Analyzing Characteristics of Convection and the Relationship with its Environment

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UNIVERSITY OF MIAMI

ANALYZING CHARACTERISTICS OF CONVECTION AND THE RELATIONSHIP WITH ITS ENVIRONMENT

By

Si Won Song

A DISSERTATION

Submitted to the Faculty of the University of Miami in partial fulfillment of the requirements for the degree of Doctor of Philosophy

Coral Gables, Florida

December 2015
UNIVERSITY OF MIAMI

A dissertation submitted in partial fulfillment of
the requirements for the degree of
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WITH ITS ENVIRONMENT

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Analyzing Characteristics of Convection and the Relationship with its Environment (December 2015)

Abstract of a dissertation at the University of Miami.

Dissertation supervised by Professor Brian E. Mapes.

No. of pages in text. (152)

Interactions between atmospheric deep vertical convection and larger-scale flow have been examined in three diverse modeling frameworks: a traditional climate model with cumulus parameterization scheme, an extremely computation intensive global convective-resolving model, and a simplified global primitive equation model with a linearized anomaly convective scheme. The first model is operational but has some chronic problems that call for more research, the second is promising but produces vast amounts of output data that are hard to interpret or even handle, and the third is satisfying for interpretation of the interaction processes, but illustrate some continuing challenges of atmospheric modeling, ultimately because this important process occurs in convective cells much smaller than the size of our planet.

More specifically, in chapter 2, hindcast bias growth in the Climate Forecast System (CFS) is analyzed. Errors in the Cold Tongue – Intertropical convergence Complex (CTIC) are apparently initiated by the convection, since they appear very rapidly. The errors are interpreted as indicating weak sensitivity of the convection scheme to moisture, a common problem in climate models. This initial convective error apparently seems
coupled feedback processes which gradually spread bias errors to other regions and components of the CFS.

In chapter 3, explicitly simulated tropical convective rain events were examined, from a pioneering global nonhydrostatic 5-km mesh model (NASA GEOS-5). The data examined are like perfect observations – they are samples at full resolution – but unlike observations the values are known exactly. Composite profiles of larger-scale temperature and humidity evolution across the rain events show good agreement with published composites of observations, but not every case has all the composite characteristics. Diverse interaction mechanisms between convection and its environment are seen in the various cases, as in nature, indicating the model’s realism in that broad sense.

In chapter 4, a linear matrix formalism for convection-large scale state interaction is explored. Heating and moistening rate anomalies are cast as weighted sums of temperature and moisture anomalies, based on pioneering work of a collaborator, Dr. Z. Kuang of Harvard. His matrix $M$ is tested as a diagnostic model to help interpret the composite data from chapter 3. The composite humidity anomalies are found to be more consequential than temperature in shaping the evolution of convection, based on $M$’s weighting factors. The static stability implied by composite-averaged cool air at the surface, which is a consequence rather than cause of heavy rain, reduces the $M$-predicted rainrate, indicating one of the challenges of framing causality in terms of large-scale variables alone.

In chapter 5, $M$ is applied as an anomaly convection scheme in a global primitive equation model. Five experiments are described with modified versions of $M$. The modifications are motivated by a wish to understand the roles of $M$’s eigenmodes, and on
the hypothesis from chapter 2 that free tropospheric moisture sensitivity is an important aspect of convection schemes. The experiments show substantial differences in large-scale M-coupled phenomena, although these first-ever model simulations are too unrealistic to judge in terms of observations.

Several general conclusions emerge. Clearly the treatment of convection is important to large-scale climate and weather, especially in the tropics. Explicitly resolving cloud systems appears to be a promising approach as computation power grows, but will not be affordable for all climate problems. Still, output datasets from such models can be a useful resources for trying to improve parameterizations through better understanding of interaction processes as played out in the range of weather scenarios that occur in the tropics. Based on the linearized matrix approach, some clean interpretations can be deduced. For example, moisture sensitivity of convection does indeed appear to be a key issue for convection interactions, and having a new model where the sensitivity can be cleanly modified and tested could lead to knowledge that may feed back into improvements in conventional cumulus parameterizations.
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Chapter 1: Introduction

A subtitle of this dissertation might be “Three views of convection-global interaction”. These are distinctive, they are not the only views, and the thing they have in common is their overall subject.

1.1 Seamless nature vs. discontinuously simulated convection

The subject of this dissertation is the role of convection in the large-scale flow of the seamless and interactive atmosphere. The atmosphere cools by infrared radiation and is heated and moistened from below. Vertical convection occurs because of this destabilization, often in moist form as saturated or cloudy updrafts. Convection has basic cellular scales related to PBL and troposphere depth (1-10 km). On the other hand, the atmosphere also has equator to pole heating gradients, which drive planetary scale ($10^4$ km) flows, and these are baroclinically unstable to synoptic eddies ($10^3$ km). In between convective and synoptic scales, cloud systems often have mesoscale size, which ‘meso’ means ‘middle’ ($10^2$ km). Mesoscale motions are partly the fine scales of planetary and synoptic flows, as deformation of airmasses makes narrow filaments with sharp gradients. But they are also partly the aggregated patterns of convective activities.

Conventional global models need convective parameterizations including mesoscale because their coarse grid (~ 100 km) cannot resolve small and meso scale motions. Grid spacings near 10km are called the “grey zone” of resolution (Gerard et al. 2009, Yu and Lee 2010). Mesoscale flows are resolved, but convective cells are not quite resolved, but
also grids do not contain a statistical ensemble as assumed for cumulus parameterization. It is unclear or grey what is the right strategy for models with such resolution: cumulus parameterization or not? Especially, Mesoscale Convective Systems (MCS) are observed to link basic cellular convection with many significant larger-scale phenomena like Convectively Coupled Equatorial Waves (CCEW), Monsoons, Madden-Julian Oscillation (MJO), or Inter-Tropical Convective Zone (ITCZ), but MCS are poorly simulated in coarse grid global model because meso and convective scales are both missing (Del Genio 2012). Although there are a few attempts to parameterize the MCS effects like thermodynamic characteristics (Donner 1993, Donner et al. 2001) or organizing new convection (Mapes and Neale 2011), it is quite new and ongoing subject.

1.2 Abundant observations and modeling works over the past decades

For better understanding about convection in the large-scale framework, this dissertation makes three main efforts to use uniquely available new data wisely. Over the past decades, there have been abundant observations and modeling works for understanding atmospheric phenomena. However, whether or not any single observation (or model) can capture (or simulate) aspects of convective phenomena reasonably, each data set itself could not be perfect because all data has its own limitations: 1) the conventional convective parameterization in the traditional global models is designed by its author’s specification for the role of convection in the unresolved scales (Arakawa 2004). 2) Unlike the conventional convective parameterization, superparameterization can resolve convective scales explicitly as well as synoptic scales (Randall 2013), but it still has a mesoscale gap enforced by the unrealistic assumption like the simplified two-dimensionality and periodic lateral boundary conditions. 3) Limited-area cloud-resolving
models (CRMs) have artificial boundary conditions unlike nature. 4) Global cloud-resolving models cannot provide the causality and are still in progress to resolve all the scales. 5) The field data is limited to a few time periods as well as locations. 6) Satellite data also have an issue about spatial and temporal coverage as well as the not-sufficiently fine resolution, especially in the vertical, depending on satellites. And finally, 7) causality between convection and its environment is difficult to interpret or infer in observations or simulation outputs because of mixed feedback.

1.3 Three approaches based on different modeling frameworks

This dissertation surveys three model-based approaches to understand the interaction of convection with larger scales: (1) evaluating a classical GCM, showing its limits, (2) a direct new approach of examining global high resolution data, and (3) an indirect new approach with matrix linking inputs (thermodynamic profiles) and outputs (tendencies).

In chapter 2, errors in a traditional parameterized General Circulation Model (GCM) for convection-global interaction are shown, as published in Song and Mapes (2012). Specifically, I examined a 20-year hindcast set from version 1 of the NCEP Climate Forecast System (CFS), illustrating the possible error sources and feedbacks that followed those initial errors through the systematic error growths.

In chapter 3, samples from the big data of a global cloud-resolving model are analyzed like observations. Specifically, I examined the NASA Goddard Earth Observing System global atmospheric model (GEOS-5) simulation with a 5-km horizontal resolution by (Putman and Suarez 2011). In this grey zone simulation, convection parameterization was used, but in restricted form. The eight heaviest rain events at 2.5 degree scale were selected and the spatial development around the event time is analyzed
to examine the interaction with its environmental condition. Although we cannot expect
to derive full information about causality between convection and its environment
through this analysis, it is a direct approach to use the explicitly simulated convection
data like observation and provides fairly realistic behavior of convection to evaluate and
study.

In chapters 4 and 5, convection-environment interaction are analyzed using linear
algebra, and this linear relationship is applied to diagnosis of the observation(-like) data
(chapter 4) as well as to a prognostic global model (chapter 5). The global model used in
this study (called “Jalopy with matrix”) separates the mean flow effects (maintained by
fixed case-independent and time-independent forcings) from large-scale weather
variability (created by large-scale dynamics coupling with anomalous tendencies due to
anomalous convection), so it would be clearer than classical full physics GCMs to isolate
mechanisms of convection-dynamics interaction. It is an idealized and indirect approach,
but provides a clever way to understand their interaction through the power of linearity
for decomposability.

Background, data and methodology are in each chapter, and summary and conclusion
through all three independent approaches are in chapter 6. Appendices include several
cases of systematic error growth in NCEP CFS (Song and Mapes 2012) other than
convectively deduced case in chapter 2, derivation of composite and regression methods,
and a description of an unsuccessful but educational effort to discover linear matrix
relationship from field data.
Chapter 2. Fingerprints of convective errors on GCM

This chapter is one of the cases in Song and Mapes (2012) that is about the several robust aspects with interesting error growth with lead time. This chapter is about the “Cold Tongue – ITCZ Complex (CTIC) error feedback” initiated by the short-lead “convective” bias. Rest of error growth aspects in Song and Mapes (2012) are in appendix A.

2.1 Background

2.1.1 Systematic error

Numerical forecasts of the atmosphere-land-ocean system, initialized from observational analyses, tend to have growing errors (differences from the verifying analyses). Forecast errors may be divided into systematic (mean bias) and random (case-dependent) parts. Case-dependent errors can be hard to identify and fix, as they may depend on the weather state, the quality of initial analysis, etc. Systematic forecast errors on the other hand may be direct indicators of model errors, so understanding them may help inform model development efforts – if they can be interpreted. This paper describes one attempt to do so.

Forecast errors tend to grow with lead time until, at some asymptotically long lead, the skill due to memory of the initial condition is gone. At this point, the error is statistically as big as the error of taking a random state of the free running model as a forecast. This no-skill lead time will differ between an atmospheric and a coupled model. Atmospheric errors grow quickly, with weather time scales of hours to days, while oceanic processes are slower due to large thermal and mechanical inertia, so aspects of

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1 The rest of aspects except convective error initiation in Song and Mapes (2012) are the different error time scales depending on the land and the ocean, errors of hemispheric-scale jet stream (polar vortex) dynamics via the thermal wind balance relationship, and errors that decay with lead time instead of increase or are saturated.
the initial conditions can remain to add skill for several months. Land processes have a
timescale in between these. Considering systematic errors as a function of forecast lead-
time thus gives a way to interpret the sources of errors according to the time scale of their
emergence. However, systematic errors do not necessarily grow monotonically: for
example, ocean errors can compensate errors from faster-responding components of the
system.

2.1.2 Advantages of NCEP CFS climatology forecast

For diagnosing the error sources in GCMs, climatology forecast data from the NCEP
CFS could be used properly. The NCEP CFS version 1 (Saha et al. 2006) is a fully
coupled ocean-land-atmosphere dynamical prediction system, but with sea ice specified
and Sea Surface Temperature (SST) prescribed poleward of 74°S and 64°N. A hindcast
set from the years 1981–2004 is available, with model initializations at various days (365
days) throughout the year. The main use for such a hindcast set is to allow CFS’s climate
forecasts to be expressed as anomalies from the model’s own imperfect climatology. But
these imperfections of the hindcast climatology also contain interesting information about
the model and how the coupled components work together. Artifacts from the
assimilation and initialization schemes are beyond the present scope, but are minimized
by use of the CFS’s own 1-day forecast state as the verification.

2.2 Data and methodology

The hindcast climatology of 30 atmospheric variables on a 2.5° × 2.5° longitude-
latitude grid was obtained from http://cfs.ncep.noaa.gov/cfsv1.html, where many more
details of the system are documented (Johansson et al. 2007, available at
http://cfs.ncep.noaa.gov/menu/doc/). For each initialization date, the model wrote out
data every 12 hours at 539 forecast times, from 24 hours up to 6480 hours (9 months). We made daily means by averaging 00 UTC and 12 UTC output (noting and correcting for an apparent error in the 00Z vs. 12Z labeling of the data files as downloaded in 2009). We shifted the data arrays from this model-output structure (organized by initialization date and forecast hour) to a more convenient form for interpretation (by verification date and lead time). We then averaged the verification dates into seasons (DJF, MAM, JJA, SON, and annual mean), since systematic errors are thought to be related to physical processes and thus behave similarly throughout a season.

