Lecture 3 Topics

- El Niño/Southern Oscillation (ENSO) tropical teleconnections
- Weak temperature gradient approximation
- Quasi-equilibrium Tropical Circulation Models (QTCMs)
- Convective margins
Current Conditions

Observed Sea Surface Temperature (°C)

Observed Sea Surface Temperature Anomalies (°C)

OLR ANOMS Pentad Centered on 23 Jan 2016

7-day Average Centered on 20 January 2016

Latimes.com 08/25/15

Climate Prediction Center
Blame El Niño [or La Niña]!

El Niño-Driven Storms Bring High Winds to California, Heavy Snow to Arizona

Did El Nino Make This Weekend’s Colossal Snowstorm Worse?

Wall Street Journal 01/07/16
Gizmodo.com 01/21/16
ENSO fundamentally involves intensification (and spatial redistribution of rainfall) in the Pacific: *What are the implications of this for climate elsewhere in the Tropics?*
Who’s driving the Pacific?

1. The ocean is the driver: The tropical Pacific has a pronounced zonal asymmetry in SST, with the warmest water in the west and coldest water in the east. Intense deep convection is favored over the warmest water in the west. With the intense deep convection in the west, there is associated ascent, low-level convergence, and low surface pressure; in the east, there is descent, low-level divergence, and high surface pressure. The resultant surface pressure gradient is associated with easterly flow along the equator.

2. The atmosphere is the driver: The easterly wind stress applied to the ocean surface leads to a shoaling of the thermocline and upwelling in the eastern Pacific. These conditions are associated with cool SST in the eastern Pacific. By contrast, the west has a deep thermocline and experiences little upwelling, leading to warm SST in the west. [Aside: we do need to consider differences in the ocean dynamics and currents along the eastern and western margins of Pacific, associated with how the applied surface stress is ultimately balanced in the vorticity.]

It is clear, in fact, that both the ocean and atmosphere in the equatorial Pacific are driving each other. The coupling between the ocean and atmosphere as it pertains to the large-scale features of Pacific climate is known as the Bjerknes feedback.
ENSO teleconnections schematic

**El Niño Phase ↔ “Warm SSTs”**

- **WARM EPISODE RELATIONSHIPS**
  - December - February

**La Niña Phase ↔ “Cool SSTs”**

- **COLD EPISODE RELATIONSHIPS**
  - December - February

**WARM EPISODE RELATIONSHIPS**

- June - August

**COLD EPISODE RELATIONSHIPS**

- June - August

Climate Prediction Center

NCEP
Recent forecasts
Observed El Niño Ts and Prec Anomalies

Throughout the remote tropics, the land surface warms by ~1-3K, while the ocean surface warms by ~1K.

Precipitation reductions (regionally up to 2 mm/day) are widespread throughout the remote tropics, but with some exceptions (e.g. western Indian Ocean).

*Chiang & Lintner [2005]*
There is a pronounced tendency in simulations as well as observations for tropical Pacific- and tropical remote-mean precipitation anomalies to anticorrelate.

More rainfall over the Pacific during El Niño⇔less rainfall elsewhere in the tropics

How do we understand this large-scale antiphasing?
The Gill (1980) Model

The Gill model is a dynamical framework that reduces the horizontal tropical circulation to a single vertical baroclinic mode (damped linear shallow water equations).

For the ENSO teleconnection problem, we consider how an imposed diabatic (convective) heating anomaly zonally localized along the equator affects the dynamics: primarily, it produces ascent colocated with the imposed heating and subsidence elsewhere.

http://www.asp.ucar.edu/colloquium/2000/Lectures/branstator.html
An alternative view: the tropospheric temperature mechanism

The Gill Model (1980) is essentially a dynamical construct. But the teleconnection problem also involves thermodynamics.

TT across the tropics is observed to be well correlated with NINO3 region sea SST anomalies (green box), particularly slightly after the peak of SST anomalies in NINO3 (e.g. bottom left and top right).

Chiang & Sobel [2002] recast the problem in terms of tropospheric temperature: the basic idea is that anomalous convection in the Pacific during El Niño sets tropical temperatures (in the free troposphere), with wave dynamics propagating that signal around the tropics.

