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.
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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
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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
Basics of Tropical Climatology

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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

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

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

<table>
<thead>
<tr>
<th></th>
<th>HH</th>
<th>Observed</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jet ($m/s$)</td>
<td>85</td>
<td>85</td>
</tr>
<tr>
<td>$L$ ($^\circ$lat)</td>
<td>25</td>
<td>30</td>
</tr>
<tr>
<td>$H$ ($hPa$ms$^{-1}$)</td>
<td>1.4</td>
<td>3.4</td>
</tr>
</tbody>
</table>

- 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:

\[ M = (\beta_0 \cos \phi + \nu) \alpha \cos \phi \]

1. A parcel at rest on the equator (\$\beta_0\$) has no angular speed

\[ \nu = \beta_0 \sin \phi \cos \phi = \frac{\beta_0 \phi}{\alpha} \]  

at latitude \( \phi = 90^\circ \) (for \( \alpha > 0 \)).


<table>
<thead>
<tr>
<th>( \phi )</th>
<th>0°</th>
<th>20°</th>
<th>30°</th>
</tr>
</thead>
<tbody>
<tr>
<td>M (cont.)</td>
<td>0.95</td>
<td>0.65</td>
<td>0.45</td>
</tr>
<tr>
<td>Observed (DPS)</td>
<td>-0.1</td>
<td>-0.07</td>
<td>-0.05</td>
</tr>
</tbody>
</table>

M is not conserved because of instability in the (horizontal) shear of the flow.

Under this simplest example, we introduce the heuristic model of the Hadley Circulation.

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

[presentation based on James, 1976]

Thermodynamic Equilibrium: \( \Theta = \Theta_e \)

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

Held and Hou (1974) assumed

\[ \Theta_e = \Theta_0 - \frac{1}{2} (\Delta \Theta) P_e (\sin \phi), \]

where \( P_e = (\alpha \sin \phi - 1)/\alpha \)

For solution narrow compared to the Earth's radius,

\[ \Theta_e = \Theta_0 - \frac{\Delta \Theta}{\alpha^2} \phi \]  

But differential solar heating drives a circulation, so \( \Theta \neq \Theta_0 \), in general.
HT assumed that adiabatic processes act to restore \( \Theta \) to \( \Theta_0 \) in a linear fashion on a time scale \( T_E \). Hence,

\[
\frac{D\Theta}{D\tau} = \frac{\Theta - \Theta_0}{T_E}
\]  

(1.2)

Assume poleward moving air leaving the equator with \( u = 0 \) conserves angular momentum. Then

\[
u = u_m = \frac{\Omega y}{a}
\]  

(1.3)

The lower level wind is then to be small by writing so that

\[
\frac{\partial u}{\partial \tau} = \frac{\partial u_m}{\partial \tau} = \frac{\Omega y}{Ha} \frac{\partial y}{\partial \tau}
\]  

(1.4)

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

\[
-\frac{\partial \Theta}{\partial z} = \frac{\Omega y}{a} \times \frac{\partial (\rho y) \Theta_m}{\partial \tau}
\]  

(1.5)

Plotting (1.5) and integrating yields the actual equilibrium temperature distribution \( \Theta_z \) under circulation (dynamics)

\[
\Theta = \Theta_0 = \Theta_0 + \frac{\Lambda y}{2 \Omega a} \Theta_0 y^2
\]  

(1.6)

Unknowns:

1. The width \( y \) of the cell, \( y = L \);  
2. The potential temp at the equator, \( \Theta_0 \).

Expected Solution

Constraints

1. The width \( y \) of the cell is where \( \Theta_0 = \Theta_0 \) \ (poleward of L, the solution requires heating again \( \Rightarrow \) not physical).

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

\[
\int_{0}^{L} (\Theta_z - \Theta_0) \, dy = 0
\]  

(1.8)

Solving (1.8), (1.6-1.8) yields the solution for the width \( y \) of the cell \( L \).
Equator - Pole radiation temp. diff. \( \Delta \theta \approx 60^\circ K \\
\theta_0 = 255^\circ K; \quad \theta = 15^\circ K \\
\Rightarrow \theta = 245^\circ K \quad (ok)

Dynamics acts to cool at the equator compared to the radiative temperature by

\[ \theta = \theta_0 - \theta = -\frac{L^2}{a} \frac{\Delta \theta}{a} \approx 1.2^\circ K \]

Summary of Held-Hou-Hasten cell
- Circulation due to viscosity gradient
- Conserve energy and angular momentum
- Yields width of cell \( L \)

\[ L \sim (\Delta \theta, x, y) \]

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**Fig. 4.5.** Showing \( \theta_1 \) and \( \theta_2 \) as a function of poleward distance for the Held and Hou model. \( \theta_1 \) 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 \( \theta_1 \), \( \theta_2 \), and \( \theta_3 \) as well as the equatorial temperature \( \theta_0 \) are to be determined.
Notes

(1) for \( |y| < L \) ⇒ \( u = \frac{2y}{L} \)
   for \( |y| > L \) ⇒ formal wid ed at \( \theta = \theta_0 \)

Jet at edge of cell.