A number of simplifying assumptions should be admitted right away: First, we used only atmospheric variables since daily output was available. In particular we take 2-meter temperature (T2m) as a proxy of SST and land surface temperature errors. We used the CFS’s own 1-day forecast climatology as reference for verification, for fields other than clouds and radiation (which are known to be biased even at initial times relative to satellite observations) (Huang et al. 2007). Therefore, error is defined here by subtracting the climatology of the 1-day forecast (treated as verification data) from climatological fields at longer lead times. To further reduce the large data volume and focus on time scale (a logarithmic concept), errors at lead times of 2, 4, 8, 16, 32, 64, 128, and 256 days are extracted. That reduced data set was examined comprehensively, in the form of animations of error maps with lead time, for verification times averaged over each season and the annual mean. These animations may be viewed at http://www.rsmas.miami.edu/users/bmapes/CFS1biases.html.
2.3 Results and discussion

2.3.1 Cold Tongue – ITCZ Complex (CTIC) error

The Intertropical Convergence Zone (ITCZ) is a feature of the Hadley circulation, but its detailed structure involves coupling also with oceanic processes in the equatorial upwelling zones or ‘‘cold tongues’’ of the eastern Pacific and Atlantic (Pike 1971, Mitchell and Wallace 1992, Murakami et al. 1992, Chang and Philander 1994, Waliser and Somerville 1994, Xie and Philander 1994, Philander et al. 1996, Wang and Wang 1999, Richter and Xie 2008). It has been studied in CFS data by Misra et al. (2009). The CTIC involves interactions among the atmospheric ITCZ (a system of winds and tropical convection, sensitive to SST and its gradients) and the oceanic cold tongues (cool upwelled water, sensitive to wind stress, wind-driven evaporation, and cloud-modulated radiation influences). Positive feedbacks are involved, as evidenced by the strong spatial contrasts in the mean CTIC. With positive feedbacks acting, modest errors in an atmosphere model can grow with time and reach large amplitudes in a coupled setting, making this a very sensitive and challenging system to simulate and predict.

The mean CTIC structure is shown in Fig. 2.1, using CFS’s 1-day forecasts of surface temperature (blue shading), precipitation rate (red contour), and wind at 10 m, in the CTIC region and peak season (SON). Deep convection in the ITCZ is centered around 5–10°N, fed by convergence of low-level winds, in each ocean basin. There are also cold-water regions off the west coasts of the continents, in regions occupied by stratocumulus cloud decks.
Systematic errors in the CTIC forecasts are shown in Fig. 2.2, in the form of differences from Fig. 2.1. Forecast leads of about a day, a week, a month, and a season are shown. Poleward precipitation and wind convergence shifts are seen already at day 2 (Fig. 2.2a), although surface temperature error is small at that stage. This could be interpreted as an atmospheric model error, presumably involving the deep convection scheme. It will be discussed in section 2.4.

The low-level wind and precipitation errors grow in the subsequent days, with wind converging into the enhanced precipitation areas and diverging from reduced precipitation areas (Fig. 2.2b). Meanwhile, SST error (as indicated by T2m error) grows with lead time, with a broad smooth pattern that is similar at all leads; mainly a warm bias in stratocumulus areas and cool elsewhere. The broad warm bias over the stratocumulus areas develops in response to too much sunlight reaching the ocean there (as shown in Huang et al.'s (2007) comparisons to satellite data). The temperature errors are almost saturated after 3 months (Fig. 2.2d), with similar magnitude at 256d (not shown). Presumably this error saturation reflects a steady balance between excess shortwave input of about 50 W m$^{-2}$ (reading off typical values from Huang et al. (2007)
compensated by excess surface fluxes plus entrainment at the base of the ocean mixed layer. If these compensating flux biases develop in response to surface temperature anomalies of about 2 K, the implied negative feedback factor of about 50/2 = 25 W m\(^{-2}\) K\(^{-1}\) is well within the typical 10–50 range depending on various ways the atmospheric state adjusts to SST errors (Frankignoul et al. 1998). Net longwave flux (upward minus downward) is surprisingly large in CFS (Huang et al. 2007, Fig. 14), presumably partly because of the dry bias in surface air humidity identified by Wu et al. (2007). This dry bias also serves to make the latent heat flux feedback \(\partial(LHF)/\partial(SST')\) larger. Detailed diagnosis is not possible with the fields available, but the point here is that the saturation of SST errors, although driven originally by cloud errors, induces errors in other aspects of the model state.

Another notable feature of Fig. 2.2 is a localized equatorial Pacific cold bias that appears from day 8 to day 32. This cold tongue bias appears dramatically for lead times longer than a month (Figs. 2.2c and d), in a region where the first-week bias was warm (Fig. 2.2b). Intense wind divergence errors are seen in association with this excessive cold water on the equator, perhaps through Lindzen and Nigam (1987) type boundary layer dynamics. These winds diverge from the equator and converge into the off-equatorial enhanced precipitation regions, reinforcing the initial ITCZ error. In other words, a slow (months to develop) oceanic upwelling and wind error acts as a positive feedback to enhance an initial (first day to week) error in the atmospheric deep convection scheme, similar to the mechanisms shown in Terray (1998) and Iizuka et al. (2003).
2.4 Discussion and summary

This chapter has made an examination of systematic error growth with lead time in the climatology of NCEP CFS hindcast outputs, especially focused on the initiated error by convection. The positive error feedback in the coupled CTIC complex appears to amplify initial atmospheric convection errors in the central Pacific (Fig. 2.2a), where off-
equatorial ITCZ wind convergence errors (Fig. 2.2b) and wind-driven equatorial upwelling of cold water synergize (Fig. 2.2d).

It shows two considerable points. At first, based on the rapid development of convective error, it presumably involves the deep convection scheme. Vertically integrated buoyancy (convective available potential energy or cloud work function) of weakly entraining air parcels has little sensitivity to free tropospheric dry air and shallow inversions. As a result, weak moisture sensitivity is an endemic problem (see, e.g., Derbyshire et al. 2004 and others in that special issue) of global model convection schemes based on those concepts. The CFS’s simplified Arakawa-Schubert scheme is of this type, so it is not surprising that its convection is excessive in the low-latitude edge of the trade winds just upstream of the ITCZ, where dryness in the free troposphere and a thin trade inversion are key limiters of convection in nature. Unfortunately, simply increasing entrainment in model convection has other unwanted consequences leading to unstable stratification bias in the mean state, so this error is a symptom of a compromise or “dilemma” that is common in atmospheric models with plume convection (Mapes and Neale 2011). One major change in CFS from version 1 (analyzed here) to version 2 is a dryness-dependent entrainment formulation in the scheme, a popular simple way to increase dryness sensitivity while minimizing side effects.

Secondly, because of the error initiated by the imperfect convective scheme, we can even expect to see the impact of those errors to the large-scale feedback at the interactive view. Even though it leads to large-scale false feedback, we can catch a glimpse of the interactive mechanism between different variables through the diagnosis of error growth in the atmosphere-land-ocean system like CFS. Along this line, it also shows that model
bias errors in different components of the CFS emerge at different rates. Atmospheric moist physics biases grow strong within a day (Fig. 2.2a) while stratocumulus cloud fields are inadequate even initially (Huang et al. 2007), and ocean model biases take months to fully saturate (Fig. 2.2c and d). Therefore, the examination of systematic error growth with lead time suggests the useful approach to see how different components interact each other in the large-scale and how those interactions lead to the noticeable atmospheric aspects.
Chapter 3: Examining explicit convection with a full global flow

One approach to the old shortcomings of parameterized convection in GCMs is to drastically decrease horizontal grid spacing to 10km or less, and use nonhydrostatic equations instead of the primitive equations. The hope is that explicitly resolving at least the mesoscale aspects of deep moist convection, including its connections to the synoptic and planetary scales, will be a major step forward even if the actual cumulus scale remains barely resolved. Such models are sometimes called “cloud permitting” rather than cloud resolving, or there are other terms such as the “grey zone” of resolution, where it is not clear if cumulus parameterization is needed, or whether explicit but rather coarse gridscale saturated updrafts can do the job. (Gerard 2007, Gerard et al. 2009, Yu and Lee 2010).

In this chapter, we examine some aspects of these pioneering simulations, in this case with data from GEOS-5 (Putman and Suarez 2011) at 5km (where we have just a few key fields) and 0.0625 degrees (the “Nature Run,” with many model fields available online in an OpenDAP server at http://opendap.nccs.nasa.gov/dods/OSS/G5NR/Ganymed/7km).

3.1 Background

Since high-resolution global models have become available, convection can be simulated explicitly without lateral boundary layer condition. Especially, convective systems may show realistic interactions with larger-scale conditions and flows. Moreover, because high-resolution model output provides seamless data with properly fine time resolution, individual case studies are also available for seeing detailed internal mechanisms of MCSs. In this chapter, selected MCSs in high resolution GEOS-5 data are analyzed in both composite and individual case studies.
3.1.1 Mesoscale Convective Systems (MCSs)

Convection-containing weather systems offer an opportunity for helping us to understand the interactions of different scales of motion. If various larger-scale convective systems are the stretched form of the ‘building block’ MCS (stretched building block hypothesis, Mapes et al. 2006), then understanding MCSs could provide important information for developing new convective parameterization.

Half of the sizes of cloud clusters by deep convection are over 100 km² (Mapes & Houze 1993, Houze 2004 review paper). The broad term MCSs includes quite large convective storms (10⁴-10⁵ km²) with various convective cells from small cumulus to cumulonimbus within it, and MCSs account for a large portion of precipitation in the tropics. Once MCSs link with large-scale dynamics such as equatorial waves (Wheeler and Kiladis 1999), this coupling extends their spatial scale and duration, and wave motions in the clear air between can modulate the population of clouds and MCSs. For these larger-scale modulations, triggering new convection convective system development is the key mechanism, since individual MCSs have a life cycle and die out. Therefore, triggering mechanism will be a key for new convective parameterizations.

3.1.2 Triggering mechanisms for new convection in MCSs and waves

Mechanisms of triggering new convection include both cold pool and non-cold pool mechanisms. Both mechanisms can generate new convection, both near (by cold pool) and far (by wave or other effects) from the previously matured ones (Mapes 1993, Mapes and Houze 1993).
3.1.2.1 Cold pool mechanism (at the edge of density current)

As evaporatively cooled downdrafts descend into the PBL, a density current is created at the leading edge of the cold pool. If this density current forces conditionally unstable environmental air up to its level of free convection (LFC), new convection can be triggered and can coalesce with previous convection to join a MCS (Houze 2004).

Such forced lifting of air is suggested as one of the triggering mechanisms for new convection, and the environmental stability and wind shear are examined as the lifting conditions in many studies (Moncrieff and Miller 1976, Moncrieff 1981, 1992, Thorpe and Moncrieff 1982, Moncrieff and Liu 1999). However, adiabatic ascent of clear air is also very helpful for destabilizing a mesoscale region and thus enhancing such effects in a mesoscale region to permit cell coalescence (Fig. 3.1 adopted from Crook and Moncrieff 1988). The nature of such forced ascent remained unclear in Crook and Moncrieff (1988), as they simply specified divergent forcing in the horizontal momentum equation to make this figure.
Two different methods of initiation of convection as lifting the air to the condensation level, warm bubble (upper) and large-scale convergence (lower panel). Solid is cloud regions, and hatched region indicate that air parcel need less than 100 m to be lifted for condensation. Adapted from Crook and Moncrieff (1988).

3.1.2.2 Non-cold pool mechanism (discrete propagation ahead of MCS)

While new convective cells may be almost always locally triggered by cold pool mechanism, sometimes the mesoscale ascent that destabilizes the column to facilitate such an updraft cell instability occurs ahead of matured MCS with discrete line (Fig. 3.2). Gravity waves (also called buoyancy waves or internal waves) are one cause for clear-air vertical motions in a stratified atmosphere, and rotational effects such as vortex secondary circulations (like QG forcing) are another. Both of these are candidates for non-cold pool mechanisms of enhancing new convective cell development in mesoscale and larger scale patches.

Mapes (1993) showed how a gravity wavefront (or “buoyancy bore”) propagates away from a top-heavy heating source ($Q$) which emulates tropical MCS. Such wavefronts or bores propagate horizontally at a speed proportional to vertical wavelength,
\[ c = \frac{\lambda}{\tau}, \quad (3.1) \]

where \( c \) is phase speed of gravity wave, \( \lambda \) is vertical wavelength, and \( \tau_b \) is buoyancy period \( 2\pi/N \) (\( N \) is Brunt-Väisälä frequency). In the tropical troposphere, \( \tau_b \) is about 10 minutes so wavenumber \( \frac{1}{2} \) of a 15 km deep troposphere propagates at a speed of about 30 km/10 min \( \sim 50 \) m/s. Such fast moving wavefronts in clear air are very hard to observe.

Two main major vertical modes are excited by the top-heavy heating profile typical of MCSs, called convective or wavenumber \( m = 1/2 \) or \( “1^{st} \) mode” (monopole \( T' \) in all vertical levels) and stratiform or wavenumber 1 or \( “2^{nd} \) mode” (a vertical dipole of \( T' \) of the troposphere).

![Figure 3.2 Fine line is gust front, and new cells are discretely triggered convective cells from the gust front. Radar reflectivity pattern in Oklahoma in June 2003. Courtesy of Dr. R. Fovell. Adapted from Houze (2004). A nice animation example is at Click Here.](image)

In Fig. 3.2, \( 1^{st} \) mode disfavors new convection because deep downward displacement stabilizes the environment. On the other hand, \( 2^{nd} \) mode creates favorable condition for new convection because upward displacement of the lower atmosphere destabilizes the
environment, which outweighs the importance of upper-level descent since convection is especially sensitive to the lower troposphere (see Chapter 4 about convection sensitivity functions). Lac et al. (2002) successfully demonstrated simulated new convection triggered by gravity wave ahead of MCS, but it is performed in strictly controlled conditions in the regional model. Generally, it is difficult to get observational or modeling evidences of gravity wave mechanism because there are typically many convective sources, like handfuls of pebbles thrown in a lake making wave fields that are very transient and complex around so many sources.

