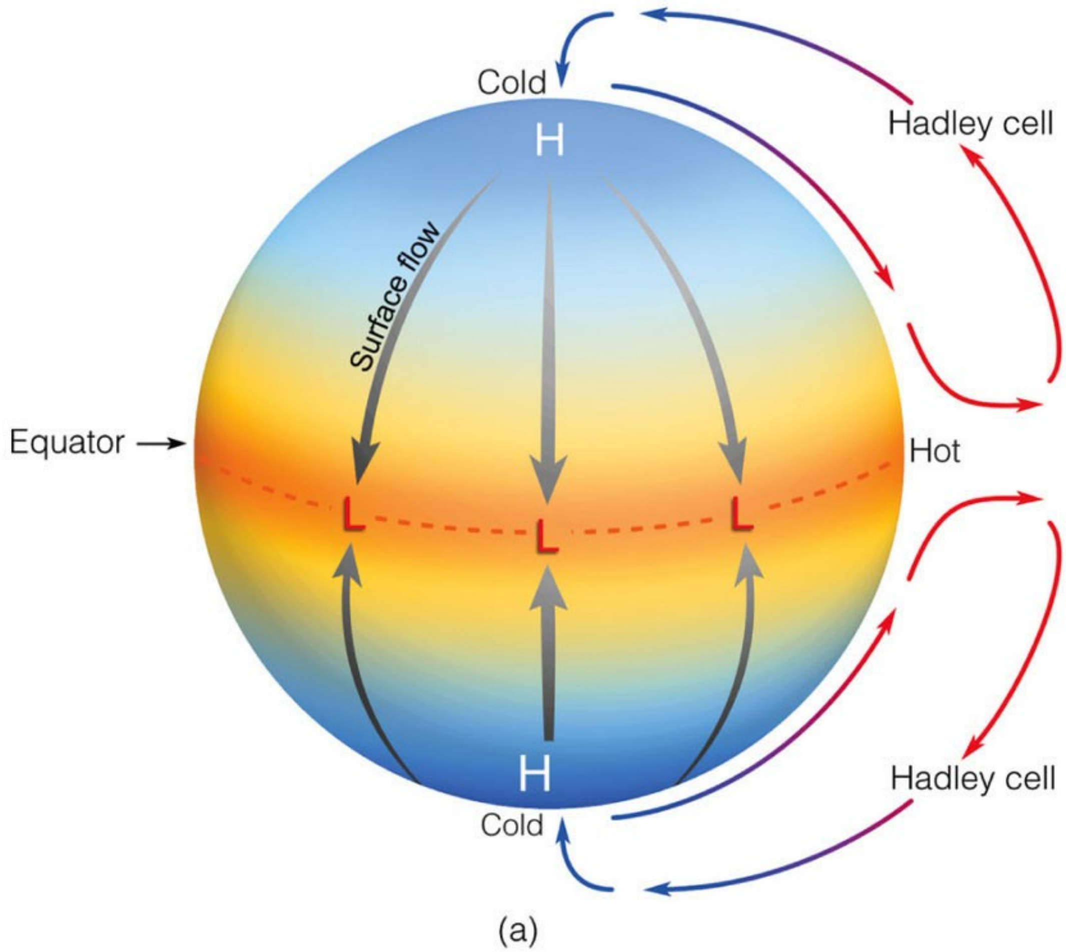
- General Circulation of the Atmosphere
  - **In that discussion it was shown that there was an uneven heating of the earth.**
    - **The energy balance is not maintained for each latitude as there is a net gain of energy at the tropics and a net loss at the poles.**
    - **As a result the poles are cool and the equator is warm.**
    - **The question is how does the earth restore the balance since we know that the poles are not at absolute zero and the equator does not melt lead.**
  - **To restore balance, the atmosphere transports warm air poleward and cool air equatorward.**
  - **We have seen before that, averaged over the entire earth, incoming solar radiation is roughly equal to the outgoing earth's radiation.**

