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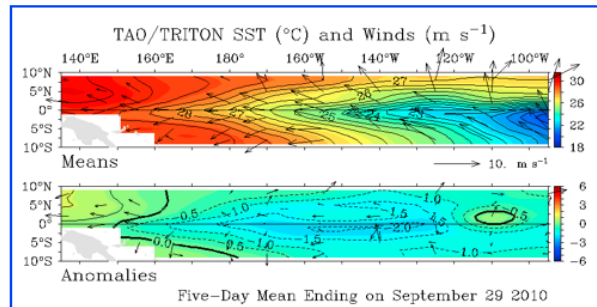
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ATM S 560/Ocean 560, Autumn Quarter 2010

Coupled Atmosphere - Ocean Interactions



Most recent TAO / TRITON data for the Tropical Pacific
(from <http://www.pmel.noaa.gov/tao/jsdisplay>)

Class Meeting Times and Location: Tuesday/Thursday from 1:30 to 2:45 pm in Room 310C in the [Atmospheric Sciences Building](#)

www.atmos.washington.edu/~david/ATM560_2010/

Atmosphere-Ocean Interactions

(ATM/OCN 560)

Topics (I think) we will be discussing:

- The Basic Physics of the Mean Climate (1.5 weeks)
 - Review the essential elements (physics and geometry) that are responsible for the gross features of the mean climate state of the tropical atmosphere and oceans.
 - Goal: to build an intuition for the processes responsible for variability in the climate system, from seasonal to decadal time scales.
- Uncoupled Atmosphere and Ocean Variability (three+ weeks)
 - overview of the dynamics of the uncoupled tropical oceans and atmosphere.

Atmosphere-Ocean Interaction

(ATM/OCN 560)

Topics (I think) we will be discussing (cont)

- Coupled Atmosphere-Ocean Variability in the Tropics (six weeks)
 - the dynamics of the El Nino/Southern Oscillation phenomenon and the Meridional Modes: two coupled phenomenon in the climate system
 - ENSO is the dominant pattern of natural climate variability on interannual time scales, and has been shown to have some impact on weather outside of the tropics.
 - The Meridional Modes are found in both the Atlantic and Pacific basins. They are the dominant forcing for variations in hurricane activity in the Atlantic and are the primary energy source for ENSO.

Atmosphere-Ocean Interaction

(ATM/OCN 560)

1. Atmosphere
2. Ocean Circulation and coupled feedbacks
3. Tropical Atmosphere Response to SST
4. Tropical Ocean Response to Winds
5. Observations of ENSO
6. ENSO Physics and Teleconnections
7. Low Frequency ENSO
8. ENSO in the mid-Holocene
9. Physics of Meridional Modes Coupled Modes
10. Meridional Mode and ENSO
11. Meridional Mode in the Atlantic

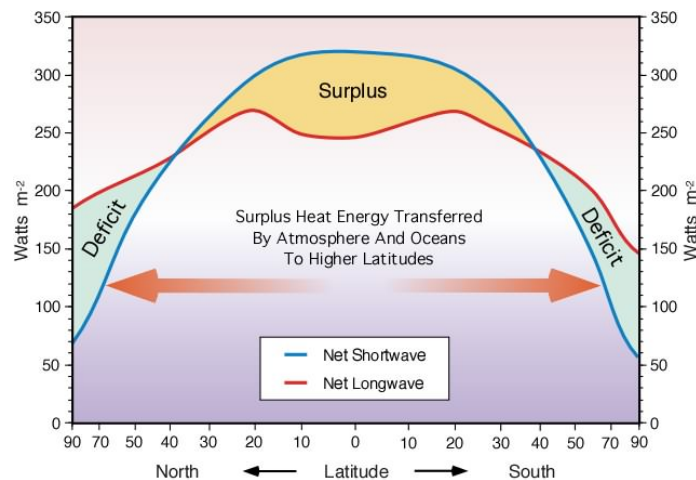
Basics of Tropical Climatology

- Hadley Circulation
 - Instigated by meridional gradients in absorbed solar radiation
 - But fundamentally shaped by (and governed by?) eddies
- East - West Asymmetries and the Walker Circulation
 - Due to coupled atmosphere-ocean processes
- North-South Asymmetries
 - Mainly to the effects of asymmetry in geometry (orography and land distribution) on the atmosphere