Figure 3.3 Schematic of the buoyancy bores (closed contour), horizontal wind (u1 and u2), and integrated displacements of material lines (solid line). Heating source \( Q \) is located in \( x=0 \). Adapted from Mapes (1993), which is originally adapted from Nicholls et al. (1991).

3.2 Data and Methodology

3.2.1 Model setup and data

GEOS-5 simulations examined here are from the cubed-sphere non-hydrostatic atmospheric general circulation model (Suarez et al. 2008) on very fine grids of 5-10 km (Putman and Suarez 2011). Since the quasi-uniform grid in cubed-sphere frame is suitable for mesh refinement and ideal for 2D X-Y decomposition, GEOS-5 performs the non-hydrostatic global simulation with high spatial resolution as fine as \( \approx 1.7 \) km (Putman and Lin 2007, 2009). For simulating convection explicitly, the moist scheme in GEOS-5 consists of two kinds of parameterization – “convective” and “large-scale cloud
condensation” parameterization. The “Convective” parameterization is the relaxed Arakawa-Schubert (RAS) scheme, which estimates convective mass fluxes to the idealized convective plume at each time step based on Moorthi and Suarez (1992). The “Large-scale cloud condensation” scheme (PrognoCloud) consists of two types of cloud – “anvil” and “large-scale condensation”. “Anvil” cloud is directly determined from detraining fluxes of the anvil cloud type in RAS, whereas “large-scale condensation” is independently estimated by an assumed probability distribution function (PDF) of sub-gridscale RH, so that some fraction of the grid is assumed to be saturated and therefore has latent heating and cloudiness. In Putman and Suarez’s simulation (2011), the RAS convective scheme is modified to suppress convective plumes by limiting plume entrainment (i.e. a stochastic Tokioka constraint, Tokioka et al. 1988). Using the Tokioka constraint effectively suppresses deep convection and engages the grid-scale condensation scheme (Lee et al. 2008), which makes the convection explicitly simulated.

A 5-km horizontal resolution simulation spanning 20 days from February 2 to 22 in 2010 is used in this study (Putman and Suarez 2011). While the original vertical array is 72 layers from the surface to 0.01 hPa (Rienecker 2008), 37 layers to 1 hPa are used with the following resolution: 25- to 700, 50- to 100, 30- to 70, 20- to 50, 10- to 10, 3- to 7, 2- to 5, and 1-hPa to 1 hPa.

The top 10 heavy rain events in 2.5 degree coarse-graining of the data were used as central points for extracting space-time data cubes of $6 \times 6$ degrees $\times$ 37 hours for study. Detailed methods follow in 3.2.2.
3.2.2 Methodology

Since GEOS-5 high-resolution model is simulated without lateral boundary conditions, unlike a regional cloud model, convection will emerge from interaction with its environment in freely evolving weather conditions. In this study, we focus on 10 heavy rain events, similar to the case study analysis of Mapes and Houze (1992). The 10 cases will be analyzed as a composite mean, and as individual cases, in section 3.3.1 and 3.3.2, respectively. Below we describe how the 10 cases were selected and explain the method for compositing the events as well as analyzing them as individual cases.

3.2.2.1 Case selection

Heavy rain events were selected in the following manner. The 5-km precipitation was rebinned to 80-, 160-, 250-, and 500-km sizes, then spatially averaged in each bin. The ten maximum values (“events”) in each rebinning size were selected within 15°S to 15°N during the whole simulated time period. Events in 250-km size are chosen for this study since these relate well to the size definitions of the term “MCS” (100s of km). Figure 3.4a shows the location and time of the 10 events in the GEOS-5. Events are named case 0 to 9.

Since case 0 and 3 are tropical cyclones according to their spatial patterns (not shown), they were excluded from this analysis to focus on non-TC mechanisms of convective organization in the model, so only 8 cases are analyzed. Case 1 and 4, and 2 and 5 are consecutive events of the same weather system, so they are merged, leaving us with 6 truly distinct weather systems. Cases 1 and 4 are over land, cases 6 and 9 are coastal, and the rest of cases are over ocean. For clarity in the land cases, UTC time is converted to local time.
As an observational baseline for comparison, similar regridding and peak finding were used on the Tropical Rainfall Measuring Mission (TRMM) 3B42 dataset during the same 20-day time period. The location and time of those observed 10 events are shown in Fig. 3.4b. Since there are no cyclones among the 10 TRMM 3B42 events (not shown), all 10 cases are used for comparison of composite precipitation in section 3.3.

![Figure 3.4](image)

Figure 3.4 Location and event local time ($t = 0$ hr) of 10 heavy rain events in (a) GEOS-5 and (b) TRMM 3B42. Yellow box indicates location and case number (0 to 9). Box represents $6^\circ \times 6^\circ$ size. Time format “DD HH:MM” indicates day (DD), local hour (HH), and minute (MM).

### 3.2.2.2 Method for composite study

Once events were selected, other variables occurring within $\pm 18$ hours and $\pm 3$ degrees of the precipitation event center were extracted as 250-km rebinned values. Since the time interval of rebinned precipitation (30-min) is different from other rebinned variables (3-hr), other variables are extracted around the closest time (dotted vertical line in Fig. 3.5) from the event time (solid vertical line in Fig. 3.5). The time of the other variables was used as the “central time” for compositing because the precipitation event times were not available in all the cases for the other variables. The choice for shifting the central composite time is acceptable given that precipitation values at both times (solid and dotted line) do not change much.
Figure 3.5 Precipitation in each case of GEOS-5. The x-axis is the time index of precipitation, where one index represents a 30-minute interval. Black squares are total, red xs are large-scale, green triangles are anvil, and blue pluses are convective precipitation. The solid vertical line is the maximum precipitation time and the dotted vertical line is the closest time of the other variables from maximum precipitation time. “Time diff” is the time difference of other variables to the maximum precipitation.

Temperature ($T$), specific humidity ($\textit{qv}$), and vertical velocity omega ($\omega$) were chosen for analysis of the general characteristics of heavy rain events (section 3.3.1).

3.2.2.3 Method for case study

For the detailed case studies, space-time data are analyzed in section 3.3.2. Space-time samples are extracted with a $670 \times 670 \text{ km}^2$ domain size centered at the event location. A $670 \text{ km}^2$ size is chosen to ensure MCS patterns in the vicinity of the event location are covered. Space-time samples are also extracted within $\pm 18$ hours of the event time. Note, the time interval of the spatial data is one-hour resulting in a total time-
array of 37 elements. Even though the central times in half the cases are not synchronized with the event time, the time discrepancy can be ignored owing to only ± 30-minute time differences. Precipitation at the surface (prec), temperature at 2m \((T2m)\), specific humidity at 2m \((qv2m)\), and temperature at 850 hPa \((T850)\) in spatial-time array are chosen for analysis of detailed characteristics of individual heavy rain events (section 3.3.2).

3.3 Results

All studies of composite as well as individual cases are available as an animation in http://www.rsmas.miami.edu/personal/ssong/research/HR_250kmevents.htm. Putman and Suarez (2011) also provide the movie for the whole simulated period, 20 days in the HTML, doi:10.1029/2011GL048438 (http://onlinelibrary.wiley.com/store/10.1029/2011GL048438/asset/supinfo/grl28355-sup-0002-ms01.mov?v=1&s=74c01e28116c37523b2736aa33fddbd237adc6f1).

3.3.1 Composite analysis of case study

Figure 3.6 shows the composite of precipitation for the 8 heavy rain events shown in Fig. 3.4. Convective (blue pluses) and anvil (green triangles) precipitation constitute only a small portion of the total precipitation (black square), whereas the large-scale condensation (red x) scheme explains most of the precipitation. The dominance of the large-scale condensation scheme to total precipitation confirms that deep convection is simulated explicitly with suppressed convective scheme.
Figure 3.6 Same as Fig. 3.5, but composite of 8 heavy rain events. X-axis is time centered at the event time.

Figure 3.7a shows the rainfall evolution of each GEOS-5 case, as well as the composite of the GEOS-5 and TRMM events. Rainfall evolution is plotted ±18 hours relative to the event time. The correctness of our heavy rain event selection process is confirmed by the fact that maximum precipitation occurs at time zero for each case and the maximum value decreases as the case number increases. Composites of TRMM 3B42 and GEOS-5 precipitation (thick black asterisk in Fig. 3.7a) show similar evolution patterns with time, but TRMM is almost twice as large as GEOS-5 (~ 13 mm/hr vs. ~ 6 mm/hr, respectively). TRMM 3B42 tends to overestimate light rain in the ocean and underestimate heavy rain in coastal and land areas owing to the rainfall retrieval algorithm with passive microwave measurement (Wilheit 1986, Huffman et al. 2007, Chen et al. 2013), but this could not explain the peak difference. While evaluation of precipitation in the GEOS-5 explicit run is beyond the scope of this study, the discrepancy is noted here.
Figure 3.7b shows the same 8 events as Fig. 3.7a, but according to their local time evolution. Most of the event times (i.e. peak rainfall) are concentrated on the midnight to the early morning except case 9 in Fig. 3.7b, and all 10 cases in TRMM 3B42 are on the same range as well (not shown). Night-to-early morning intensive convection is observed over the oceans (e.g., Gray and Jacobson 1977), and also can happen in coastal area due to topographic effects (e.g., Mapes et al. 2003a), various geographical conditions are explored in section 3.3.2 through individual case study.

![Comparison of GEOS-5 with TRMM 3B42](image)

![Precipitation in GEOS-5 cases](image)

Figure 3.7 (a) 250-km rebinned precipitation centered at the event time in GEOS-5. Individual cases are represented by color. Thick black squares are the composite in GEOS-5 and thick black asterisks are the composite in TRMM 3B42. (b) is same as (a), but with local time.

The composite average of other variables are also examined to analyze the general characteristics around and during MCSs. The detailed method for compositing is
explained in Appendix B. Figure 3.8 shows the time-height composite patterns of $T'$, $qv'$, and vertical velocity $\omega$, which are rebinned values over the 250-km area. Anomalies of $T$ and $qv$ ($T'$ and $qv'$) are calculated by subtracting the time mean at each vertical level.

Figure 3.8 Composite of (a) temperature [K], (b) specific humidity [g/kg] anomaly, and (c) original vertical velocity $\omega$ [hPa/s]. X-axis is time centered at the event time (t=0) [hr] and y-axis is vertical height [hPa]. All variables are rebinned values over the 250-km area.
Temperature exhibits 3 main layers of coherent variation the troposphere (Fig. 3.8a): near the surface, lower troposphere (600-900 hPa), and upper troposphere (200-600 hPa). 

$T$ is cooler during and after rain event, especially at low levels where rain-cooled air descends and accumulates near the surface. Most of this post-rain tends to be during the daytime according to Fig. 3.7b. One interesting precursor signal to maximum rainfall is a cold tongue near 700-900 hPa which emerges about 3-4 hours before the cold pool at the surface. The cold tongue reduces the convective inhibition (CIN) by reducing the temperature difference between the convective parcel and its environment, making the environment favorable for convection (Mapes 2000). The cold tongue’s reduction in parcel to environment temperature difference is consistent with the “stratiform instability” which is one possible lower free troposphere control mechanism by temperature for organizing convection (Fig. 5 in Mapes 2000).

Humidity also varies in 3 main coherent layers in Fig. 3.8b: roughly 850-950, 600-850, and 400-600 hPa. Moist anomalies deepen from the lower to middle layer about 7-8 hours before the event time. This moistening tendency is consistent with the notion of reduced inhibition of cumulus clouds with the appearance of the 800 hPa cold tongue. This humid-cool tendency deepens once more from the middle to the upper layer about 3 hours before the event time at 650 to 550 hPa. Zipser (1977) showed that the role of a dry-warm layer in an onion sounding is to inhibit deep convection via the vertical advection by adiabatic descent in MCSs. Although the mechanism behind a dry tongue (Mapes and Zuidema 1996) is different from an onion sounding, the dry tongue shows the same conditions as a dry-warm layer for inhibiting deep convection. The humid-cool
layer in Fig. 3.8a and b, therefore, is the evidence for enhancing deep convection via the vertical advection by adiabatic ascent in contrast to the dry-warm layer.

The 3-layer pattern with two time jumps is broadly consistent with patterns in observation-derived regression composites\(^2\) in Mapes et al. (2006). The 3-layer pattern can be interpreted in terms of three MCSs development stages, which is based on three types of tropical cumulus clouds – shallow cumulus, congestus, and cumulonimbus – with trimodal characteristics of divergence, cloud detrainment, and fractional cloudiness (Johnson et al. 1999). In addition, this 3-layer pattern is stretched over various convective space and time scales, which Mapes et al. (2006) refers to as the “stretched building block” hypothesis. This multiscale variability is modulated by large-scale waves.

Vertical motion \((\omega)\), shown as original values instead of anomalies, has a smoother tilted shape, which indicates deepening with time (Fig. 3.8c). Unlike observations, \(\omega\) is accurately measured and known in the model data, which is a great advantage of models for studying the interaction between convection and the large-scale environment. However, \(\omega\) is still not easy to interpret because it contains both diabatic and adiabatic parts. While a large “diabatic” part balances the strong latent and other heating processes in MCSs (Mapes and Houze 1995), a small residual or “adiabatic” component causes the dynamically-crucial stratification adjustments that may govern the dynamical evolution of typical life-stages in MCSs, akin to geostrophic and ageostrophic flow decomposition in quasi-geostrophic theory (p.497-499, Gill 1982).