- General Circulation of the Atmosphere (Single-Cell Model)
  - **The british meteorologist George Hadley to try to understand how the earth comes into balance with a simple model.**
  - **Single-Cell model assumes that:**
    - **Earth’s surface is uniformly covered with water (so that differential heating between the land and water does not come into play).**
    - **The sun is always directly over the equator (so that the winds will not shift seasonally).**
    - **The earth does not rotate (so that the only force we need to deal with is the pressure gradient force).**

- General Circulation of the Atmosphere (Single-Cell Model)
  - **For the single-cell model described above, the air near the equator heats and as it does it becomes less dense.**
  - **The air near the equator then begins to rise and the surface pressure drops.**
  - **Conversely, the air near the poles cools and becomes more dense.**
  - **The air near the poles then begins to sink and the surface pressure rises.**

- General Circulation of the Atmosphere (Single-Cell Model)
  - **At this point, a pressure gradient force develops between the poles and the equator and the cold air flows from the high pressure at the poles to the low pressure at the equator.**
  - **The opposite occurs in the upper portion of the troposphere. The rising air at the equator strikes the tropopause and must flow both north and south towards the poles which creates high pressure aloft.**
  - **At the poles, the sinking air creates low pressure aloft and another pressure gradient develops in the upper part of the troposphere between the poles and equator.**

- General Circulation of the Atmosphere (Single-Cell Model)



- General Circulation of the Atmosphere (Single-Cell Model)
  - **Obviously, this circulation does not really occur on the earth.**
    - **Earth is not uniformly covered with water.**
    - **The sun does not always stay above the equator.**
    - **Coriolis forces will turn the winds to the right.**
  - **However, Hadley's *model* did provide grounds for further study.**

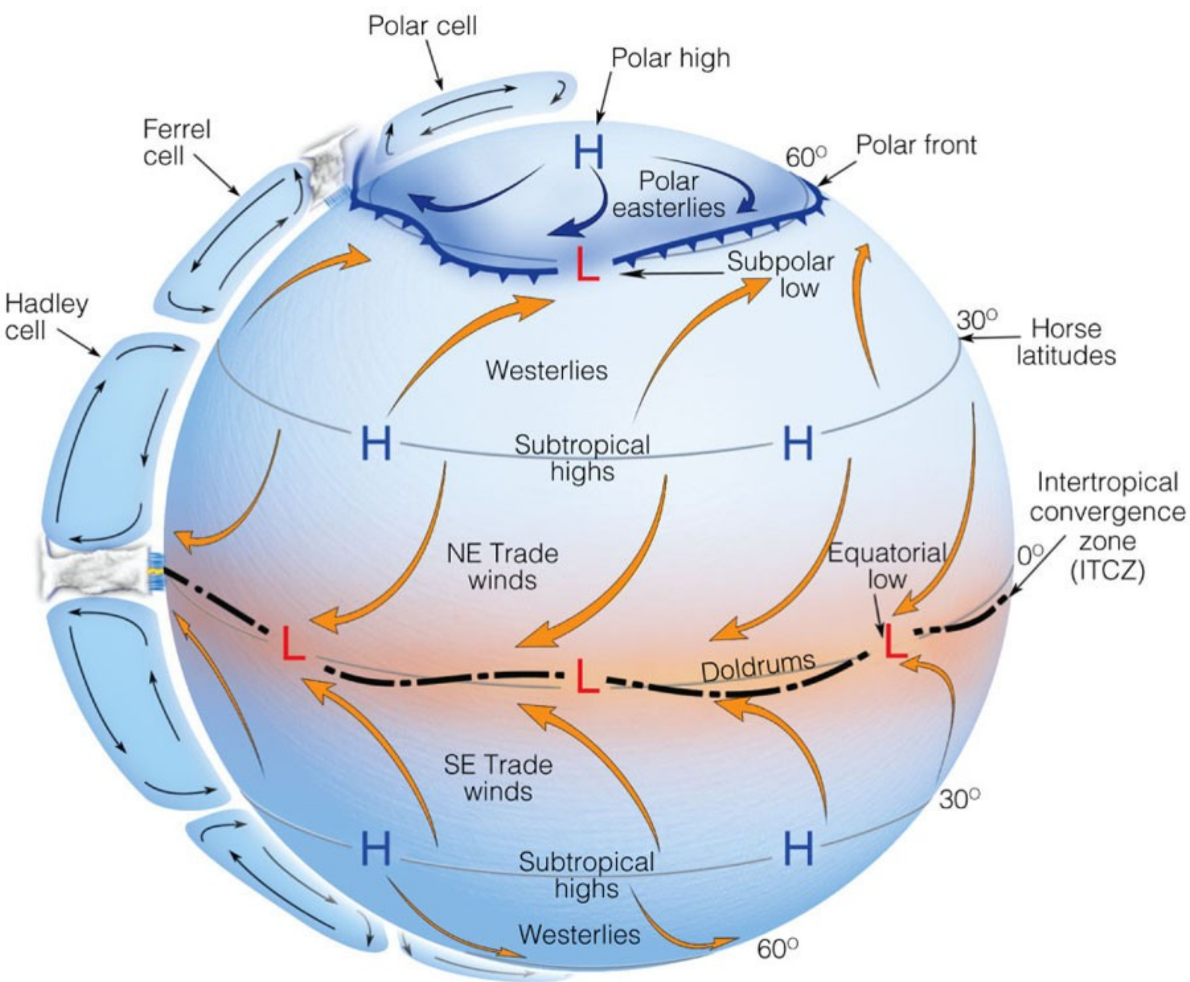
- General Circulation of the Atmosphere (Three-Cell Model)
  - **In the three-cell model, the first step is to remove the restriction of a non-rotating earth.**
  - **Although this model is much more complex than the single-cell model, there are some similarities.**
  - **As with the single cell model, the air at the equator is heated by the sun, becomes less dense, and rises to the tropopause creating a low pressure at the surface.**
  - **In addition, above the equator, the rising air creates high pressure aloft.**
  - **The poleward flow of air at upper-levels is then deflected by the Coriolis force toward the right in the NH and to the left in the SH.**