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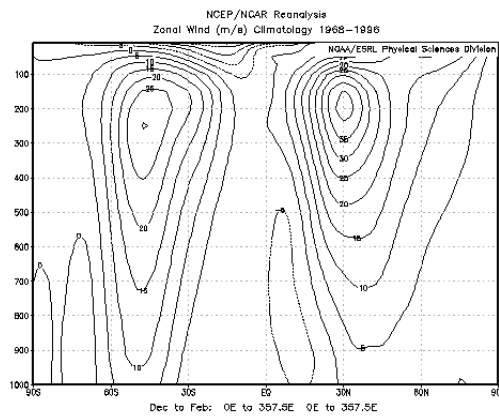
Annual Radiation Balance



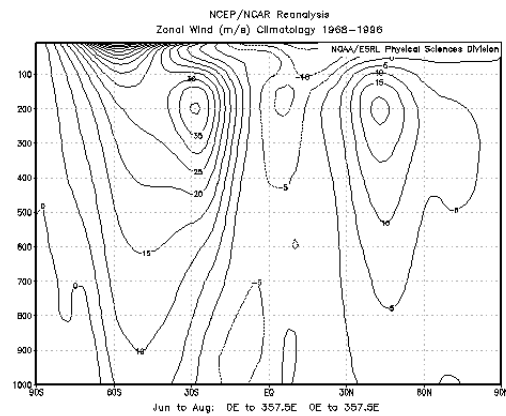
The meridional gradient in Absorbed Solar Radiation (ASR) drives circulation.

Climatological Zonal Averaged Zonal Wind

DJF



JJA



The Hadley Circulation

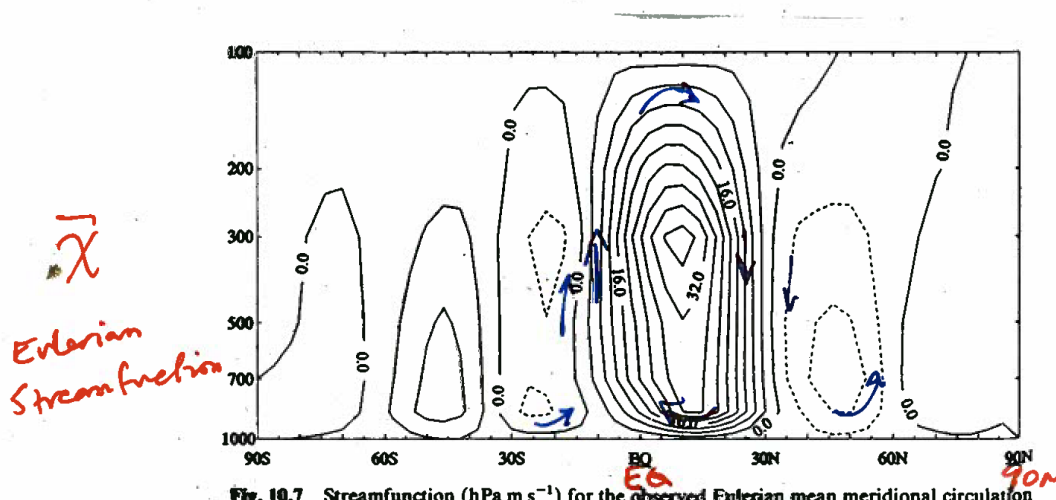
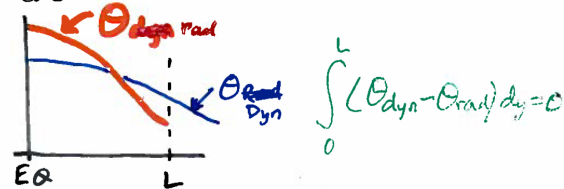


Fig. 10.7 Streamfunction (hPa m s^{-1}) for the observed Eulerian mean meridional circulation for Northern Hemisphere winter, based on the data of Schubert *et al.* (1990).

Held-Hou (1980) Heuristic Model of the Hadley Circulation

Conservation of Angular Momentum, Thermal Wind,
no energy gained in circuit



The HH model is a theory of the cell width L and strength H , and of the strength of the Jet J

	HH	Observed
Jet (m/s)	85	35
L (°lat)	25	30
H (hPa m s ⁻¹)	1.4	34

← i.e., no Hadley Cell or surface winds!