These general features of the composites are all consistent with previous observational studies (Frank 1978, Sherwood and Wahrlich 1999, Mapes et al. 2006),

\(^2\) Regression is simply an anomaly-weighted composite (see Appendix B).
where only composites can be examined due to data weaknesses. However, high-resolution model data in GEOS-5 has the advantage for individual case study analysis with an accurate and complete data set. In section 3.3.2, we shall see that the composite characteristics do not occur in all cases, rather diverse mesoscale dynamics cause individual extreme rain cases.

3.3.2 Individual analyses of case studies

The six distinct model case studies are now presented to illustrate the diverse dynamical processes operating to organize the mesoscale. Cases where MCSs developed from obvious external forcings are considered first, followed by the subtler cases where MCSs may be more self-organizing. We begin in section 3.3.2.1 with case 6 over western Colombia (Fig. 3.4a), where geographical surface contrasts are a strong source of forcing with diurnal timescales (Mapes et al. 2003a, b). Next, section 3.3.2.2 examines forced-convection cases involving extratropical interaction – upper-level midlatitude troughs impacting low latitudes. Weaker forcing cases over ocean and land complete this survey in section 3.3.2.3 to 3.3.2.6.

3.3.2.1 Case 6 – Coastal mountain convection

Case 6 is located in the Panama Bight, a coastal area surrounded by Panama and Colombia, west of the Andes (Fig. 3.4a). Figure 3.9 illustrates the weather evolution of 2-meter temperature ($T_{2m}$, top), humidity ($qv_{2m}$, middle), and enthalpy ($C_pT + Lqv$, bottom) at the surface from -6 hours to the event time at 2-hr interval (from left to right). Since $z$ is small and nearly constant, the enthalpy is nearly equal to moist static energy ($MSE = C_pT + Lqv + gz$) in most cases, but not over the mountains whose surface altitude we do not have on the same grid. Still the acronym MSE is used for convenience.
Around midnight ($t = \text{-6 hr}$), small convective cells emerged from the Andes and propagated offshore to the west, where they merged to form a mesoscale system in the early morning. Near-surface temperature and humidity indicate that convection favors highly humid (high MSE) areas offshore (middle row in Fig. 3.9). Once the MCS is formed, it tends to stay and follow the edge of temperature gradient (top row in Fig. 3.9), but this is a highly humid area, too (middle row in Fig. 3.9). In the bottom row in Fig. 3.9, MCSs developed and propagated while consuming high MSE values, and dissipated as high MSE values are flattened. This MSE patterns also resembles more humid areas than temperature (middle and bottom rows in Fig. 3.9). Consuming high MSE (the fuel for buoyancy in deep convection) is a common mechanism of convective organization, on very large scales too (Fig. 9 in Gallée et al. 2004).

![Figure 3.9 Color shading: Temperature anomalies (top), humidity anomalies (middle), and 2m enthalpy (bottom row) in case 6. Blue-black shows the Andes mountains. Local time 1:30 ($t = \text{-6}$)](image)
3:30 (t = -4), 5:30 (t = -2) and 7:30 am (t = 0) from left to right. All of them are surface data. Black contour: precipitation.

Coincidently, there is an enhanced case study with observations and modeling analyses about MCSs’ propagation in the same location as case 6 (Mapes et al. 2003a, b). In the Mapes et al. (2003b) case study, gravity waves forced by the diurnal cycle of land heating was discussed as the propagation mechanism (their Fig. 7a). As the cold anomaly patterns at 800 hPa propagated offshore during the night and into the morning, precipitation responded to the cold temperature anomaly (Fig. 6a and 7a in Mapes et al. 2003b). By comparison, temperature anomalies at 850 hPa in case 6 are averaged over the same latitude belt, 5°N – 7°N, and plotted in the longitude-time dimensions in Fig. 3.10. The 24-hour mean is removed as the daily mean (daily time-mean window, from -12 to + 11 hours). In addition, only the oceanic part (west of the 283°(= 77°W)) is used for calculating the mean as Mapes et al. (2003b) did.
Figure 3.10 Latitudinally averaged (5-7°N) temperature anomaly patterns with time at 800 hPa by model (Mapes et al. 2003b) and (b) at 850 hPa in case 6 of GEOS-5. (a) Black contours are temperature anomalies, (b) color shading is temperature anomalies and the black contours are precipitation. The grey line is the propagation speed of temperature in GEOS-5. The red line is adopted from (a) Mapes et al. (2003b), ~15 m s⁻¹.

Mapes et al. (2003b, Fig. 7a) showed cool daytime and warm nighttime propagation of temperature anomalies at 850 hPa over the ocean. In contrast, case 6 has warm daytime and cool nighttime temperatures over the ocean with little evidence of cool wave propagation (sloping feature) in Fig. 3.10. Of course, one specific day need not agree with a 10-day mean and the two models may behave differently. Nevertheless, case 6 shows similar propagation patterns between precipitation and temperature, like Mapes et al. (2003b). Despite having opposite temperature anomaly signs during the local day time
in the ocean, a diurnal cycle is clear in both case 6 and Mapes et al. (2003b). Figure 3.7b shows this diurnal cycle of precipitation. Only coastal cases, including the present case 6 (light-blue line) and case 9 (purple line) in Fig. 3.7b, show cyclical patterns instead of bell shapes with time.

The other noticeable difference between our case 6 and Mapes et al. (2003b) is the propagation speed. Case 6 propagates at \( \sim 6 \text{ m s}^{-1} \) (grey slope in Fig. 3.10) compared with \( \sim 15 \text{ m s}^{-1} \) (red slope in Fig. 3.10) in Mapes et al. (2003b). The difference may depend on speed or direction of background flows. According to the humidity patterns in Fig. 3.9 (middle row), there is a continuous background flow from north to south (https://www.rsmas.miami.edu/users/ssong/research/HR_movies/250events_case6_var.mov). This flow meets gravitational propagation from inshore and the two airmasses collide along the coastal line. Scattered convective cells from the mountain, therefore, appear to stay and organize as persistent MCSs near the coastal line. Perhaps these dynamics, rooted in low-level advective processes, lead to the slower propagation speed in case 6 than the case of Mapes et al. (2003b), which shows the dominant sea breeze from west to east.

3.3.2.2 Case 2 & 5 – Synoptic scale midlatitude trough intruding to 15°N over the eastern Pacific

Case 2 and 5 are consecutive cases affected by large-scale dynamics over the ocean at the subtropical edge of our latitude belt (15°S to 15°N). In these cases, precipitation occurred with strong background flow, horizontal shear in the cyclonic sense and vertical shear implied by thermal wind balance over a strong mean northwest-southeast horizontal temperature gradient seen at both surface and 850 hPa during the whole evolution period.
Figure 3.11 shows temperature (color fill) and precipitation (black contours) patterns with time at the surface ($T_{2m}$, upper) and 850 hPa ($T_{850}$, lower panels). Since case 2 precedes case 5 by only 6 hours, the analysis focuses only on case 5 in rest of this section. A subtropical trough to the west (implied by black to blue shading) carries precipitation northeastward along its edge.

Figure 3.11 Temperature at 2 m ($T_{2m}$) (upper panels) and 850 hPa ($T_{850}$) (lower panels) in case 5 are colored, and precipitation is black contour. Each column is drawn at -6, -4, -2, 0, and +2 hours from left to right. Colorbar range of $T_{2m}$ is different from $T_{850}$, but its interval is same.

Animation makes clear that the westward motion is advection in the mean easterly flow at low levels, while the large-scale eastward motion is presumably from the advection of vorticity in a trough aloft by upper-level westerlies, whose existence is implied by thermal wind balance in the strong north-south $T$ gradient (https://www.rsmas.miami.edu/users/ssong/research/HR_movies/250events_case5_var.mov). In the quasi-geostrophic (QG) theory, upward motion associated with the positive vorticity advection (PVA) can be explained by the QG omega equation below,

$$\left(\nabla^2 + \frac{f_0^2}{\sigma} \frac{\partial^2}{\partial p^2}\right)\omega = -\frac{f_0}{\sigma} \frac{\partial}{\partial p} \left[ -v_g \cdot \nabla_p \left( \zeta_g + f \right) \right] - \frac{R}{\sigma p} \nabla^2_p \left( -v_g \cdot \nabla_p T \right),$$  

(3.2)
where $\sigma$ is static stability, $\zeta_g$ is vorticity, $f$ is $f_0 + \beta y$, and $\beta$ is the constant meridional gradient in the Coriolis parameter $\partial f / \partial y$. According to Eq. 3.2, “positive” vorticity advection (PVA) at upper-levels (first term on the right in Eq. 3.2) will induce strong upward motion (i.e. “negative” $\omega$) ahead of a trough axis (left term in Eq. 3.2), where PVA is maximum as Fig. 3.12 shows. Then, strong cold advection from this midlatitude trough will be coupled with the development of a low-level frontal boundary and interact with tropical convection (Kiladis and Weickmann 1992). Case 2 and 5 shows this interaction between tropical convection and midlatitude synoptic weather systems as well.

![Figure 3.12](image)

Figure 3.12 (a) Schematic diagram of vorticity advection along the ridge and trough in upper level. It is adapted from the class note at department of Earth & Space Sciences in University of Washington by Gerard Roe. (b) Temperature patterns at 850 hPa (colored) with precipitation (black contour) at 4 hours before the event time in case 5. Pink line indicates midlatitude trough.

Low $q_{2m}$ areas (low MSE) can have a critical effect to weaken convection. The later stages of the case are shown in Fig. 3.13. Gradual weakening of the trough in the thermal structure is seen, and then Precipitation is suddenly weakened as dry areas in the tropics (green closed area in upper panels in Fig. 3.13) approach the convection from the southeast.
Figure 3.13 MSE (upper panels) and humidity (lower panels) at the surface in late stages of case 5 is colored, while precipitation is indicated by black contours. Panels show -1, +1, +4, +8, +13 hours from left to right.

3.3.2.3 Case 7 – Confluence along the latitude 4°N in the ocean

Case 7 is characterized by confluence along 4°N during the whole development period. While the main convection emerged along the 4°N latitudinal line (black oval, upper left panel), secondary convection was triggered southwestward after the event time (black oval on the upper right panel, Fig. 3.14). This secondary propagation developed via discrete development of smaller convective cells, not continuous propagation.

Figure 3.14 MSE (upper) and temperature at 850 hPa (lower panels) in case 7 are colored, and precipitation is in black contours. Each column is drawn at local time 2:30 ($t = -2$), 4:30 ($t = 0$),
6:30 (t = 2) and 8:30 am (t = 4) from left to right. Two white vertical lines indicate the averaged area in Fig. 3.15.

In the previous cases (i.e. case 2, 5, and 6), convection propagated toward high MSE areas. In case 7, however, convection does not propagate to the high MSE region (i.e. north of 4°N; upper panels in Fig. 3.14), but along the confluence line 4°N latitude. Examination of the animations suggests that dynamical conditions (large-scale flow) may be more important than thermodynamic conditions (i.e. near-surface MSE) (https://www.rsmas.miami.edu/users/ssong/research/HR_movies/250events_case7_var.mov). Since the location of case 7 is in the ITCZ, any dynamical mechanisms related to the ITCZ may affect the horizontal confluence around 4°N.

Secondary convective propagation is not along the 4°N confluence line, like the main convective activity, but to the south. Reduced inhibition by cooling above the boundary layer (i.e. a cold tongue) might be responsible, as in the multi-case composite in Fig. 3.8a. Lower panels in Fig. 3.14 show T at 850 hPa, where a cold anomaly pattern extends from the organized convective systems toward the south in advance of convective development. This propagation pattern is shown in Fig. 3.15 as a time-latitude pattern averaged over 230° to 231° longitudes (between two while vertical lines in Fig. 3.14). Several longitude windows were tested, but results did not vary as there was little change in propagation speed. Cold anomalies at 850 hPa seem to radiate from the convective signals in both the north and south directions in Fig. 3.15 with various speeds from ~2.5 to ~10 ms\(^{-1}\). The various speeds may be due to complex reasons including combined background flow and various wave speeds. Even though dry (without precipitation, black arrows) or moist (with precipitation signal, red arrows) conditions are not the only effect to change propagation speed, propagation of temperature anomalies become slower when
precipitation is involved, e.g., 9.9 (black solid) to 6.8 m\(^{-1}\) (red solid), and 11.3 (black dashed) to 2.5 m\(^{-1}\) (red dashed). Southward propagation of a warm anomaly (black solid on the warm anomaly) precedes the dominant cold anomaly (black solid on the cold anomaly) propagation with similar contour slopes, and is perhaps emanating from the decay of a previous system (i.e., from the termination of a cooling process which is an anomalous heating process with sinusoidal fluctuation in the forced gravity wave equation).

![Figure 3.15 Time-latitude pattern of temperature anomaly averaged in 230-231 longitude areas. Figure is flipped right and left for intrinsic reason because convective propagation is westward with time. Propagating speed is 9.9 (black solid), 6.8 (red solid), 11.3 (black dashed), and 2.5 m s\(^{-1}\) (red dashed).](image)

New convective cells from the main convection propagate with jumping tendencies ahead of the previous convective cells in Fig. 3.14. These new convective cells are not
caused by cold pools at the edge of the MCSs. Rather, the new cell’s distance from the parent cell indicates discrete propagation. There are observations and hypotheses of discrete propagation instead of cold pool mechanism including radar reflectivity observations (Fig. 27 in Houze 2004), GATE analyses (Houze 1977), larger-scale lifting mechanism for MCS (Crook and Moncrieff 1988), the interaction between the sea-breeze front (SBF) and front-parallel horizontal convective rolls (HCRs) (Fovell 2005), and etc. One of the noticeable mechanisms for discrete propagation is destabilization by gravity wave propagation.