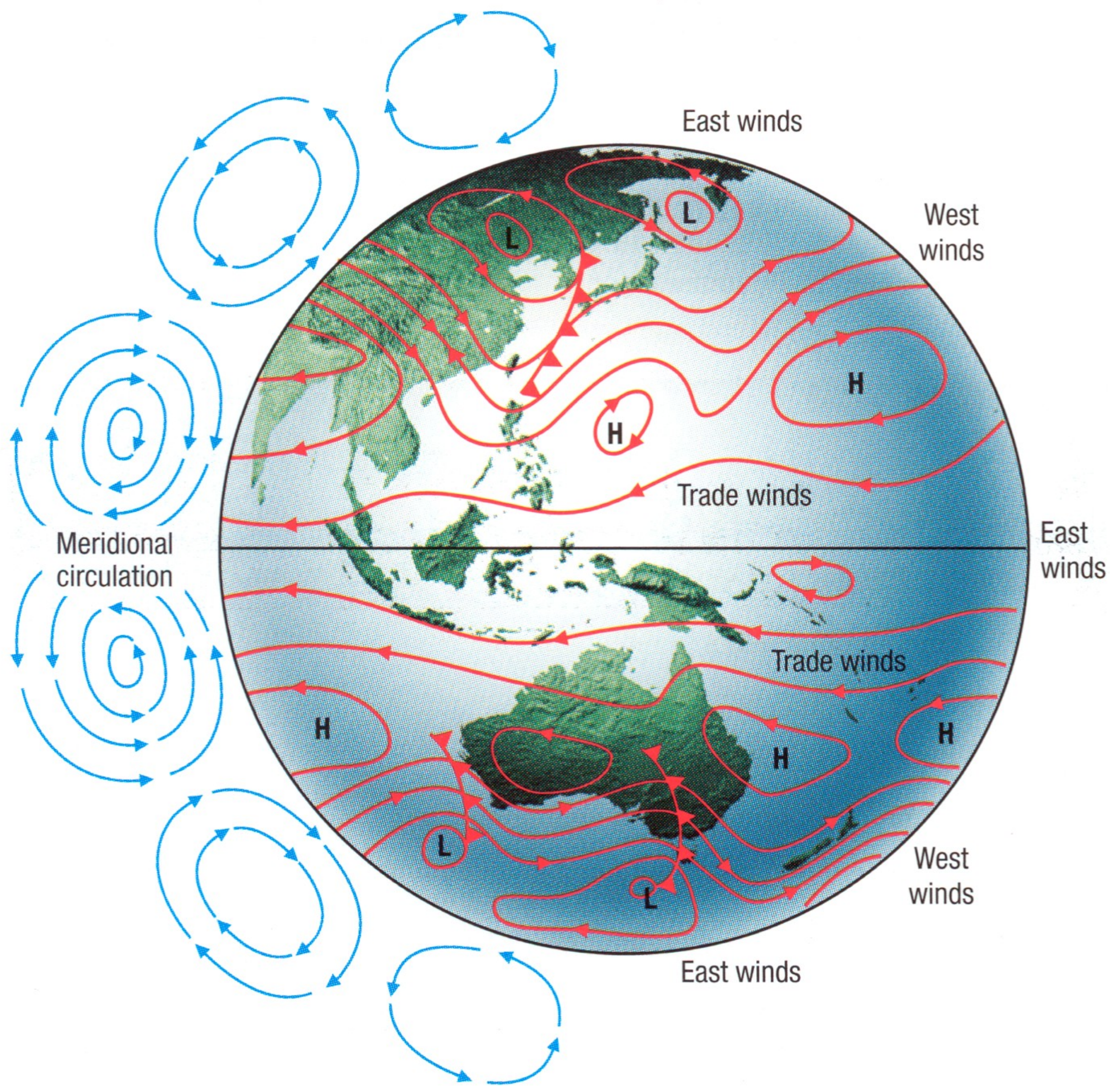
- General Circulation of the Atmosphere (Three-Cell Model)
  - **As the air aloft moves northward, it's volume is slowly reduced.**
    - **The volume is reduced because of the spherical surface.**
    - **As you move closer to the poles the latitude lines get closer together to reflect the reduced radius.**
  - **The slowly reducing volume creates a region of convergence near 30°.**
  - **The air becomes so compacted at 30° that it must sink.**
  - **The sinking air strikes the surface and spreads out in all directions.**
  - **This creates a belt of high pressure called subtropical highs.**

