Lecture 4 Topics

• Wrap up from Lecture 3
• Overview of representation of convection in models, especially parameterization
• Errors and biases in models
Cloudscape in the SW Pacific
Research related to the SPCZ

- Synoptic interplay of circulation, moisture, and precipitation along the SPCZ in observations, reanalysis, and CMIP5 models [Matt Niznik, former PhD student]
- Some preliminary results from applications of self-organizing maps (SOMs) to daily precipitation in the SPCZ [Max Pike, current PhD student]
- “Principal uncertainty patterns” to characterize spread in simulated SPCZ behavior across CMIP5 models [analysis by Baird Langenbrunner (UCLA)]
Geography and climate of the South Pacific

From Chapter 2 of the Pacific Climate Change Science Program’s *Climate Change in the Pacific: Scientific Assessment and New Research.*
Context: Flooding in Fiji, April 2012
Context: ENSO and “Zonal SPCZ” events

Cai et al. [2012]
SPCZ research motivation and background

• **Question:** How is the SPCZ generated and maintained and what factors lead to its distinct spatial characteristics?

• **Takahashi & Battisti [2007]** distinguish two broad paradigms:
  – **Western control:** SPCZ develops in response to convective heating over the western Pacific warm pool *[Lindzen & Nigam 1987; Kiladis 1989; Matthews et al. 1996; Widlansky et al. 2010]*
  – **Eastern control:** SPCZ characteristics can be understood primarily in terms of advection of low moist static energy (cool & dry) air masses from the southeast tropical Pacific and the distribution of SSTs and large-scale subsidence there *[Derbyshire et al. 2004; Takahashi & Battisti 2007; Lintner & Neelin 2008]*

• **Issues in model simulation:**
  – Well-known biases in the broader Pacific include the double ITCZ and eastern Pacific cold tongue *[Li 2004; Lin 2007]*
  – Simulated SPCZs are often too zonal, with deep convection penetrating too far to the east *[Brown et al. 2011]*
Composite Analysis

Daily (SSMI)
Winds from NCEP reanalysis

*Compositing index is $u_{925}$ mb averaged over a 140°W-120°W and 20°S-10°S. Vectors (top) show the composite difference in $(u_{925},v_{925})$.

4 mm day$^{-1}$ (+ve)
4 mm day$^{-1}$ (-ve)

Lintner & Neelin [2008]
Longitudinal cross-sections

(a) Zonal wind
- Trades relax

(b) Column water vapor
- Moisture increases

(c) Precipitation
- Margin shifts

Lintner & Neelin [2008]
Within the convecting region \((P>0)\)

1. & 2. \[ \nabla \cdot \vec{v} = M_c^{-1} (R_{net}^{\text{clear}} + CP - \nu_q \partial_x q) \]

If \(P\) is assumed of the form \(\tau_c^{-1}(q - q_c)\), for sufficiently small \(\tau_c\), \(q \approx q_c\) and \(M_q \approx M_{q_p} q_c\) and thus:

\[ q(x) \approx q_c + (1 - e^{\lambda_c(x-x_c)}) \tau_c [M - M_q C]^{-1} [M_s E + M_q R_{net}^{\text{clear}}] \]

where \(\lambda_c^{-1} = -\nu_q \tau_c [1 - M_c M_s^{-1}(1 + C)]^{-1}\) is a characteristic length scale; \(C\) is a cloud-radiative feedback factor; and \(R_{net} = R_{net}^{\text{clear}} + R_{net}^{\text{cloud}} = R_{net}^{\text{clear}} + CP\) is the total clear-sky and cloudy-sky column radiative forcing.

The precipitation field is just:

\[ P(x) \approx (1 - e^{\lambda_c(x-x_c)}) [M_c - M_q C]^{-1} [M_s E + M_q R_{net}^{\text{clear}}] \]
Simple 2D prototype

a) Prec, q and u, v

b) Prec, q comp diff on u

c) Prec, q and u, v (q_c reduced)

d) Prec, q comp diff on u (q_c reduced)
Relating SST and precipitation
Relating SST, precipitation, and winds

There are essentially two paradigms:

1. Deep tropospheric thermodynamic control: SST determines the vertical profiles of temperature and moisture which determine precipitation/convective heating; the convective heating determines the winds.
   → Essentially, what we have discussed thus far (CQE, WTG, QTCM1) aligns with this paradigm.