Deep convection acts as a source for gravity waves in many ways – diabatic heating effects, obstacle-in-shear effects, vertically oscillating effects, and other nonlinear advection processes. First, diabatic heating effects regard deep convection as a heating source, which induces instability thermally and generates gravity waves (Bretherton 1988, Lin et al. 1998, Chun and Baik 1998). Since Schmidt and Cotton (1990) showed two dominant gravity wave modes propagated from a model-simulated MCS, diabatic heating from MCSs have been used as a gravity wave source in many subsequent studies (Nicholls et al. 1991; Mapes 1993, Mapes and Houze 1995, Mapes 2000, Lac et al. 2002, Tulich and Mapes 2010, etc.). Secondly, in the presence of shear clouds can act as obstacles and interrupt air flow and become sources of gravity waves (Clark et al. 1986). Thirdly, the oscillating convective updraft hypothesis proposes the generation of gravity waves via mechanical forcing as deep individual convective updrafts oscillate when they hit the tropopause (Pierce and Coroniti 1966, Fovell et al. 1992, Lane et al. 2001, Fovell et al. 2006). Most generally, the nonlinear advection term can contain additional forcings
for gravity waves in deep convection (Lane et al. 2001), with oscillator and obstacle effects being special cases.

Since only a limited data set is available in this study ($T$, $qv$, $\omega$, and precipitation at only surface and 850 hPa), the above mechanisms could not be tested or isolated in detail, but Fig. 3.15 makes clear that propagating T850 signals were present and may have helped shape the convective development pattern.

3.3.2.4 Case 8 – No strong wind shear over the ocean

Organized mesoscale convection in case 8 formed at midnight over the ocean without strong background flow, then propagated northeastward and dissipated. Figure 3.16 shows the evolution of convection along with the MSE pattern. Before the event time (February 4th in 2010 at 23:30 LT), several small convective cells “jumped” into the high MSE region from the southwest. These cells merged and propagated to the northeast as an organized line shape about 4 hours (from 4th 22:30 to 5th 02:30 on February 2010), then became scattered, although still broadly organized in a NW-SE line until the end of this event. As in the previous cases, convection preferentially consumes high MSE, and the humidity pattern sets the MSE pattern (not shown).
As convection passes, surface cooling is seen in close association with precipitation, indicative of precipitation-driven downdrafts. Temperature at 850 hPa keeps a relatively similar broad cool temperature pattern (Fig. 3.17), moving broadly eastward like the mesoscale pattern of convection. Might the cold temperature at 850 hPa be shaping the mesoscale convection pattern via reduction of CIN in a cold tongue near 700-900 hPa as seen in the composite mean (Fig. 3.8a). The cold tongue emerged about 7-8 hours before the event time and disappeared about 5-6 hours after it in the composite (Fig. 3.8a), whereas the cold areas ahead of convection (i.e. spatial cold tongue) at 850 hPa did not disappear ahead of the new convective cells (lower panels at time +7 and +9 in Fig. 3.17).
Figure 3.17 Temperature at 2 m (upper panels) and 850 hPa (lower panels) at -9, -7, -5, -3, -1 hours from the event time (left to right).

Figure 3.17 (Continued) at +1, +3, +5, +7, and +9 hours from the event time (left to right).

To see the difference in propagation speed around the event time clearly, space-time patterns of temperature anomalies at the surface and 850 hPa are shown in Fig. 3.18.³ Before convection was organized (negative time index), smaller convective cells moved eastward at a moderate speed ~2.8 ms⁻¹ in conjunction with temperature anomalies at the surface. This eastward motion is the prevailing motion of the near-surface flow as

³ Since convection is propagated across the northwest-southeast (NW-SE) axis of last panel in Fig. 3.16, this domain is rotated 45° anticlockwise before calculation of Fig. 3.16. Since 200 km in latitude around the center of domain are averaged, the warm flow pattern before the event time represents near-surface flow (white arrow, ~2.8 ms⁻¹), and the cold flow pattern with precipitation after the event time represents cold pool surges (yellow arrow, ~5.3 ms⁻¹) in Fig. 3.18.
indicated by the white arrow in Fig. 3.18a. Once convection is organized, the cold pool surges eastward at a faster speed, \( \sim 5.3 \text{ m s}^{-1} \), while precipitation continues at its front edge. These cold pool surges at the surface are consistent with cooling tendencies at 850 hPa as indicated by the yellow arrow in both Fig. 3.18a and b. (https://www.rsmas.miami.edu/users/ssong/research/HR_movies/250events_case8_var.movie).

3.3.2.5 Case 1 & 4 – Amazon basin

Case 1 and 4 are consecutive events over the Amazon. These cases are located in northwest Amazonia \((286^\circ-292^\circ, 6.5^\circ S-2^\circ N)\), where the basin is surrounded by Andes Mountains along the west coast. The Amazon River crosses east to west and smaller
rivers branch off from it - Jupurá, Putumayo, and Napo in Fig. 3.19. The size of Amazon river is huge, ~10-50 km width (as dry and wet seasons) and ~6400 km length. This geographical characteristic – a basin with abundant moistures supported by huge rivers – would be the reason why land area can be selected as one of the 8 heavy rain events.

Figure 3.19 (Left) Domain of case 4. (Right) Red box is same as left panel. (adapted from Wikipedia).

Since case 1 precedes case 4 by only 4-hours, analysis is focused on case 4. Convection developed during the nighttime and peaked in the early morning, as is clear from the diurnal evolution of surface $T$ patterns in Fig. 3.20. As a precondition for convection, nocturnal conditions are often considered unfavorable owing to their cold and stable boundary layer, but in case 4 a major organized convective event happened under such condition.
Figure 3.20 shows that a breeze developed toward the rivers (which are slightly lower and warmer than the surrounding land), resulting in air mass convergence and the development of new convective cells (Feb. 4$^{th}$ 23:30 to 5$^{th}$ 3:30 in Fig. 3.20). Individual convection gathered and organized around the 5$^{th}$ at 3:30 and peaked in the early morning (i.e. 7:30, which is event time). Once organized mesoscale convection emerged, it propagated southwestward with a line shape because of the dominant northeasterly in this region, and finally weakens and exits the domain of the figures.

Figure 3.20 Temperature pattern at 2m (color shading) in case 4. Black contour is precipitation. All times are local times, as the diurnal temperature cycle makes clear. Time is from 4$^{th}$, 21:30 to 5$^{th}$ 19:30. 7:30 is the event time.

From left to right columns in Fig. 3.21, temperature at 2m in front of MCS (west-south corner area in the first row) is getting warmer while temperature at 850 hPa (west-south corner area in the second row) is getting cooler. This shows the stable inversion layer is broken as sun rises. At the same time, favorable conditions for high MSE are
formed in southern areas with higher temperature and moisture at the surface (first, third, and last rows are $T_{2m}$, $qv_{2m}$, and MSE respectively in Fig. 3.21). Unlike case 2 & 5 or case 8, MCS in case 1 & 4 is not formed by favorable thermodynamical condition (i.e., dominantly high MSE), but dynamically enforced condition (i.e., land-river breeze and convergence with moisture supplied from rivers at night). Higher MSE is formed after the event time (5th 7:30, second column in Fig. 3.21) and widely spread out on the south area. That is not continued to maintain the MCS, but spreads out it as smaller pieces (last column in Fig. 3.21).

Figure 3.21 Temperature at 2m (first row) and 850 hPa (second row), humidity at 2m (third row), and MSE at 2m (the last row) in case 4. Black contour is precipitation. All time are local time – 5th 4:30, 7:30 (event time), 10:30 and 13:30 from left to right.
Case 1 & 4 is consistent with the study about the diurnal cycle of precipitation in Amazon basin in Tanaka et al. (2014). They explained the peak of MCS in the early morning that “the river surface is warmer than the continental surface, and hence upward movement over the water surface and subsidence over the continent are present. That is, convective clouds develop over the river surface in the early morning hours. As the morning hours progress, the predominant equatorial easterly flow over the region (Fig. 3 in their paper) drifts the convective clouds westward and at 05 LT”. They concluded that “organized and small-scale convections are responsible for rain events in the Amazon Basin” and “local effects like the river breeze control the diurnal cycle of stations closer to the river. (Tanaka et al. 2014)”

3.3.2.6 Case 9 – Peak at the afternoon near the coast in the ocean

Typically, convection in maritime area shows the peak at the afternoon in land while the peak in the early morning in the ocean (Fig. 3 in Sato et al. 2009, Fig.1 in Cronin 2012). Although thermal land-sea breeze is strong enough to trigger convection in land at the afternoon, it is not in the ocean in the early morning. Gravity wave effect by coastal mountains during the night is one of the major causes for the latter one (Fig. 11 in Mapes et al. 2003b).

Although case 9 happens near the coast in the ocean, this is opposite to the typical convection as above – peak convection at the afternoon (event time = 16:30) in the ocean. While southwesterly sea breeze continually blows at this time, convection is somewhat stagnant in the middle of the domain, and smaller convective cells eventually merge to form MCS (black contour in Fig. 3.22). This is caused by counter flow to the southwesterly sea breeze, which is the dominant northeasterly wind in Panama Bight.
According to the Fig. 3.23 (adapted by Xie et al. 2008), “during the boreal winter (January to April) when the ITCZ is south of the region, the pressure gradient from the Caribbean to the Pacific is strong and sustains strong northeasterly low-level winds over Panama which are intensified through the mountainous terrain (Cheng et al. 2011).” Therefore, neither higher MSE nor convection cannot move toward the land because of northeasterly winds, even convection cannot move backward to consume higher MSE (redish colors in last row in Fig. 3.22) because of southwesterly sea breeze. That is why the convection converges along some steady line and organizes MCS in the ocean at the afternoon other than the typical land-sea breeze effect in maritime area.

Figure 3.22 Temperature at 2m (first row), humidity (second row), and MSE at 2m (the last row) in case 9. Black contour is precipitation. All time are local time – 6th 10:30, 13:30, 16:30 (event time), 19:30 and 22:30 from left to right.
Figure 3.23 Climatology of surface wind velocity (m s\(^{-1}\)), SST (black contours, °C), and precipitation (grey shade and white contours, mm month\(^{-1}\)) for February, based on Quick Scatterometer (QuikSCAT), Advanced Very High Resolution Radiometer (AVHRR), and Tropical Rain Measuring Mission (TRMM) satellite observations, respectively. Adapted by Fig. 1 in Xie et al. (2008). Red box is the area in case 9.

3.4 Discussion and summary

This chapter has shown a composite (section 3.3.1) and a diversity of mechanisms in the GEOS-5 high-resolution model for organizing convection on the 2.5-degree scale (section 3.3.2). These mechanisms include: lower boundary forcings (Amazon river and Panama Bight, case 1 and 4), the ITCZ and its mean confluence (concentrated along lines of longitude, case 7), intrusion of midlatitude troughs and their quasi-geostrophic lifting into the latitudes near 15°N (case 2 and 5), gravity waves lifting the “cap” of inhibition near 850 hPa (case 8), and coastal mountain diurnal convection (case 6). Such diverse mechanisms for MCS development are realistic, in that all are observed in nature. The hardest to observe has been fast-moving gravity waves above the surface, where clear-air observations of high frequency are rare. The International H2O Project (IHOP-2002) showed several good example of bore propagation. IHOP-2002 gathered observation data
from many instruments – e.g., spaced-antenna radar, FM-CW radar, S-band radar for deriving refractivity fields, interferometer, Raman lidar, Doppler lidar, aerosol backscatter lidar, and a surface mesonetwork – for improving our understanding about water vapor field (Koch et al. 2003a, b), and the very recent PECAN-2015 with many lidars has observed many more such instances of low-level gravity wave effects this summer, along with many other mechanism and phenomena.

Compositing is a good method to see the general characteristics of interesting atmospheric features, i.e. MCSs in this study, especially for observational data sets with the inevitable limitation of incomplete time or space sampling. Composite analyses of MCSs with GEOS-5 data showed good agreement with observational composite studies in section 3.3.1, which is encouraging. However, not every case has all the composite characteristics, on account of diverse forcing mechanisms. For example, in the composite, a cold tongue near 850 hPa slightly leads cold pools at the surface about 2 hours ahead of the peak rain (Fig. 3.8a). One might infer that the cold tongue near 850 hPa makes a favorable condition for convection by reducing CIN, while a cold pool at the surface supports a triggering mechanism of new convection by lifting boundary layer air to its (possibly wave-lowered) LFC. However, these conditions do not take place all together on each case in section 3.3.2. Case 1 and 6 (geographical forcing cases in Fig. 3.24a and b) do not have significant cold tongues near 850 hPa ahead of cold pool at the surface, and case 8 (gravitational wave-like propagation in Fig. 3. 24c) has a very weak cold pool ahead of event time compared with the composite $T'$ pattern in Fig. 3.8a. Therefore, we may get general characteristics of MCS via composite analysis, but should not assume that all the results are applicable to each case.
Figure 3.24 Time-height temperature anomaly ($T'$) in (a) case 1, (b) case 6, and (c) case 8.