- General Circulation of the Atmosphere (Three-Cell Model)
  - **The air from the high pressure belt that is directed south curves to the right and eventually flows into the equatorial surface low (trade winds).**
  - **The air from the high pressure belt that is directed northwards is again deflected to the right (westerlies).**
  - **The subsiding air produces generally clear skies and warm surface temperatures. Here are where the major deserts of the world are found.**
  - **This region is know as the horse latitudes.**

- General Circulation of the Atmosphere (Three-Cell Model)
  - **The mild air from the high pressure belt that is called the westerlies moves poleward and encounters cold air moving down from the poles.**
  - **The cold air from the poles meets the air from the subtropical high in the horse latitudes.**
  - **This convergent zone causes the air to rise once again, and is called the polar front.**

- General Circulation of the Atmosphere (Three-Cell Model)




East winds
West winds
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Trade winds
East winds
Meridional circulation
Trade winds
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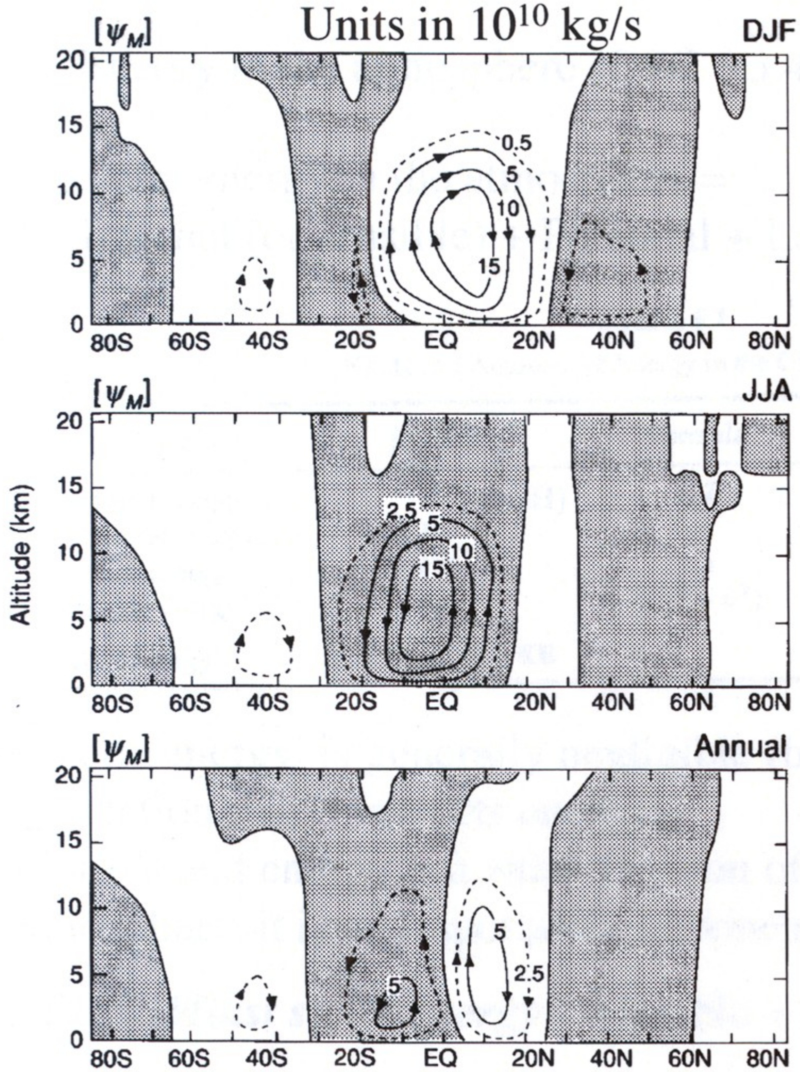
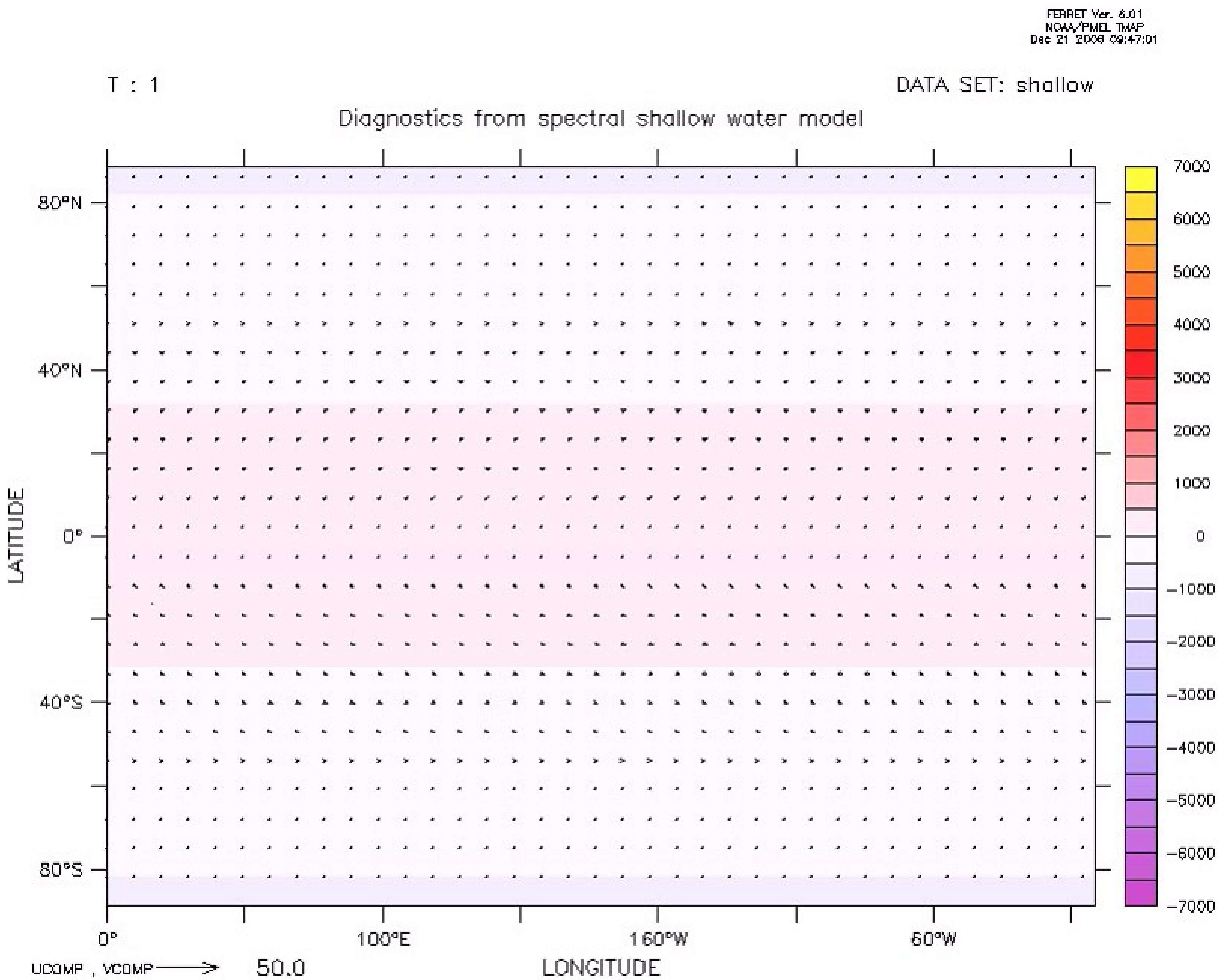
West winds
East winds