*Chiang & Sobel [2002]*
1. Increased convection over the equatorial Pacific during El Niño heats the tropical troposphere.

2. The warming is rapidly propagated to the rest of the tropical free troposphere by equatorial planetary waves.

3. Over the remote tropics, tropospheric warming induces ABL moist static energy to increase, mediated by moist convective processes.

4. Remote tropical surface temperature increases as a consequence of increased ABL moist static energy. Over the remote oceans, latent heat flux acting through ABL moisture variations regulates the surface warming.
Some basic considerations

In the mid-latitudes, quasi-geostrophic advection dominates, leading to baroclinic (and barotropic) conversion processes. Thus, while diabatic heating is obviously important, it is typically not dominant in the time-evolution of the atmosphere on daily timescales.

In the tropics, baroclinic instability is typically not important (since horizontal temperature gradients are weak—more on this in the next slide). Barotropic instability is important in the formation of tropical waves and cyclones.

Within ~5° of the equator (the “deep tropics”), geostrophic balance does not hold. Other (more complicated) types of balance, e.g., nonlinear, apply.
Weak temperature gradient (WTG)

A consequence of the small Coriolis parameter near the equator as well as efficient equatorial wave dynamics is that horizontal temperature gradients in the tropical free troposphere are small. The weakness of tropical temperature gradients provides constraints on tropical large-scale dynamics and their relationship to diabatic processes.

We can take advantage of this to build a balanced dynamical framework for the tropical circulation (or at least the divergent portion thereof). This framework is known as the weak temperature gradient (WTG) approximation.

Recall from Helmholtz’s theorem, or the fundamental theorem of vector calculus, that we can decompose an arbitrary vector $\vec{F}$ as:

$$\vec{F} = -\nabla \Phi + \nabla \times \vec{A}$$

where $\Phi$ and $\vec{A}$ are scalar and vector potentials, respectively.

In particular, we can use this result to write the velocity field in terms of rotational (nondivergent) and divergent (irrotational) components:

$$\vec{v} = \vec{v}_{rot} + \vec{v}_{div}$$

Professor Adam’s Sobel Lecture at the 2005 GCC summer school.
Thermodynamic equations

Starting from thermodynamic equations for energy and moisture:

\[
\frac{\partial s}{\partial t} + \mathbf{u} \cdot \nabla s + \omega \frac{\partial s}{\partial p} = Q_c + Q_R \\
\frac{\partial q}{\partial t} + \mathbf{u} \cdot \nabla q + \omega \frac{\partial q}{\partial p} = Q_q
\]  
(1)  
(2)

In (1), \(Q_c\) is the convective heating and \(Q_R\) is the radiative heating, and in (2), \(Q_q\) is the convective drying. Note that integrating the convective heating and drying over the depth of the troposphere, i.e., from the surface to the tropopause, yields:

\[
\int_{p_s}^{p_t} Q_c \, dp = P + (\omega \, s)^{p_t}_{p_s} = P + H \\
\int_{p_s}^{p_t} Q_q \, dp = -P + (\omega \, q)^{p_t}_{p_s} = -P + E
\]
Thermodynamic balance for tropical precipitating conditions

\[
\frac{\partial s}{\partial t} = -u \cdot \nabla s - \omega \frac{\partial s}{\partial p} + Q_c + Q_R
\]

\[
\frac{\partial q}{\partial t} = -u \cdot \nabla q - \omega \frac{\partial q}{\partial p} + Q_q
\]

In the $s$ equation, the dominant balance is typically:

\[
\omega \frac{\partial s}{\partial p} = Q_c + Q_R
\]

For tropical deep convection regions, $Q_c \gg Q_R$, and assuming $\frac{\partial s}{\partial p} \sim \text{constant}$, so $Q_c \propto \omega$. Since $P \propto \int Q_c dp$, specifying the vertical velocity profile effectively determines the rate of precipitation. The “conventional” view is that vertical velocity is known \textit{a priori}, or at least determined by the momentum equations.
WTG equation and perspective

On the other hand, suppose we regard $\omega$ as completely determined by the diabatic heating. Then the s equation reduces to

$$\omega \frac{\partial s}{\partial p} = Q_c + Q_R$$

The WTG equation is diagnostic for $\omega$, rather than prognostic for dry static energy (or temperature).