One lesson of this chapter is that MCSs are linked to larger-scale flow gradients spanning the 6-degree scale of our whole regions, and thus could not be reproducible in doubly-periodic CRMs of even 600 km scale. Traditional periodic CRM simulations are thus impoverished in their mesoscale mechanisms up to now, but global CRMs like GEOS-5 will open the door to a fuller view of how the mesoscale fits into larger scales.
In summary, the GEOS-5 high-resolution global model is realistic in the sense that it reproduces many mesoscale mechanisms. The composite temperature, humidity, and vertical velocity fields of 8 heavy rain events show general characteristics of MCS consistently compared with previous studies (Frank 1978, Sherwood and Wahrlich 1999, Mapes et al. 2006). The 3-layer patterns with two time jumps in both temperature and humidity are consistent with previous study as well (Mapes et al. 2006), indicative of successfully simulating three types of tropical cumulus clouds as shallow cumulus, congestus, and cumulonimbus in the tropics (Johnson et al. 1999).
Chapter 4: A new approach to convective tendencies in a linear framework

4.1 Background

4.1.1 Linearity

The linear concept is simply Calculus: all differentiable curves are locally linear. This concept is applied to the sensitivity of a steadily convecting column of the atmosphere to the spatial-mean vertical profiles of environmental (overbar = large-scale) variables, temperature $T$ and specific humidity $q$, ($\overline{T}(z)$ and $\overline{q}(z)$). If $\overline{T}$ and $\overline{q}$ have a small perturbation like from advection, what is the response of convection? The answer will finally help to improve cumulus parameterization as well as our knowledge of convection, and it is the motivation of this chapter. Convective responses are usually evaluated by precipitation ($P$), convective heating ($Q'_{ic}$) and moistening ($Q'_{2c}$) tendencies on some larger scale average or a hypothetical Reynolds ensemble average (the expectation value).

The linear relationship between convection and its environment is shown in several CRM studies. Mapes (2004) examined sensitivities of convection to the large-scale stimuli in the CRM with 1600-km 2D domain (Grabowski et al. 1996). The horizontally uniform vertical profiles (cooling and moistening) on the strong MCS day during GATE campaign were used as the background forcing and it is decomposed as two vertical modes (deep and dipole modes). Constant radiative cooling value was added for avoiding complications of cloud-radiative feedbacks ($Q_R$). On this background forcing, large-scale stimuli (= horizontally uniform perturbation) were added during 12 hours with 10-min interval, and $P'$ responses were monitored. As the results, $P'$ responses from the $\pm$ forcing with $T'$ and $q'$ vertical mode were opposite-signed, but similar magnitude in both
vertical mode cases. It indicates the quasi-linear convective responses from the environmental perturbation forcing.

Tulich and Mapes (2010) advanced this work with more kinds of perturbations in 3D CRM, the System for Atmospheric Modeling (SAM) version 6.2 (detailed in Khairoutdinov and Randall 2003). Its domain size is 128 km with 2-km interval, and vertical height is 28 km with 50- to 500-m interval. It had periodic horizontal boundaries and time-independent background forcing, spatially uniform profiles of cooling and moistening by TOGA-COARE. They applied $T'$ or $q'$ vertical forcing separately as well as together, and compared $P'$ responses between them. Results showed that the sum of the $P'$ responses from the separated $T'$ and $q'$ forcing was almost same as the response from the both forcing together, indicating that linearity holds. Therefore, linear algebra is the appropriate approach to capture their relationship.

4.1.2 Heat and moisture budgets

Heat and moisture budgets are used to define convective tendencies in this study. Based on Yanai et al. (1973), conservations of dry static energy and moisture are defined as below,

$$\frac{\partial \bar{s}}{\partial t} + \nabla \cdot \bar{s} \bar{V} + \frac{\partial \bar{s} \omega}{\partial p} = Q_R + L(c - e), \quad (4.1)$$

$$\frac{\partial \bar{q}}{\partial t} + \nabla \cdot \bar{q} \bar{V} + \frac{\partial \bar{q} \omega}{\partial p} = (e - c), \quad (4.2)$$

where $s \equiv c_pT + gz$ (dry static energy). $Q_R$ is the radiation heating rate, $c$ is the condensation rate, $e$ is the evaporation rate of cloud droplets, $c_p$ is specific heat for air, $L$ is latent heat for vaporization, $g$ is gravitational constant, and $z$ is the height. If horizontal
mean (overbar) and its perturbation (prime) are applied, and smaller term (i.e., horizontal transport term) is ignored, then

\[
Q_i = \frac{\partial \overline{s}}{\partial t} + \mathbf{V} \cdot \nabla \overline{s} = Q_R + L(c - e) - \frac{\partial \overline{s}' \omega'}{\partial p} = Q_R + Q_{ic}, \tag{4.3}
\]

\[
Q_2 = -L \left( \frac{\partial q}{\partial t} + \mathbf{V} \cdot \nabla q + \omega \frac{\partial q}{\partial p} \right) = L(c - e) + L \frac{\partial q'}{\partial p} = Q_{2c}, \tag{4.4}
\]

where \(Q_i\) is defined as the apparent heating, and \(Q_2\) as the apparent moisture sink tendencies. “Apparent” means heating and drying (moisture sink) as it appears in a diagnosis from large-scale (averaged) fields, where averaging (overbar) is applied before multiplication to produce the apparent advective tendencies on the left hand side of the equation. Subscript \(R\) is radiative, \(C\) is convective. To calculate \(Q_i\) and \(Q_2\) in a cyclic boundary CRM, \(Q_R\) and large-scale apparent advection are specified to be uniform over the domain, and the CRM produces the domain-averaged time changes, the \((c - e)\), and eddy flux divergence terms.

Following Yanai et al. (1973), it is customary to express both \(Q_i\) and \(Q_2\) in K/day units using \(L\) and \(c_p\) appropriately. Subscript \(C\) for convective is defined \(Q_{ic} = Q_i - Q_R\) and \(Q_{2c} = Q_2\) in this study\(^4\), and prime will be redefined below as perturbations in time relative to a RCE basic state, rather than sub-CRM-scale horizontal perturbations as above. It will be revisited in section 4.2.1.

**4.1.3 Linear response matrix, M**

While Tulich and Mapes (2010) examined responses from the forcing applied over only 2 thick layers (mid- and low-level), Kuang (2010) calculated it at each of a model’s

\(^4\) Radiation is disabled and replaced with constant cooling in Radiative-Convective Equilibrium (RCE) model of Kuang’s studies (2010, 2012); \(Q_R\) is part of the forcing of the horizontally-periodic CRM and we focus on the model’s response to that forcing.
vertical levels (75 m to 1 km intervals until the top height 32 km, Kuang, 2008a, 2010, 2012) in CRM. Based on the quasi-linearity of convective response from the environmental perturbation forcing, the CRM’s response can be expressed as a linear matrix $\mathbf{M}$ below,

$$\frac{d\mathbf{x}}{dt} = \mathbf{Mx},$$  \hspace{1cm} (4.5)

where $d\mathbf{x}/dt$ is anomalous convective tendency ($Q’_{1c}$ and $Q’_{2c}$ as convective heating and moistening profiles), and $\mathbf{x}$ is anomalous state vector ($T’$ and $q’$ profiles which represent large-scale environmental perturbation). Therefore, $\mathbf{M}$ can be named as linear response function (or linear response matrix), which shows linear convective responses from the small perturbation of environmental forcing. For clarity below, we often recast $\mathbf{M}$ in energy units using $c_p$ and $L$ factors.

Conceptually, $\mathbf{M}$ is derived like below in CRM with periodic lateral boundary conditions. In each of $2N$ experiments where $N$ is number of levels in the troposphere, Kuang forced the CRM to make small localized $d\mathbf{x}/dt$ perturbations (= anomalies) at one level vertical, by putting opposite signed perturbations on the forcing profiles (representing advection terms and $Q_R$) and running for a long time until the convective tendency matches the new forcing profiles, in equilibrium state. In each equilibrium run with a different local anomaly on the $T$ or $q$ forcing profiles, the corresponding domain-averaged anomalies $T’$ and $q’$ profiles (and thus $\mathbf{x}$) can be obtained. After this process was repeated for all $2N$ cases with a vertically localized bump on the forcing profile of the $T$ and $q$ equations at each vertical level, the complete set of corresponding pairs between $\mathbf{x}$ and $d\mathbf{x}/dt$ were obtained. The matrix version of Eq. 4.5 is

$$\mathbf{Y} = \mathbf{MX},$$  \hspace{1cm} (4.6)
where matrix $X$ and $Y$ are sets of column vectors of $x$ and $dx/dt$ respectively.

According to the Kuang (2010), based on the assumption that “the domain-averaged temperature and moisture profiles completely describe the system” in Eq. 4.5, “the cumulus ensemble can be considered in statistical equilibrium with its large-scale environment at all times.” However, if the small perturbation of $T$ and $q$ is added in CSRM at a level, as in Tulich and Mapes (2010) approach, then this assumption is (slightly) violated, so Kuang’s time-averaging of $x$ from a steady run with perturbed forcing is a little closer to the ideal of linearizing around an equilibrium convecting state. To obtain $M$ from Eq. 4.6, right-multiply both sides by $X^{-1}$ to give $M = YX^{-1}$. $X$ is invertible because its columns are all independent, since each perturbed forcing (column of $Y$) gives a unique perturbed state vector column $x$.

If the bumps on the forcing used to probe the CSRM are small enough, the result may be viewed as a linearization around the background forcing. Finite sized bumps had to be used to create a finite response $x$ that was measured as a time average over a finite time running the CSRM at statistical equilibrium, but the system appears usefully linear with respect to such finite perturbations (as tested in Appendix of Kuang 2010).

$M$ was derived by SAM version 6.4 in Kuang (2010) for the first time. Its domain size was 128 km × 128 km with 2-km horizontal grid spacing, and 64 vertical levels with 75 m to 1 km vertical grid spacing until the model top height of 32 km. Kuang also compared 2D CSRM to this 3D CSRM, and checked 2 cases with different mean specified background forcing profiles (RCE with specified $Q_R$ as the only forcing, and another case with $Q_R$ forcing augmented by vertical advection terms estimated from TOGA-COARE rawinsonde array data). Constant surface wind (5 m s⁻¹) and SST
(29.5°C) were imposed in the surface flux calculation, so that the only forcing was the free tropospheric forcing, without unwanted complications from surface flux feedbacks in the energy budget. Comparing with transient CSRM responses like Tulich and Mapes (2010) had done shows that Ms “were able to reproduce those behaviors of the CSRM with some quantitative accuracy” (Appendix of Kuang 2010), although the convective responses through Ms exhibit somewhat different tendencies depending on the mean specified background forcing profiles.

The same way to derive M by CSRM was applied (Kuang 2012) in large ‘2.5D’ (between 2D and 3D) 2048 km x 64 km periodic domain (2048-km case M). In that case, the CRM produces an organized convective system in one part of the domain rather than the random cumulonimbus clouds all over the domain which is the case in an unheared, isotropic 3D 128 km x 128 km CRM domain (leading to the 128-km case M), Figure 4.1 shows the difference in the quadrant of M between these two types of convective cloud system. In this case, Ms are interestingly different. For example, deep positive sensitivity to T’ is seen at lower to mid levels (red columns near 950 - 650 hPa in right panel) is apparently because that layer is part of the warm inflow to the organized MCS. On the other hand, negative sensitivity to 600-700 hPa T’ is seen for the isolated cumulonimbus clouds (blue-dominated columns in the left panel in Fig. 4.1), since those layers function as environmental inhibition reducing the buoyancy (proportional to $T_{updraft} - T_{env}$) of surface-based parcels. Kuang (2012) mentioned that “for strongly organized convection with deep inflows, these sensitivities are consistent with a layer mode of convective overturning (2048-km case M), instead of the parcel mode as in unorganized convection (128-km case M), resulting in a weaker second baroclinic component” when he used
those $M$ matrices as the convection parameterization in his 2D Walker cell-like calculations of a large-scale overturning driven by an SST gradient.

![Figure 4.1](image.png)

Figure 4.1 Adopted from Fig. 8 in Kuang (2012). They are the temperature quadrant of the linear response matrices ($M$s), which are the response of convective heating to the environmental temperature perturbation at each level. Left panel is 3D 128-km case $M$ (representing the response of a periodic CRM with isolated cumulonimbus clouds) and right panel is 2.5D 2048-km case $M$ (representing the response of a mesoscale convective system within the long, narrow periodic CRM). Strictly speaking, these are finite-time response matrices for 4-hour time averaged tendencies (see section 4.2.1 for time averaging procedure).

Similar matrix calculations have been attempted to characterize the response function of single column models (SCMs) using parameterized convection (Herman and Kuang 2013). Some SCMs fail to give a well-behaved steady state in response to steady forcing (Mapes 2015, personal communication), but Herman and Kuang managed to succeed. Their single column models have a problem with deciding on the deep convection owing to the deficiency of their own convective schemes, which are “modulating the strength of deep convection by effects occurring over a limited region in the lower troposphere,” and “avoiding midtropospheric stable layers in the determination of the depth of the deep convecting layer. … Such differences have implications to the simulation of large-scale convective phenomena, such as the growth and propagation of convectively coupled waves (CCW).” They conclude that “the technique employed herein can be used as a basis for tuning and modifying convective parameterization schemes.”
In this chapter, I look into the detailed characteristics of $M$ derived from the small isotropic 3D domain containing a population of isolated cumulonimbus clouds. Since $M$ is the sensitivities of convection to the small perturbation of environmental condition, its eigenmodes are interesting (analyzed in section 4.3) and characteristics of noticeable modes are interpreted in section 4.4. A diagnosis of observation-like data GEOS-5 (from chapter 3) with $M$ is performed in section 4.5. For those analyses, a regridding of Kuang’s $M$ to 10 levels was used, as described in section 4.2.

4.2 Methodology

4.2.1 Finite-time response function (1-hour response)

The matrix $M$ for this study was provided by Prof. Kuang (personal communication), which is 128-km case $M$ in Kuang (2012) for a 3D 128 km $\times$ 128 km isotropic domain that contains isolated cumulonimbus clouds (middle column of Fig. 8a-d in that paper). Since there is an agreement between the basic features of higher (64 levels) and lower (28 levels) vertical resolution $M$s, a low-resolution version is used for the convenience of calculation, and to connect this work to a 10-level GCM (Chapter 5).