   → This paradigm is not without challenges: for example, first baroclinic-type models (e.g., QTCM1) have trouble simulating ITCZs.

2. ABL dynamic control: SST determines the low-level winds through pressure gradients, the associated low-level moisture convergence feeds precipitation.

   → For example, in Lindzen & Nigam [1987], surface fluxes determine ABL temperature, which determines surface pressure directly from hydrostatic balance.

   → In this paradigm, moist physics and radiation are largely irrelevant for determining the winds [except to the extent that these contribute to setting the SST in the first place].

While these paradigms are distinct in terms of causality [convection→winds in 1 vs. winds→convection in 2], they are quite difficult to tease apart because in reality, convection ↔ winds. We can regard this as the convection-convergence feedback.

For a nice discussion, see the introduction to Sobel & Neelin [2006].
The vertical expansion in QTCMs can be extended to an arbitrary number of vertical modes. For QTCM2, a second set of basis functions were implemented by inserting an explicit, fixed depth, prognostic atmospheric boundary layer (ABL) beneath the free troposphere (FT).

**Theory:** This approach is directly motivated by wanting to enable both paradigms described on the previous slide to be operational in the model.

**Methodology:** The basis functions for the ABL assume a well-mixed layer and so are vertically uniform. The basis functions in the FT are modified forms based on QTCM1.

*Sobel & Neelin [2006]; Lintner et al. [2012]*
QTMM2: Better (more complete) physics... but is it better?

Lintner et al. [2012]
Why do we parameterize convection?

As we discussed in Lecture 2, the purpose of convection parameterization is to represent the net effect of the unresolved, subgrid-scale physics of convection in terms of the resolved, grid-scale dynamics and thermodynamics.

A convection parameterization yields precipitation as well as convective heating and drying profiles and vertical momentum transport that feed into the large-scale evolution. Parameterizations also provide information relevant for diagnosing cloud cover and radiation, and in the context of atmospheric chemistry, vertical transport of trace constituents.

As Plant [2010] notes “The parameterization of precipitating convection for both general-circulation and numerical weather prediction models is a notoriously stubborn problem.”
The components of a convection scheme

There are three components of model convection scheme: **triggering**, **closure**, and the **cloud model**.

**Triggering** (or triggering function) determines whether convection will occur. The triggering evaluates the stability of the resolved scale profile using parcel method techniques: for conditional instability, convection will occur.

**Closure** determines the strength of convection. We will discuss various methods for closure in the following slide.

A **cloud model** represents the physics of convection, including updrafts, downdrafts, and their transport of properties like temperature, moisture, momentum, and other scalars. The cloud model also represents the lateral interactions between cloudy area and its cloud-free environment. The latter relates to the processes of **entrainment** and **detrainment**.
Convec7ve closures

There are many possible ways of closing the convection problem.

One possible closure is based on low-level moisture convergence. This closure tends to produce a positive feedback between low-level moisture convergence and precipitation. The *Kuo* [1965] and *Tiedtke* [1989] schemes use this type of closure.

CAPE-based closures are used in many schemes. These closures are grounded in CQE assumptions, albeit with CAPE relaxed over an adjustment timescale. As we have discussed, a range of values exists for the adjustment to occur.

Some more recent closures are based on boundary layer processes and are effectively tied to a comparison of turbulent kinetic energy in the boundary layer to CIN. The *Bretherton et al.* [2004] scheme is an example.
Single plume mass flux

In a single plume mass flux, the bulk effect of clouds with a gridbox are represented by models of a single updraft (and downdraft). The governing equations for the updraft are of the following form:

\[
\partial_z M^u = E^u - D^u
\]

\[
\partial_z (M^u \phi^u) = E^u \bar{\phi} - D^u \phi^u + S
\]

Here the superscripts \( u \) indicate properties of the updraft (cloudy region) while overbars represent properties of the cloud-free environment. Rates of entrainment \( E \) and detrainment \( D \) represented as products of entrainment or detrainment coefficients and updraft mass.
Shallow convection

Thus far, we have been focusing on deep convection, but convection can also be shallow.