Although the time evolution of temperature cannot be predicted from this equation, it may be reasonable to use for many purposes under which temperature does not vary much. On the other hand, we can immediately relate the 3D divergent circulation [consider mass continuity] to the diabatic heating. We can solve Poisson’s equation (using $\omega$) to obtain a WTG-consistent horizontal divergent wind field.
Relating to Lecture 2...

\[
Q_c \sim (w - w_c)^\beta
\]

\[
Q_c \sim \frac{(T - T_{Neutral})}{\tau_{adj}}
\]

**RCE:**

\[
Q_c(T_{RCE}) = -Q_R(T_{RCE})
\]

\[
\omega \frac{\partial s}{\partial p} = Q_c + Q_R
\]

**CQE:**

\[
\omega \frac{\partial s}{\partial p} - Q_R = Q_c \Rightarrow \left( \frac{dA_\lambda}{dt} \right)_{LS} = - \left( \frac{dA_\lambda}{dt} \right)_C
\]
WTG and the ENSO tropical teleconnection

Approach: Force QTCM1 with prescribed 1997-1998 SSTs over the Pacific domain, coupled to a “slab ocean” in the remote tropics. Compute 30S-30N areal-mean temperature evolution assuming mass conservation (“what goes up comes down”) over the domain.

Lintner & Chiang [2005]
WTG and Amazonian Seasonality*

Anber et al. [2015]*: Consider a high resolution cloud-resolving model (CRM), in this case the Weather Research and Forecasting (WRF) model. The CRM domain is coupled to a “parameterized” large-scale. This is accomplished by relaxing potential temperature toward a target profile over a timescale characteristic of gravity wave propagation out of the domain.

*We will discuss the results of this analysis in more detail in Lecture 5.
Quasi-equilibrium Tropical Circulation Model (QTCM): Overview

QTCM is an intermediate level complexity model developed by David Neelin’s group at UCLA [Neelin & Zeng 2000; Zeng et al. 2000].

**Theory:** Approximate analytic solutions for vertical structure in tropical deep convecting regions consistent with convective quasi-equilibrium (QE) constraints can be determined.

**Implementation:** Primitive equations are projected onto analytic solutions in a truncated Galerkin expansion in the vertical.

CQE constrains \( T \Rightarrow \text{vertical structure of baroclinic } \rho \text{ gradients } \Rightarrow \text{vertical structure of } v \Rightarrow \text{vertical structure of } \omega \\

QTCM includes full complement of GCM-like parameterizations (e.g., radiative transfer, surface turbulent exchange, Betts-Miller convection); is computationally efficient; and has been applied to an array of problems in tropical climate dynamics (e.g., ENSO teleconnections, monsoons, global warming,...).
Primitive equations and vertical expansion

Temperature:
\[
(\partial_t + D_T)T + \omega \partial_p s = Q_c + g \partial_p R^\uparrow - g \partial_p R^\downarrow - g \partial_p S + g \partial_p F_T
\]

Moisture:
\[
(\partial_t + D_q)q + \omega \partial_p q = Q_q + g \partial_p F_q
\]

Horizontal Momentum:
\[
(\partial_t + D_V) \mathbf{v} + f \mathbf{k} \times \mathbf{v} + g \partial_p \tau = -\nabla \int_p^{p_s} \kappa T d \ln p - \nabla \phi_s
\]

Continuity:
\[
\omega = \omega_s + \int_p^{p_s} \nabla \cdot \mathbf{v} dp
\]

\[
T = T_r(p) + \sum_{k=1}^{K} a_k(p) T_k(x, y, t)
\]

Temperature projection. Note the reference profile, which is chosen based on a “typical” tropical convective region sounding.

\[
\mathbf{v}(x, y, p, t) = V(p) \mathbf{v}_T(x, y, t)
\]

\[
V(p) = (a_1^+(p) - \hat{a}_1^+)
\]

\[
a_1^+(p) = \int_p^{p_s} a_1(\hat{p}) d \ln \hat{p}
\]