(2) Solution is baroclinically neutral

\[ f(\frac{\partial \theta}{\partial y}) = 0, \]

but convergence of meridional (spheric) terms

flow instability at jet location ⇒ eddies will

affect solution here.

(3) Strength of Cell

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

\[ \frac{dE}{dz} = \frac{dW}{dz} = \frac{dW}{dz} (\text{adiabatic}) \]

numbers:

\[ \theta = 1.25°; \quad N_{\text{Eq}} = 10^{-3}; \quad D = 1 \text{ day} \quad \Rightarrow \quad W = 4 \times 10^{-4} \text{ m s}^{-1} \]

(used = \( 10^{-3} \text{ m s}^{-1} \)). Hence, the NH model gives strong 4-weekly cell.

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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 (\( z = \frac{H}{P} \)) \( \beta \)-plane Taylor or (see

Hoke (1992 ch. 10) extension to spherical domain

straightforward):

\[ \frac{D\theta}{Dz} - f \frac{\partial \theta}{\partial y} + \frac{V \theta}{c_p} = \frac{J}{M} \quad (1.10) \]

\[ \frac{\partial \theta}{\partial y} = \frac{\phi}{H} \quad (1.11) \]

hydrostatic balance + ideal gas

\[ V \cdot \vec{u} + \frac{1}{\rho_0} \frac{\partial (\rho \vec{u})}{\partial t} = 0 \quad (1.12) \]

continuity, compressible

\[ \frac{\partial}{\partial t} + \frac{V \cdot \vec{u}}{c_p} = \frac{\partial}{\partial z} \quad (1.13) \]

thermodynamics

where

\[ \frac{D}{Dz} \approx \frac{\partial}{\partial z} + \frac{V}{c_p} \frac{\partial}{\partial y} \quad \frac{\partial}{\partial x} \]

\[ \rho_0 = \rho_0(z) \quad f = f_0 \cos \omega \]
Write variables in terms of a zonal mean component \( \overline{A} \), where
\[
\overline{A}(y, z, t) = \frac{1}{L} \int_{0}^{L} A(x) \, dx \quad (1.18)
\]
where \( L \) is the distance around a latitude circle, and a perturbation component \( A' \) such that
\[
A = \overline{A} + A'.
\]

The average \((1.18)\) is taken at a fixed place \( \xi \), and \( \xi \) tends to the zonal mean
\[
\overline{\nu}_x = \frac{\partial \overline{\nu}}{\partial x} = 0,
\]
so the zonal mean meridional flow is geostrophic.

Hence, to a good approximation
\[
|\overline{\nu}| << |\nu|,
\]
where to good order
\[
f \overline{\nu} = -\frac{2 \overline{\nu}}{\partial y}
\]
Combined with the hydrostatic equation \((1.19)\) yields
\[
\text{Zonal mean zonal wind relationship}
\]
\[
-f \frac{\overline{\nu}}{\partial z} + \frac{\partial \overline{\psi}}{\partial y} = 0 \quad (1.18)
\]

The zonal mean of the final two equations
(zonal momentum \((1.18)\) and geostrophic \((1.19)\) yields)
\[
\frac{\partial \overline{\nu}}{\partial t} + u \frac{\partial \overline{\nu}}{\partial x} = -\frac{2 \overline{\nu}}{\partial y} + \overline{F} \quad (1.19)
\]
\[
\frac{\partial \overline{\psi}}{\partial t} + K + \frac{\partial \overline{\psi}}{\partial y} = -\frac{2 \overline{\nu}}{\partial y} + \frac{\partial \overline{F}}{\partial x} \quad (1.20)
\]
where \( \overline{F} \) is the small-scale (unresolved) zonal mean forcing \( \text{(e.g., boundary layer, gravity waves)} \).

In \((1.19)\) and \((1.20)\) we have used the observational facts that indicate:
1. advection by the mean meridional circulation is small compared to other terms;
2. the vertical eddy flux terms \( -\frac{2 \overline{\nu}}{\partial y} \) are small compared to other terms;
3. small scale air motion affects \( \overline{F} \), e.g. \( \overline{F} = \overline{F} \).

Notes:
1. Our good mean correlation \( \overline{\nu} \) is affected by zonal mean forcing and zonal mean heating (via thermal wind).
2. Forcing by eddy heat and momentum flux cause changes in both the mean flow and temperature. (1.20)
(3) The zonal mean circulation is affected by

- Heat flux convergence (midlatitudes)
- Momentum flux convergence (subpolar and midlatitudes)

E.g., Synoptic storms

(4) Steady Motion

- Adiabatic, inviscid
- Meridional circulation = forced (eddies)

Coriolis torque = divergence of \( \overrightarrow{v} \)

Adiabatic cooling = divergence of \( \overrightarrow{v} \)

(5) Eddies are crucial for midlatitude variability

Eddy momentum divergence provides a brake for the Hadley cell jet

It is fundamental for setting the climatological mean brake inside the tropics

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