Equation 4.6 can be rewritten as below

$$
\begin{bmatrix}
Q'_{1C} \\
Q'_{2C}
\end{bmatrix} = \begin{bmatrix}
d(c_p T')/dt \\
d(Lq')/dt
\end{bmatrix} = M \begin{bmatrix}
c_p T' \\
Lq'
\end{bmatrix},
$$

(4.7)

where $Q'_{1C}$ and $Q'_{2C}$ are almost the same as Yanai et al.’s (1973) famous apparent heating and moistening convective tendencies as this chapter’s Introduction, but here are converted to energy units as $d(c_p T')/dt$ and $d(Lq')/dt$ respectively. Prime indicates deviation from equilibrium state, and bold capital character represents the column vector with vertical dimension, e.g., $T' = T'(k)$, where $k$ is vertical levels. The quadrants of $M$...
could be written in Jacobian form or for easier verbal expression, using the word “from” to indicate the Jacobian denominator, as in Eq. 4.8:

$$M = \begin{bmatrix} \frac{\partial Q'_{1c}}{\partial (c_p T')} & \frac{\partial Q'_{1c}}{\partial (Lq')} \\ \frac{\partial Q'_{2c}}{\partial (c_p T')} & \frac{\partial Q'_{2c}}{\partial (Lq')} \\ \frac{\partial Q'_{2c}}{\partial (c_p T')} & \frac{\partial Q'_{2c}}{\partial (Lq')} \end{bmatrix} = \begin{bmatrix} Q'_{1c} \text{ from } c_p T' & Q'_{1c} \text{ from } Lq' \\ Q'_{2c} \text{ from } c_p T' & Q'_{2c} \text{ from } Lq' \end{bmatrix}. \quad (4.8)$$

As Kuang (2012) mentioned, “instantaneous tendencies are dominated by the fastest decaying eigenmodes and more prone to error” in the matrix inversion process. Therefore, a finite-time response matrix with 1-h averages is calculated for “reducing the dominance of these fast-decaying eigenmodes because their precise decay rates do not significantly affect the interaction of convection with the large-scale flow”. To calculate such a finite-time response matrix, notice that the solution of Eq. 4.5 is

$$x(t) = e^{M_t}x(0), \quad (4.9)$$

which can be expanded for clarity to

$$\begin{bmatrix} c_p T' \\ Lq' \end{bmatrix}(t) = e^{M_t} \begin{bmatrix} c_p T' \\ Lq' \end{bmatrix}(0). \quad (4.10)$$

To remove fast decaying modes, $t = 1$ hour is applied to Eqs. 4.9 or 4.10, then convective tendencies averaged over the hour can be calculated like Eq. 4.11,

$$\begin{bmatrix} \Delta c_p T'/1h \\ \Delta Lq'/1h \end{bmatrix} = \begin{bmatrix} \exp(M \cdot 1h) - \exp(M \cdot 0h) \\ \end{bmatrix} \begin{bmatrix} c_p T' \\ Lq' \end{bmatrix}(0) = \frac{\exp(M \cdot 1h) - I}{1h} \begin{bmatrix} c_p T' \\ Lq' \end{bmatrix}(0) = M_{1h} \begin{bmatrix} c_p T' \\ Lq' \end{bmatrix}(0), \quad (4.11)$$

where $M_{1h}$ is the finite-time response matrix with 1-h averages. One effect of time averaging is to reduce the distracting near-diagonal tripole (+ − +) features in $M$ reflecting local diffusion in the plots. Diffusion is expressed in fast decaying vertically
localized modes that are not essential to couple with large-scale conditions such as through the 1-hour time step of the simple GCM in chapter 5. For the convenience, because only $M_{1h}$ will be analyzed in this chapter, $M$ will indicate $M_{1h}$ in rest of this chapter.

4.2.2 Regridding $M$ to 10 layers

$M$ was adjusted as below for the analysis and for use in a 10-layer GCM with layer centers at 950, 850, …, 50 hPa in chapter 5.

4.2.2.1 Converting from irregular vertical levels

The CRM vertical grid used to derive the original $M$ was irregular, but for clarity and for Chapter 5’s GCM use we want a version of $M$ for 10 layers centered at 950, 850, …, and 50 hPa. The tendencies were integrated (averaged) over these layers, while the sensitivity profiles were interpolated to the layer centers where $T'$ and $q'$ are predicted by the GCM dynamical core, and multiplied by $100/dp$ ($dp$ is the pressure difference between consecutive original vertical levels) to adjust sensitivity for 100 hPa perturbation layers. To accomplish this, both columns and rows of $M$ were first interpolated to a very fine grid (100 layers). Then columns were averaged over 100-hPa layer to preserve the vertical integral properties of heating and moistening. Then rows were interpolated and adjusted. The 100-hPa version of $M$ thus reflects 100-hPa layer integrated heating and moistening for 1000-900, 900-800, 800-700, …, 100-0 hPa as a response to $T'$ and $q'$ perturbations evaluated at the layer midpoints (950, 850, 750, …, 50 hPa) and necessarily assumed to be representative of a 100 hPa layer.

Figure 4.2 shows the 4-hour finite time matrix with 28 layers (Fig. 8 a-d in Kuang 2012), and our 1-hour $M$ with 10 layers (e-h). The horizontal axis is perturbation level
and the vertical axis is response layer with 100-hPa interval. Near-diagonal tripole (+ − +) features in original matrix (Fig. 4.2a-d) are reduced by averaging over 100-hPa layer except moistening tendency to the moisture perturbation ($Q_{2c}'$ from $q'$, Fig. 4.2h). Main warming and cooling, moistening and drying tendencies are similar.

**Figure 4.2** (a)-(d) are quadrants or original matrix adopted from the middle column in Fig. 8 in Kuang (2012). They are (a) convective temperature ($\partial T'/\partial t$) and (b) moisture ($\partial q'/\partial t$) tendencies from the $T'$. (c) and (d) are same as (a) and (b), but from the $q'$. (e)-(h) are same as (a)-(d), but it is newly interpolated $M$ with regular 100-hPa interval, and the energy units ($Q_{1c}'$ and $Q_{2c}'$) from multiplying constants $c_p$ and $L$ respectively. The horizontal axis is perturbation layer, and the vertical axis is response layer. It is from 950, 850, …, 50 hPa (left to right, bottom to top direction) in all quadrants.

### 4.2.2.2 Adjusting imbalanced Moist Static Energy (MSE)

Condensation and eddy fluxes (the definition of “moist convection” tendencies for present purposes) cannot change the column MSE. While the CRM used to derive $M$ conserves MSE and has radiation disabled and surface fluxes highly controlled, the vertical integrations in section 4.2.2.1 may have introduced small imbalances of integrated MSE tendency in each column of $M$. To remove any such small effects, each column of $M$ was adjusted to remove column mean of MSE tendency.
4.3 Analysis of eigenmodes of $M$

To characterize $M$ in another way, its eigenmodes are analyzed in this section. Prof. Kuang has not published such eigen-analyses except the least damped mode (Fig. 14 and 15 in Kuang (2010) although he has looked at them (personal communication).

Any square matrix can be decomposed into eigenmodes, which form a linear basis set. If some vector $V$ is nonzero and $M$ and $V$ satisfy

$$\mathbf{MV} = \lambda \mathbf{V},$$

(4.12)

where $\lambda$ is a scalar, then $V$ is called an eigenvector of $M$ and $\lambda$ is its eigenvalue. Together, $V$ and its corresponding $\lambda$ describe an **eigenmode** of $M$. The set of eigenmodes provides a complete, fundamental characterization of $M$. All eigenmodes are orthogonal, hence eigenmodes can be added or subtracted freely as a complete basis set, capable of describing any arbitrary profile. A real but nonsymmeric matrix like $M$ will have complex $V$-$\lambda$ pairs in general although some of those may be pure real. Using this basis set as coordinates can be called viewing the matrix in its eigenspace.

For a matrix differential Eq. 4.5 above, the solution can be written as Eq. 4.9. When that solution is considered in eigenspace, Eq. 4.9 can be written simply as a set of uncoupled ODEs for eigenvectors $V_m$,

$$V_m(t) = V_m(0)e^{\lambda_m t},$$

(4.13)

where $m$ is eigenmode number, and $\lambda_m$ is eigenvalue of $m$th eigenmode. Since eigenvalues can be complex numbers, $\lambda_m$ is written as below,

$$\lambda_m = (\alpha \pm \beta i)_m,$$

(4.14)

where $\alpha$ and $\beta$ are constant of real and imaginary part respectively. If it is applied to Eq. 4.13,
\[ \mathbf{V}_m(t) = \mathbf{V}_m(0)e^{(\alpha + i\beta)t}, \]  
(4.15)

where

\[ e^{(\alpha + i\beta)t} = e^{\alpha t}(\cos(\beta t) + i\sin(\beta t)). \]  
(4.16)

Therefore, the real part (\(\alpha\)) of an eigenvalue represents exponential growing or decaying according to signs, and the imaginary part (\(\beta\)) represents oscillatory behavior.

Eigenvectors (\(\mathbf{V}_m\)) and eigenvalues (\(\lambda_m\)) are calculated in Octave software’s \textit{eig( )} function. Since some of \(\lambda\)s are complex, modes are ordered by absolute values (= magnitude or modulus) of \(\lambda\)s with both real and imaginary parts. If two eigenvalues have an equal modulus, phase angle by imaginary part determines the order between them. This ordering is arbitrary, a choice made by writers of the \textit{eig( )} function.

4.3.1 Eigenvectors (\(\mathbf{V}\))

Since the whole matrix is 20 by 20, 20 eigenmodes exist. In the case of \(\mathbf{M}\), Only 17 eigenmodes are nonzero because convective tendencies from the \(T'\) in the stratosphere (100-0 hPa) and \(q'\) at two (200-100, 100-0 hPa) top layers were set as zero.

Figure 4.4 shows the whole set of eigenvectors (\(\mathbf{V}\)). Modes with imaginary parts always occur in complex conjugate pairs: 2 eigenvectors with same real parts, but opposite sign of imaginary parts. Each eigenvector (= each column) is a set of both convective tendencies (\(Q'_{1C}\) and \(Q'_{2C}\)) from any perturbation (\(c_pT'\) or \(Lq'\)) with both real and imaginary parts – i.e., each column in Fig. 4.3 is one eigenvector.
Figure 4.3 Eigenvectors of $\mathbf{M}$. (a) and (b) are heating ($Q'_{1C}$) and moistening ($Q'_{2C}$) tendencies in real part. (c) and (d) are same as (a) and (b), but in imaginary part. The horizontal axis is mode number and the vertical axis is vertical levels. Unit is [kJ/kg/sec].

Four modes are selected for further examination for their relatively strong (large $\lambda$) or weak tendencies (the least damped mode), or interesting features (with strong imaginary parts): 1$^{\text{st}}$ (black), 2$^{\text{nd}}$ (red), 6$^{\text{th}}$ and 7$^{\text{th}}$ (blue), and 17$^{\text{th}}$ (green box) modes. The 1$^{\text{st}}$ and 2$^{\text{nd}}$ modes are deep convective modes that convection is especially aggressive in damping out, the 6$^{\text{th}}$ and 7$^{\text{th}}$ modes are a congestus-deep convective oscillatory mode producing descending $Q'_{1C}$ and $Q'_{2C}$ features with ~1 day period, and 17$^{\text{th}}$ mode is the slowest decaying mode corresponding to column-integrated MSE perturbations which convection can hardly affect. Detailed analyses to support these interpretations are in section 4.4.
4.3.2 Eigenvalue ($\lambda$)

Eigenvalues also consist of real and imaginary parts. Referring to Eq. 4.16 above, the real part ($\alpha$) represents coefficient of exponential growing (+ sign) or decaying (− sign). In other words, it is the inverse of growing (or decaying) timescales ($e$-folding time, section 4.3.3). In this case of $\mathbf{M}$, eigenvalues of real parts ($\alpha$) are all negative, since a periodic CRM with steady forcing is a stable system that is always relaxing toward radiative-convective equilibrium, so there are no unbounded growth modes (Fig. 4.4a). Imaginary part ($\beta$) represents oscillation, as Eq. 4.16 above makes clear. Like eigenvectors, the complex eigenvalues occur in conjugate pairs (Fig. 4.4b).

![Figure 4.4](image.png)

Figure 4.4 Eigenvalues of $\mathbf{M}$ in (a) real and (b) imaginary parts. The horizontal axis represents mode number.
4.3.3 e-folding time

Since the real part ($\alpha$) of an eigenvalue represents the growth or decay rate of that mode, the e-folding time ($\tau_{\text{decay}}$) is defined as below,

$$\tau(m)_{\text{decay}} = \frac{1}{\text{real}(\lambda_m^m)}.$$  \hspace{1cm} (4.17)

e-folding time shows how long it takes convection in a periodic CRM to adjust a perturbed environment with the eigenvector profile structure toward the reference state (= equilibrium state), perhaps oscillating during the decay. The e-folding time increases gradually from an hour to a day (Fig. 4.5a), suddenly peaking at 5 days in the slowest mode (17th mode). e-folding times in other selected cases are ~45 mins, 2.8, and 6.7 hours in 1st, 2nd, and 6th & 7th modes respectively (Fig. 4.5b, enlarged version of Fig. 4.5a). In section 4.4, e-folding time will be analyzed with other characteristics of eigenvectors as well.
Figure 4.5 e-folding time ($\tau_{\text{decay}}$) of eigenmodes of $M$. (a) is for all 20 modes [day], and (b) is focused on until 10th modes [hour]. The horizontal axis is the mode number and the vertical axis is time.