Velocity projection. The vertical structure of temperature (here “first baroclinic”) constrains the velocity structure.
QTTCM1 basis functions

Neelin & Zeng [2000]
Applicability of the $T$ basis function

Convective region: $T$ constrained

Adjacent region: $\nabla T$ modest so vertical structure ok

Far from convection: Just truncated vertical representation

Radius of Deformation $\sim 25^\circ$

Fig. 1. Schematic of how the basis function for deep temperature structure chosen appropriately for a QE convective parameterization may be expected to approximate the solution when applied more generally. The solid curve indicates the model representation of temperature, $T_{\text{model}} = T_r + a_i(p)T_1$, where the reference profile $T_r$ is independent of space and time (dashed curve). In convective regions, $T_{\text{model}}$ tends to be close to actual temperature $T$ (thin curve) since both are close to the convective QE temperature (not shown). Far from convective regions, $T_{\text{model}}$ matches the vertical average of $T$, but vertical structure may deviate.

Neelin & Zeng [2000]
DJF Precipitation

Lintner et al. [2012]
Convective margins: theory & applications

Research motivating question: Why does it rain where it does, when it does, and how much it does in the tropics?

Framework discussed here: Convective margins

– Perspective is to consider the spatial transition zones between strongly convecting and non-convecting regions; we call these **convective margins**.
– The energy and water budgets within strongly convecting and non-convecting regions are quite different.
– Viewed in the context as a spatial “transition to strong convection,” rainfall within the margins can vary strongly: the mean convective environment exists “near” the threshold for strong convection to occur. [Note that the margin may be difficult to identify. ITCZs are rather sharp; continental convection zones tend to be much more smeared out.]
– Specific question: What factors determine the margins (“edges”) of tropical convection zones?

Applications:

– El Niño/Southern Oscillation (ENSO) tropical teleconnection over South America
– South Pacific Convergence Zone (SPCZ)
– Land-atmosphere interactions [Return to this on Friday]
Convective margins and society

Human Population Density (1994)

$2 \text{ mm day}^{-1} \leq P \leq 4 \text{ mm day}^{-1}$
Convective margins and society

Large potential risk to human welfare associated with droughts in tropical margins.
Convective margin variability

Ave. for ocean gridpoints (20°S-20°N)
Ave. for land gridpoints (20°S-20°N)

Ratio of monthly precipitation standard deviations to mean values points to relatively strong variability where mean precipitation is lowest.

*Computed from 2.5° x 2.5° CMAP dataset (1979-2006), with averages over 1 mm/day bins.
Current generation models show considerable spread and disagreement with the observations along certain convective margins, e.g., the South Pacific Convergence Zone (SPCZ).
Projected changes in rainfall

Intermodel comparison of JJA $P$ change under SRES2 forcing

Current generation models generally project the largest decreases in tropical precipitation along convective margins, but with disagreement with where these occur.

Neelin et al. [2006]
Convective margins prototype

Lintner & Neelin [2009]
Outline of margins framework

1. Vertically-integrated tropospheric temperature:
\[ \partial_t T = -M_s \nabla \cdot \vec{v} - \vec{v}_T \cdot \nabla T + Q_C + F_T \]
\[ F_T = R_{toa} + R_{surf} + H \]

2. Vertically-integrated tropospheric moisture:
\[ \partial_t q = -M_q \nabla \cdot \vec{v} - \vec{v}_q \cdot \nabla q + Q_q + F_q \]
\[ F_q = E \quad Q_c = -Q_q = P \]

3. Surface Temperature:
\[ \partial_t T_s = F_{T_s} \]
\[ F_{T_s} = R_{surf} + E + H \]

In the nonconvecting \((P = 0)\) region, for steady state conditions along a generalized inflow coordinate \(x\) with negligible temperature gradients:

1. & 3. \(\rightarrow\) \(\nabla \cdot \vec{v} = M_s^{-1} R_{toa}\)

Assuming a dry surface \((E = 0)\):

2. \(\rightarrow\) \(q(x) = q_0 e^{\lambda_0 x}\)
\[ \lambda_0^{-1} = v_q M_s (M_{qp} R_{toa})^{-1} \]