The mean meridional circulation (MMC)

The stronger central cell(s) are the ***Hadley cell(s)*** and they are ***thermally direct***

The weaker midlatitude cells are the ***Ferrel cells*** and are ***thermally indirect*** (rises in cold air, drops in warm air - so they actually transport energy from cold to warm regions)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DATA SET: shallow
Diagnostics from spectral shallow water model
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40°N
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160°W
60°W
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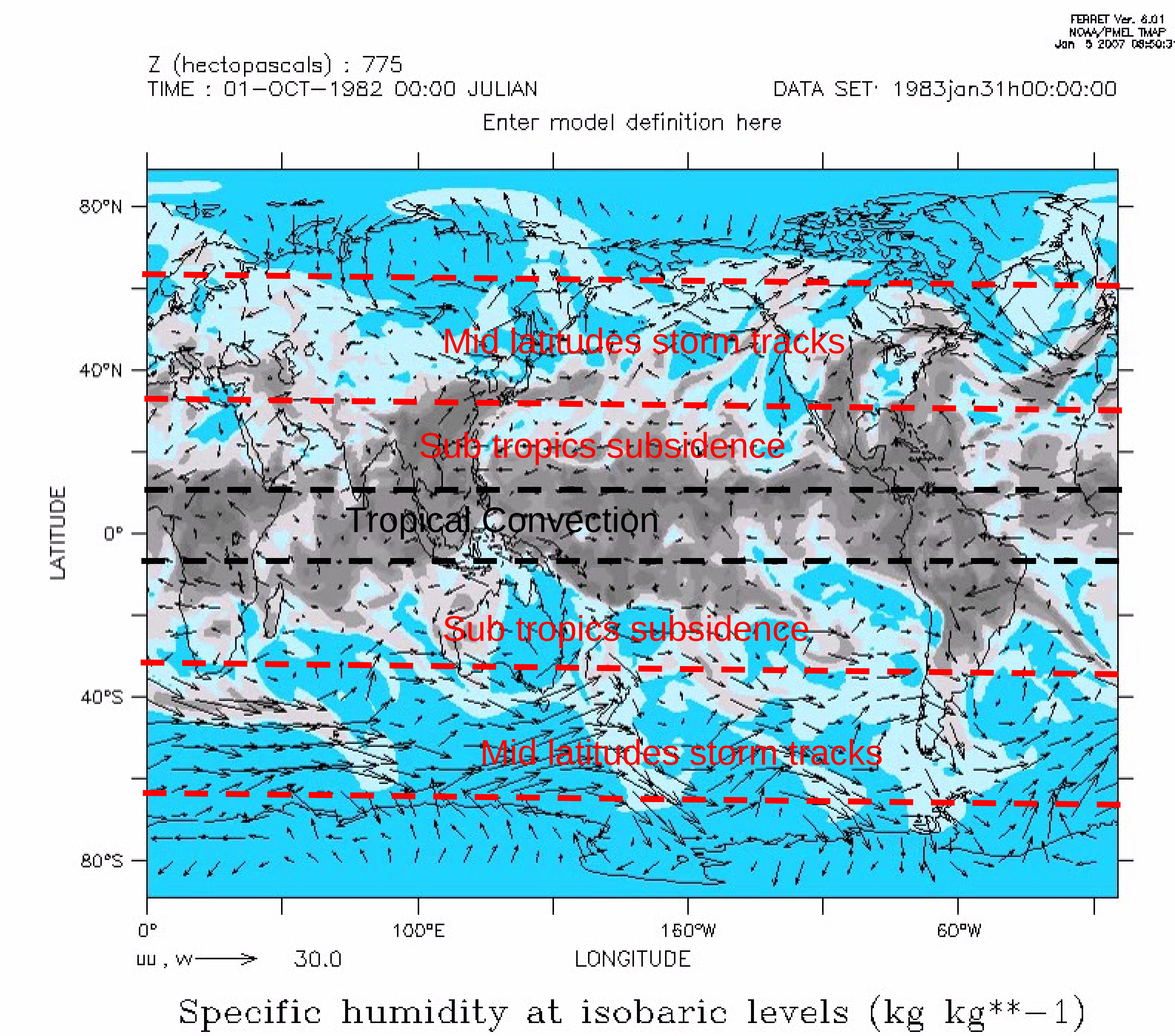
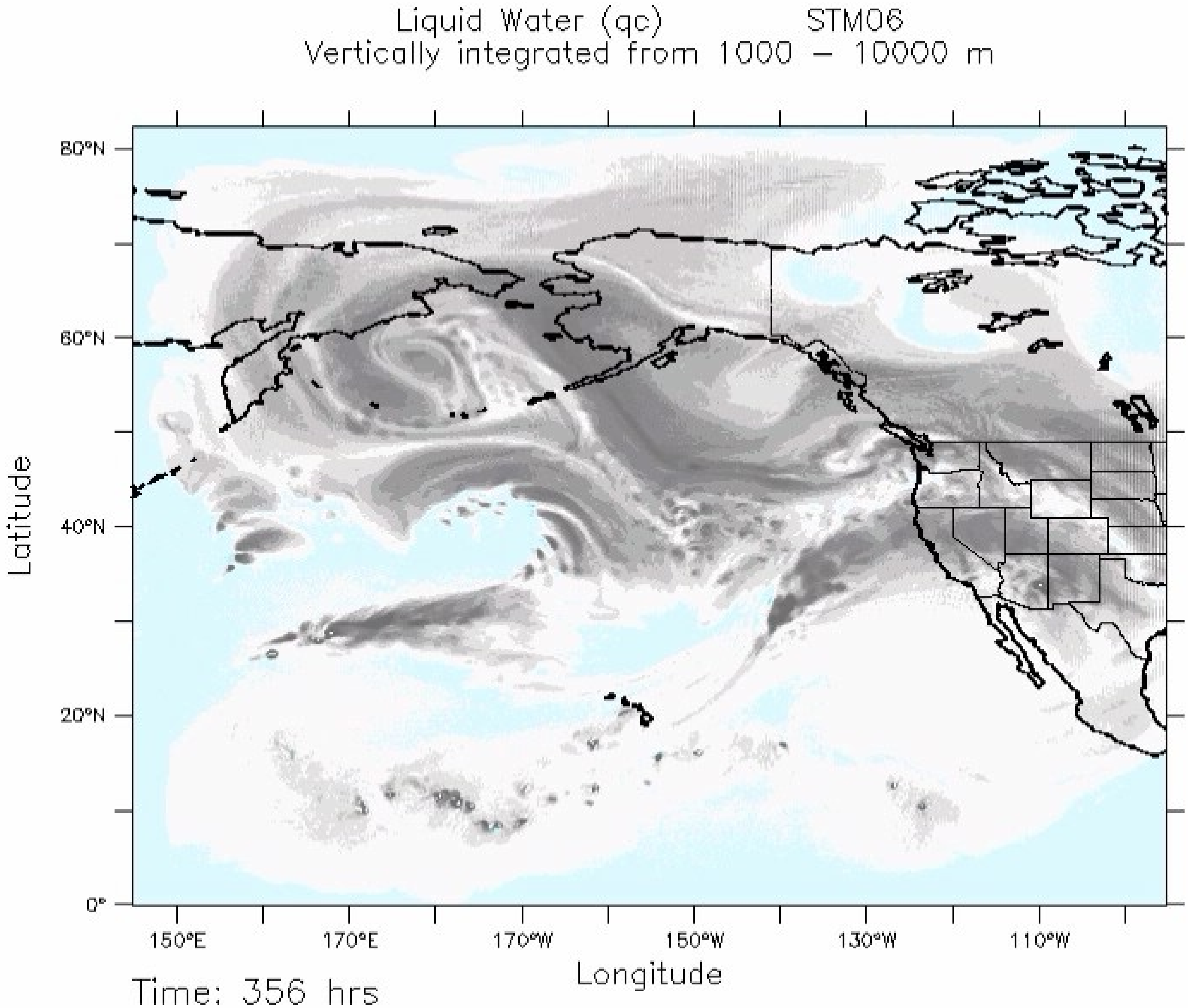
−6000
−7000