- What's missing? The large scale eddies are crucial for the mean atmospheric climate (the solution is barotropically neutral in cartesian coordinates, and unstable in spherical coordinates). No eddies - no Trade Winds!

THEORY FOR THE ZONAL MEAN CIRCULATION

Tools: Conservation of angular momentum &!

$$M = (\Omega a \cos \phi + u) a \cos \phi$$



∴ A parcel at rest on the equator ($\phi=0$) has eastward speed

$$u = \Omega a \sin^2 \phi / \cos \phi \approx \frac{\Omega y^2}{a} \quad (1.1)$$

at latitude $\phi \approx y/a$ (for $|\phi| < 1$)

ϕ	0°	20°	30°
$M = \text{const.}$	0 m/s	56	127
Observed (DJP)	~ 0	~ 20	~ 45



M is "not conserved" because of instabilities in the (horizontal) shear of the flow

Under this caveat empty, we introduce the heuristic model of the Hadley Circulation

Held + Hou model: 'angular momentum conserving', radiative equilibrium, conservation of energy, plus thermal wind.

[presentation based on James, 1994]

Thermodynamic Equilibrium: $\Theta \equiv \Theta_E$

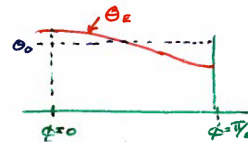
without circulation, the atmosphere would quickly obtain a radiative equilibrium balance with an equator-to-pole temperature difference $\Delta\Theta$.

Held and Hou (HH) assumed

$$\Theta_E = \Theta_0 - \frac{2}{3}(\Delta\Theta)P_2 \cos^2 \phi,$$

↑
global mean
radiative temperature

where $P_2 = (g \sin^2 \phi - 1)/2$



For solutions narrow compared to the Earth's radius,

$$\Theta_E = \Theta_{E0} - \frac{\Delta\Theta}{a^2} y^2 \quad (1.2)$$

But differential solar heating drives a circulation,

so $\Theta \neq \Theta_E$, in general.

HH assumed that diabatic processes act to restore Θ to Θ_E in a linear fashion on a time scale τ_E . Hence,

$$\frac{D\Theta}{Dt} = \frac{\Theta_E - \Theta}{\tau_E} \quad (1.3)$$

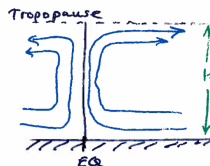
Assume poleward moving air leaving the equator with $u=0$ conserves angular momentum.

Then

$$u \equiv u_M = \frac{\Omega y^2}{a} \quad (1.1)$$

The lower level wind is taken to be small by friction so that

$$\frac{\partial u}{\partial z} \sim \frac{u_M}{H} = \frac{\Omega y^2}{Ha} \quad (1.4)$$



Finally, thermal wind (good at $\omega \rightarrow 0$) states

$$-\frac{g}{\Theta} \frac{\partial \Theta}{\partial y} = f \frac{\partial u}{\partial z} \approx \frac{2\Omega y}{a} \frac{\partial u_M}{\partial z} \quad (1.5)$$

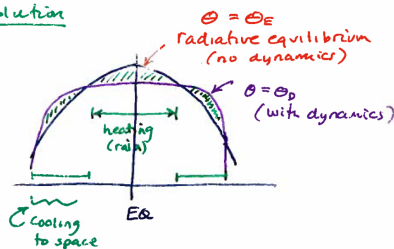
Plugging (1.5) into (1.4) and integrating yields the actual equilibrium temperature distribution with Θ_p under circulation: (dynamics)

$$\Theta = \Theta_D = \Theta_{D0} - \frac{\Omega^2 \Theta_0}{2a^2 g H} y^4 \quad (1.6)$$

unknowns:

- (1) the width of the cell, $y = L$;
- (2) the potential temp at the equator, Θ_{D0} .

Expected Solution



Constraints

- (1) The width of the cell is where $\Theta_D = \Theta_E$ (1.7)
(poleward of L , the solution requires heating again $\Rightarrow$ not physical).