For a convective “moisture threshold” \(q_c\):
\[ x_c = \lambda_0^{-1} \ln(q_c / q_0) \]

Lintner & Neelin [2007]
Consider the limit $\tau_c = 0$ and a mean margin position $x_c$ with $P = P_0$. In the presence of Gaussian distributed windfield perturbations $\delta v_q$ (of standard deviation $\sigma_{v_q}$):

$$P(x) = \frac{P_0}{2\sigma_{\delta x_c} \sqrt{\pi}} \int_{x-x_c}^{\infty} d\delta x_c \exp(-\frac{\delta x_c^2}{2\sigma_{\delta x_c}^2})$$

$$\delta x_c = \delta v_q (x_c / v_q)$$

Lintner & Neelin [2007]
Methodology: Composite analysis

For a composite index with elements $X_i$ and mean $\bar{X}$, compute its standard deviation, $\sigma_X$:

$$\sigma_X = \sqrt{(N-1)^{-1} \sum_{i=1}^{N} (X_i - \bar{X})^2}$$

Define positive ($pX$) and negative ($nX$) index phases:

$$pX_i = \theta(X_i - \bar{X} - \sigma_X) \quad nX_i = \theta(\sigma_X + \bar{X} - X_i)$$

$$\theta(x) = \begin{cases} 
1 , & x \geq 0 \\
0 , & x < 0 
\end{cases}$$

For a field $Y$, the positive ($pY$) and negative ($nY$) composite phases are given by:

$$\bar{pY} = \frac{\sum_{i=1}^{N} pX_i Y_i}{\sum_{i=1}^{N} pX_i} \quad \bar{nY} = \frac{\sum_{i=1}^{N} nX_i Y_i}{\sum_{i=1}^{N} nX_i}$$

*A composite difference is defined as $\bar{pY} - \bar{nY}$.

Lintner & Neelin [2007]
Regional anomalies during El Niño

El Niño conditions: margin displaced toward continental interior

La Niña conditions: margin displaced toward the Atlantic coast

4 mm day$^{-1}$ El Niño (NINO3 $> \sigma$)

4 mm day$^{-1}$ La Niña (NINO3 $< -\sigma$)

SON CMAP (1979-2000) Precipitation Composites

Lintner & Neelin [2007]
Regional anomalies during El Niño

Average over 3.75S-1.75S

El Niño (NINO3 > $\sigma_{\text{NINO3}}$)

La Niña (NINO3 < - $\sigma_{\text{NINO3}}$)

La Niña conditions shift the margin toward the Atlantic coast...

El Niño conditions shift margin toward the continental interior

El Niño - La Niña

Lintner & Neelin [2007]
The fraction of tropical continental land area experiencing drought scales with the strength of El Niño forcing. In the convective margins perspective, this would correspond to contraction of tropical continental convecting areas. 

*Lyon [2004]*
Diagnosing El Niño-related margin shifts

\[ \frac{\delta x_c}{x_c} \approx \frac{\delta v_q}{v_q} - \frac{\delta R_{toa}}{R_{toa}} + \left[ \ln \left( \frac{q_c}{q_0} \right) \right]^{-1} \frac{\delta q_c}{q_c} - \left[ \ln \left( \frac{q_c}{q_0} \right) \right] \frac{\delta q_0}{q_0} \]

<table>
<thead>
<tr>
<th>Term</th>
<th>Mechanism</th>
<th>Impact on ( x_c )</th>
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<tbody>
<tr>
<td>( \frac{\delta q_c}{q_c} )</td>
<td>T increases ⇒ Increased ( q_c )</td>
<td>+</td>
</tr>
<tr>
<td>( \frac{\delta q_0}{q_0} )</td>
<td>SST increases ⇒ Increased ( q_0 )</td>
<td>-</td>
</tr>
<tr>
<td>( \frac{\delta R_{toa}}{R_{toa}} )</td>
<td>T increases ⇒ Increased OLR</td>
<td>+</td>
</tr>
<tr>
<td>( \frac{\delta v_q}{v_q} )</td>
<td>Trades strengthen</td>
<td>+</td>
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Notation: + shift toward interior; - shift toward coast