H−H[X=@AVE,Y=@AVE]


FERRET Ver. 6.01
NOAA/PMEL TMAP
Z (hectopascals) : 775
TIME : 01-OCT-1982 00:00 JULIAN
DATA SET: 1983jan31h00:00:00
Enter model definition here
Mid latitudes storm tracks
Sub tropics subsidence
Tropical Convection
Sub tropics subsidence
Mid latitudes storm tracks
80°N
40°N
0°
40°S
80°S
LATITUDE
0°
100°E
160°W
60°W
LONGITUDE
uu , w
30.0
Specific humidity at isobaric levels (kg kg**−1)


Liquid Water (qc) STM06
Vertically integrated from 1000 - 10000 m
80°N
60°N
40°N
20°N
0°
Latitude
150°E
170°E
170°W
150°W
130°W
110°W
Longitude
Time: 356 hrs

Things to remember from Lecture 1.

•The energy balance is not maintained for each latitude as there is a net gain of energy at the tropics and a net loss at the poles. as a result the poles are cool and the equator is warm.

• To restore balance, the atmosphere transports warm air poleward and cools air equatorward.

•We have seen before that, averaged over the entire earth, incoming solar radiation is roughly equal to the outgoing earth's radiation.

•The heat at the tropical regions produce convection, transport heat upward and spread the heat to the subtropics, producing a circulation called the Hadley cell.

•As the air moves to higher latitudes, since the earth has shorter radius to its axies, the air has an excess of eastward angular momentum. Producing an eastward jet.