- (2) Require that a parcel going around cell gets no net heat. Hence,

$$\int_{-L}^L (\Theta_E - \Theta_D) dy = 0 \quad (1.8)$$

Solving (1.8), (1.6-1.8) yields the solution for the width of the cell L :

$$L = \left(\frac{5gH\Delta\theta}{3\alpha^2\theta_e} \right)^{1/2} \quad (1.9)$$

Typical Numbers

Equator - Pole radiative temp. diff $\Delta\theta \approx 60^\circ\text{K}$

$$\theta_e = 255^\circ\text{K}; \quad H = 15\text{ km}$$

$$\Rightarrow L = 24^\circ \text{ lat.} \quad (\text{ok})$$

$\therefore$ Dynamics acts to cool at the equator compared to the radiative temperature by

$$\delta \equiv \theta_{e0} - \theta_{p0} = -\left(\frac{L}{a}\right)^2 \frac{\Delta\theta}{6} \approx 1.2^\circ\text{K}$$

Summary of Held + Hou's Hadley cell

$\rightarrow$ circulation due to insolation gradient

$\rightarrow$ conserves energy and angular momentum

yields width of cell L

$$L \sim (\Delta\theta)^{1/2}, \Omega^{-1}, H)$$

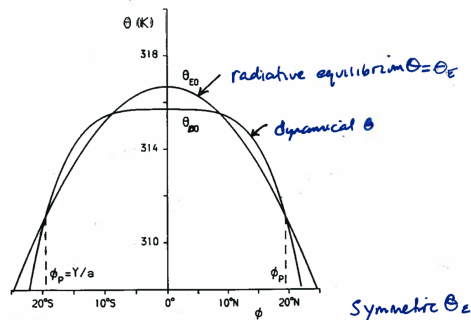


Fig. 4.5. Showing θ_e and θ_p as a function of poleward distance for the Held and Hou model. θ_p must be chosen so that the areas between the two curves are equal, i.e., so that there is no net heating of air parcels.

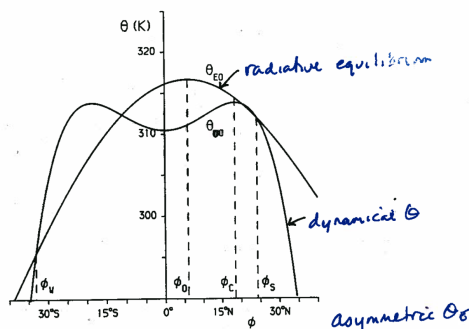
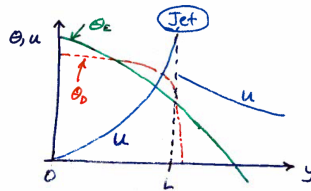


Fig. 4.6. The Held-Hou model for the case of a heating maximum away from the equator. The latitudes ϕ_U , ϕ_S , and ϕ_C , as well as the equatorial temperature θ_{e0} are to be determined.

Notes

- (1) for $|y| < L \Rightarrow u = \frac{2y^2}{L}$
 for $|y| > L \Rightarrow$ thermal wind and $\theta = \theta_E$



Jet @ edge of cell.

- (2) Solution is barotropically neutral
 $f(f - \frac{\partial u}{\partial y}) = 0$,
 but convergence of meridians (sphere) ensures
 flow instability at jet location $\rightarrow$ eddies will
 affect solution here.

(3) Strength of Cell

In deep tropics, diabatic heating is balanced
 by vertical motions (adiabatic cooling). At equator

$$\frac{\theta_E - \theta_0}{\tau} = \frac{D\theta}{Dt} \approx w \frac{\partial \theta}{\partial z} = w N^2 \theta_0 / g$$

numbers: $\theta = 1.2^\circ$; $N^2 = 10^3 s^{-2}$; $\tau = 15 \text{ days} \Rightarrow W = 4 \times 10^{-4} \text{ m/s}$
 (observed $\sim 10^{-2} \text{ m/s}$). Hence, the HH model gives wrong intensity cell

Maintenance of observed Hadley Cell and Zonal Mean circulation

planetary waves (eddies) affect the zonal mean
 flow in important (qualitative) ways!