Models may use a unified scheme, in which the model first determines whether convection should be shallow or deep (depending on the large-scale thermodynamics present). Models may also use separate schemes for each of shallow and deep convection, which in contrast to the previous schemes, allows both shallow and deep convective clouds to be present at the same time.
Comparing different convection schemes

An illustration of the impact of using a different convection scheme in the same coarse resolution model.

<table>
<thead>
<tr>
<th></th>
<th>Tiedtke</th>
<th>ECMWF</th>
<th>Zhang-McFarlane-Hack</th>
<th>Bechtold</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Closure</strong> (deep)</td>
<td>CAPE / moisture convergence</td>
<td>CAPE</td>
<td>CAPE</td>
<td>CAPE</td>
</tr>
<tr>
<td><strong>Entrainment</strong></td>
<td>turbulent and organised moisture</td>
<td>turbulent</td>
<td>turbulent</td>
<td>turbulent</td>
</tr>
<tr>
<td><strong>Closure</strong> (shallow)</td>
<td>moisture convergence</td>
<td>1.) moist static energy</td>
<td>moist static energy</td>
<td>CAPE</td>
</tr>
<tr>
<td><strong>Trigger condition</strong></td>
<td>$T_\varphi + \Delta T &gt; T_\varphi^{\text{env}}$  $\Delta T = 0.5 \text{K}$</td>
<td>$w_\kappa &gt; 0$ with $w_\kappa$ from entrainment and buoyancy (Jakob and Siebesma, 2003)</td>
<td>$T_\varphi + \Delta T &gt; T_\varphi^{\text{env}}$  $\Delta T = 0.5 \text{K}$</td>
<td>$\Theta_\varphi + \frac{\Delta T}{\varpi} &gt; \Theta_\varphi^{\text{env}}$  $\Delta T = 6 \cdot</td>
</tr>
<tr>
<td><strong>Precipitation formation</strong></td>
<td>$\Delta r = r^c_\kappa / (1 + c_\tau \cdot \Delta z)$</td>
<td>proportional to $1/w_\kappa$</td>
<td>$\Delta r = c_0 \cdot r^c_\kappa$</td>
<td>$\Delta r = (1 - \beta) \cdot r^c_\kappa$</td>
</tr>
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</table>

Source: [http://www.lmd.ens.fr/wavacs/Lectures/Tost-2.pdf](http://www.lmd.ens.fr/wavacs/Lectures/Tost-2.pdf)
Comparing different convection schemes

http://www.lmd.ens.fr/wavacs/Lectures/Tost-2.pdf
Challenges for modeling: MCSs

As discussed in Lecture 1, most tropical precipitation occurs in MCSs. However, current generation models fail to represent organized structures like MCSs.

One specific aspect that is problematic is the partitioning of rainfall in convective vs stratiform: in general, current-generation climate models typically simulate too much convective (intense) rainfall.

It is important to note that this can have deeper implications, since convective and stratiform precipitation may have distinct convective heating and moistening profiles.

Dai [2006]
Challenges for modeling: diurnal cycle

Over oceans, models tend to underestimate the amplitude of the diurnal cycle relative to observations.

Over land, models tend to have the wrong phasing, with maxima occurring too early in the day.

*Dai [2006]*
Challenges for modeling: the double ITCZ

Coupled GCMs exhibit a well-known tendency to simulate two ITCZs in the eastern Pacific as opposed to the single observed ITCZ.

While this tendency has been attributed to the coupling itself [i.e., in atmosphere only models with prescribed SSTs, the bias is reduced], there is some suggestion that incorrect moisture sensitivity in the free troposphere can contribute [Oueslati & Bellon 2013].

Niznik et al. [2015]
Free tropospheric moisture sensitivity

Approach: Apply an “entrainment-like” in the Betts & Miller QTSM2 convection scheme to alter the sensitivity to free tropospheric moisture.

Lintner et al. [2012]
Shallow convective preconditioning

Approach: Lengthen the timescale for shallow convective adjustment in QTCM2 to decrease the strength of shallow convective preconditioning, i.e., free tropospheric moistening through the effect of shallow convection.

Lintner et al. [2012]