THE EULERIAN MEAN EQUATIONS

In log-P coordinates ($z = H \ln P_0/P$; $H = RT_0/g$)

on the midlatitude β -plane the eqns are (see
 Holton (1992 ch. 10.2); extension to spherical domain
 straightforward):

$$\frac{D}{Dt} \underline{u} - f_0 \underline{k} \times \underline{u} + \nabla \phi = \underline{F} \quad (1.10)$$

horizontal momentum

$$\frac{\partial \phi}{\partial z} = \frac{R}{H} T \quad (1.11)$$

hydrostatic balance + ideal gas

$$\underline{V} \cdot \underline{u} + \frac{1}{P_0} \frac{\partial}{\partial z} (P_0 w) = 0 \quad (1.12)$$

continuity, compressible

$$\text{ad } \frac{DT}{Dt} + \frac{N^2 H}{R} w = \frac{J}{C_p} \quad (1.13)$$

thermodynamic

where

$$\frac{D}{Dt} \equiv \frac{\partial}{\partial t} + u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y} + w \frac{\partial}{\partial z} \quad (1.14)$$

$$\underline{V} \equiv \frac{\partial}{\partial x} \hat{i} + \frac{\partial}{\partial y} \hat{j} \quad N^2 = \frac{R}{H} \left(\frac{RT_0}{C_p H} + \frac{dT_0}{dz} \right) \quad (1.15)$$

P_0 = reference density = $P_0(z)$; $\underline{F}$ = friction; J = diabatic heating.

Write variables in terms of a zonal mean component ($\bar{A}$), where

$$\bar{A}(y, z, t) \equiv \frac{1}{L} \int_0^L A dx \quad (1.18)$$

where L is the distance around a latitude circle, and a perturbation component (A') such that

$$A \equiv \bar{A} + A'$$

The average (1.16) is taken at a fixed place $\rightarrow$ Eulerian $\downarrow$

Under the zonal mean

$$\bar{v}_0 \equiv \frac{\partial \bar{\phi}}{\partial x} f_0^{-1} \equiv 0,$$

so the zonal mean meridional flow is ageostrophic.

Hence, to a good approximation

$$|\bar{v}| \ll |\bar{u}|,$$

where to zeroth order

$$f_0 \bar{u} = -\frac{\partial \bar{\phi}}{\partial y} \quad (1.17)$$

Combined with the hydrostatic equation (1.12) yields the

zonal mean thermal wind relationship

$$f = \frac{\partial \bar{u}}{\partial z} + \frac{R}{H} \frac{\partial \bar{T}}{\partial y} = 0 \quad (1.18)$$

The zonal mean of the first two equations (zonal momentum (1.10) and thermodynamics (1.13)) yields

$$\frac{\partial \bar{u}}{\partial t} - f_0 \bar{v} = -\frac{\partial}{\partial y} (\overline{u'v'}) + \bar{X} \quad (1.19)$$

$$\frac{\partial \bar{T}}{\partial t} + \frac{N^2 H}{R} \bar{w} = -\frac{\partial}{\partial y} (\overline{v'T'}) + \bar{J}/c_p \quad (1.20)$$

where $\bar{X}$ is the small scale (unresolved) zonal mean drag (e.g., boundary layer, gravity waves).

In (1.19) and (1.20) we have used the observational facts that indicate:

- (i) advection by the mean meridional circulation is small compared to other terms;
- (ii) The vertical eddy flux terms $\overline{u'v'}$ and $\overline{w'e'}$ are small compared to other terms
- (iii) small scale ad radiative effects ($\bar{J}$) are in $\bar{X}$ and $\bar{J}$.

Notes

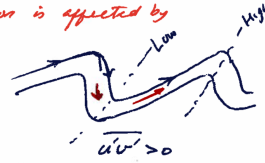
- (1) The zonal mean circulation & $\bar{u}$ is affected by zonal mean friction and zonal mean heating (via thermal wind)
- (2) Forcing by eddy heat or momentum flux causes changes in both the mean flow and temperature (by 1.18)

(3) The zonal mean circulation is affected by



heat flux convergence
(midlatitudes)

e.g., Synoptic storms



momentum flux convergence
(subtropics and midlatitudes)

↑ Hadley impact!

e.g., Jet wobbles

(4) Steady Motions (adiabatic, inviscid)

meridional circulation = fract (eddies)

Coriolis Torque = divergence of $\overline{u'v'}$

Adiabatic Cooling = divergence of $\overline{v'T'}$

(5) Eddies are crucial for midlatitude troposphere

Eddy momentum divergence provides a
brake for the Hadley Cell Jet
and

Is fundamental for setting the climatological
mean trade winds in the tropics

It is the dominant player for
setting strength of the Hadley Cell.