4.4 Characteristics of eigenmodes

4.4.1 First mode (black box in Fig. 4.3)

Figure 4.6 contains the same information as black box in Fig. 4.3. This 1st mode shows that a cool troposphere and moist PBL are very rapidly damped by convection, presumably by deep convection as anomalously moist PBL parcels experience enhanced buoyancy in an anomalously cool troposphere.

Since this is the eigenstructure, sign is arbitrary and these tendencies can be interpreted in opposite sign as well: a dry PBL and warm troposphere are rapidly damped by negative anomalies of deep convection. In physical terms, in this negative sign case,
the background surface flux and radiation would act with less opposition from convection, moistening the PBL and cooling the troposphere respectively.

Kuang calls this the “CAPE damping” mode (personal communication) and its timescale of ~45 mins (Fig. 4.5) for the ~7 mm/h rainrate background convection intensity (balancing a corresponding cooling rate and surface fluxes) may be relevant for CAPE-damping convection schemes (Betts and Miller 1986, Zhang and McFarlane 1995). However, it did not act like the damping mode when it is applied to the global model in chapter 5, so it will be discussed more carefully later.

![Figure 4.6 Eigenvector of 1st mode. It is same as the black box in Fig. 4.3, but illustrated as a profile plot. The vertical axis is vertical level [hPa]. Unit is [kJ/kg/sec]. Red vertical line is zero.](image)

Since eigenmodes are orthogonal, $\mathbf{M}$ can be reconstructed mode by mode with the sum of all reconstructed $\mathbf{M}$ by each mode. Reconstructed matrix ($\text{Recon}\mathbf{M}_m$) by each mode is calculated as,
\[ \text{Recon} \mathbf{M}_m = \mathbf{V} \cdot \Lambda_m \cdot \text{inv}(\mathbf{V}), \]  

(4.18)

where \( \mathbf{V} \) is the original eigenvector matrix, \( \Lambda_m \) is the diagonal eigenvalue matrix with only eigenvalue of \( m \)th mode, \( \text{inv}(\ ) \) is the function for inverse matrix in Octave, and \( \text{Recon} \mathbf{M}_m \) is the reconstructed matrix by \( m \)th mode. Basic structure of each column in reconstructed matrix is same as that of \( m \)th eigenvector with different magnitude and sign.

Figure 4.7 Four panels on the left are same as Fig. 4.3e-h. Four panels on the right are same as left set, but reconstructed \( \mathbf{M} \) by only 1st mode. The horizontal axis is perturbation layer, and the vertical axis is response layer. Unit is [kJ/kg/sec].

Quadrants of reconstructed matrix by only 1st mode are on the right in Fig. 4.7e-h and original matrix on the left (Fig. 4.7a-d) for comparison. We see in Fig. 4.7 the following features:

1) Warming responses from both positive \( T' \) and \( q' \) in boundary layer (1000-900 hPa): Original warming tendencies above 900 hPa from \( T' \) and \( q' \) are reconstructed by 1st mode (orange-red left column in Fig. 4.7e and g). On the other hand, cooling tendency in boundary layer from the \( T' \) (blue square in lower left corner in Fig. 4.7a) is absent in this reconstruction.
2) Cooling responses from the $T'$ in lower to middle troposphere (900-600 hPa): The inhibition effect of convection is partly reconstructed (blue columns in Fig. 4.7a and e).

3) Warming responses from the $q'$ in lower to upper-middle troposphere (900-400 hPa): Yellowish columns in Fig. 4.7g.

4) PBL and shallow convective moistening responses from $T'$ and $q'$ at various levels: Dominant blue or red rows in boundary to lower levels in Fig. 4.7f and h.

The most noticeable characteristic in this reconstruction by 1st mode is that most of the original heating (Fig. 4.7a) and moistening (Fig. 4.7c) convective tendencies are already reconstructed – color bars are same in this figure. Especially, moistening tendencies in lower levels are reconstructed as almost same as original. Therefore, 1st eigenmode could be a dominant mode for the cooperation between convection and PBL. The large absolute eigenvalue corresponds to strong tendencies and a fast decaying time. The absolute eigenvalue of 1st mode is significantly larger than others (black box in Fig. 4.3a), therefore it decays in very short time (~45 mins, black box in Fig. 4.5). This fast decaying pattern of 1st eigenvector with time is shown in Fig. 4.8.
Figure 4.8 Evolution patterns of 1st eigenvector with time. (a) is $Q'_1C$ and (b) is $Q'_2C$ in real part. The horizontal axis is time [hr] and the vertical axis is vertical level [hPa]. Unit is [kJ/kg/sec].

Since $e$-folding time of 1st mode is very short (~45 mins), both cooling (Fig. 4.8a) and moistening (Fig. 4.8b) tendencies are decayed within 1-2 time steps (~1-2 hours). There is no oscillating evolution feature with time because of no imaginary part in 1st mode.

4.4.2 Second mode (red box in Fig. 4.3)

Figure 4.9 contains the same information as the red box in Fig. 4.3: it is the 2nd mode eigenvector. It indicates that a convecting CRM would rapidly (in about 3h) damp a moist PBL with a dry anomaly at 850 hPa. With an anomalously warm troposphere above the moist PBL, this structure does not project onto the 1st mode, so the response is not simply deep convective, but rather a transfer of moisture between 850 hPa and 950 hPa levels, with some mild connection to deep convection.

The 2nd mode is orthogonal to the 1st mode, with opposite signs between boundary layer and free troposphere. In this sign, it seems like deep convective downdrafts, which is cooling (Fig. 4.9a) and moistening (Fig. 4.9b) on the 1000-900 hPa layer. On the other hand, heating and drying tendencies on the layers (850-750 hPa) above the boundary layer compensate the opposite tendencies by eddy flux to the lower layer (Fig. 5a in
Tulich et al. 2007). If this mode sign is reversed, it looks like shallow convection updrafts with compensating layers above. This is correlated with reduced deep convection with anomalous negative heating at upper levels.

Figure 4.9 Same as Fig. 4.6, but 2nd mode.

Figure 4.10 Same as Fig. 4.7, but by only 2nd mode.
Figure 4.10 shows reconstructed matrix by only 2\textsuperscript{nd} eigenmode. Convective responses to the perturbations in 950 and 850 hPa levels are reconstructed. Features include:

1) Heating responses from the $T'$ in boundary layer (1000-900 hPa): Cooling tendency in the boundary layer (blue square in lower left corner in Fig. 4.10e), which is missing from 1\textsuperscript{st} mode (Fig. 4.7e), is lightly reconstructed here. Top-heavy warming tendencies in the upper levels (left column in Fig. 4.10a and e) are emphasized by 2\textsuperscript{nd} mode.

2) Heating responses to the $T'$ in 900-800 hPa: Top-heavy cooling tendencies in the upper levels (second from left columns in Fig. 4.10e) are emphasized.

Figure 4.11 shows the decaying patterns of 2\textsuperscript{nd} mode with time from an initial state of its eigenvector with positive (Fig. 4.11a and b). Its $e$-folding time is about 2.8 hours (Fig. 4.5) toward equilibrium state, and all features are decayed in 5-10 hours. There are no oscillating features with time.

Figure 4.11 Same as Fig. 4.8, but in 2\textsuperscript{nd} eigenvector.
4.4.3 6th and 7th mode pair (blue box in Fig. 4.3)

6th and 7th modes do not have a large real part of the eigenvalue; they are only weakly damped (~6.7 hours in Fig. 4.5). But the total magnitude of the eigenvalue is large because of large imaginary part. Their imaginary eigenvalues are the largest values among others (Fig. 4.5). Therefore significant tendencies in M may associate with this mode pair. Recalling the math result that asymmetry in M is what makes eigenvalues and eigenvectors complex in section 4.3, it will not be surprising that the contributions of these modes to M are asymmetric across the diagonal. This pair has same real profiles (Fig. 4.12a and b), but opposite-signed imaginary profiles (Fig. 4.12c and d). The real profile of the mode’s heating tendencies shows cooling in upper half and mild warming in lower levels (Fig. 4.12a). Moistening tendencies are general in all levels except very upper atmosphere, which is only slightly drying (Fig. 4.12b). And, 6th mode make a pair mode with 7th mode because of imaginary part (Fig. 4.12c and d), which is exactly same with opposite sign.
Reconstructed matrix by 6th (or 7th mode) consists of real and imaginary parts. Reconstructed real parts are same in both modes (Fig. 4.13e-h) and imaginary parts are same but opposite sign (Fig. 4.14). Therefore, the reconstructed matrix will be purely real when the mode pair is used in reconstruction, and is twice the real matrix shown in Fig. 4.13e-h. In previous cases (1st and 2nd modes), all columns of reconstructed M reflect the same eigenvector structure with only different signs and magnitudes. In this case (6th and 7th modes), however, eigenvector structure (blue box in Fig. 4.3) is not fixed, but makes a tilted structure in the reconstructed M (Fig. 4.13e-h). Even though Fig. 4.13e-h shows only real parts of reconstructed M, both real and imaginary parts of eigenvector are used for this reconstruction (Eq. 4.18) and make these tilted patterns which lead to an oscillatory behavior. Tendencies straddle the perturbation levels along the diagonal (Fig. 4.13a and e, b and f), indicating vertical propagation of anomalies in time.
Figure 4.13 Same as Fig. 4.7, but (e-h) are only real parts of reconstructed $M$ by only 6th (or 7th) mode.

Figure 4.14 Same as Fig. 4.7, but imaginary part of reconstructed $M$ by 6th (a-d) and 7th (e-h) modes.

Figure 4.15 shows how the 6th (or 7th) eigenmode evolves with time. Again, this is simply a solution to Eq. 4.5 above, but with $V$ and $\lambda$ as complex numbers now (equation 4.15 and 4.16). 7th mode is exactly same as 6th mode but opposite sign in only imaginary part. Imaginary part of eigenvalue affects both real and imaginary evolution with time (as shown in Fig. 4.13 and 4.14), so there is oscillating (at each level) or, more to the point,
descending evolution patterns with time in all figures. While $e$-folding time is about 6.7 hours, the period of an oscillation is about 50-60 hours despite very weak signals after 10-20 hours. Evolution of heating tendency with time (Fig. 4.15a) shows the transition from cooling to warming tendencies in the upper half layer, and at the same time it is opposite in lower half layers. Because it involves shifts of the top height of the total heating profile, this mode of variability may be viewed or interpreted as an oscillation between middle-height (cumulus congestus) and deep (cumulonimbus) convection, and could be called the “congestus-deep oscillator” mode.
Figure 4.15 Same as Fig. 4.8, but in 6\textsuperscript{th} and 7\textsuperscript{th} eigenvector. (a) and (b) are real parts of 6\textsuperscript{th} or 7\textsuperscript{th} mode (same). (c) and (d) are imaginary parts of 6\textsuperscript{th} mode, and (e) and (f) of 7\textsuperscript{th} mode (opposite sign to the c and d).
4.4.4 The least-damped mode – 17th mode (green box in Fig. 4.3)

Figure 4.16 Same as Fig. 4.6, but 17th mode.

Figure 4.16 is the eigenvector of the slowest decaying mode. This is the same mode as the slowest decaying mode in Kuang (2010, his Fig. 14), therefore the slowest eigenvectors between this study and his paper (2010) are compared in Fig 4.17. Fig. 4.17a and b are same as Fig. 4.16a and b, but redrawn in same range values in the horizontal axis. Fig. 4.17a and b are the heating and moistening tendencies in 17th mode in this study, and solid-circle line in Fig. 4.17c and d are the temperature and specific heating profiles in the least-damped mode in Kuang’s matrix (2010). The eigenvector of the slowest mode in both cases shows similar patterns. According to Kuang (2010), “the temperature profile (solid-circle line in Fig. 4.17c) of this eigenmode resembles a shift of a moist adiabatic profile (dashed line in Fig. 4.17c), and its specific humidity profile (solid-circle line in Fig. 4.17d) approximately conforms to the expected change with the
relative humidity profile fixed to that of the reference state (dashed line in Fig. 4.17d).” He also mentioned that this slowest eigenmode “resembles the reference profile used in the Betts-Miller scheme,” in which “the long e-folding time scale is roughly that needed for the anomalous surface fluxes to remove the column-integrated moist static energy anomaly associated with this eigenmode” because radiation is disabled (recall that constant radiative cooling was applied to eliminate radiative feedback). The time to adjust away such a column-integrated MSE anomaly should be infinite if we succeeded in making $M$ conserve moist static energy exactly, but the smallness of this very small eigenvalue may be limited by numerical accuracy or by the fact that we treat $c_p$ and $L$ as constants here while the CRM that derived $M$ may use a slightly different formulation of thermodynamics.

Since its eigenvalue is very small (Fig. 4.4), reconstruction of matrix from this mode has very small values and is an extremely subtle and small part of $M$ (not shown).
4.5 Application of M to the GEOS-5 data

How does M work to the observation perturbation? How does M reconstruct the convective response (R’, Q1’, or Q2’) with observation? In this chapter 4.5, this M is applied to the GEOS-5 from chapter 3. Although GEOS-5 data is not observation, it is calculated without convective scheme in the global model, therefore convection is simulated like in nature. Applying M to the GEOS-5 data would give a sense how linearity calculated by cloud resolving model works on the nature-like simulated data – whether M gives meaningfully interpretable insight for convection or not.

In this examination, composite cases of T’ and q’ in chapter 3 are used as environmental forcing perturbation. At first, vertical array of composite T’ and q’ profiles with time in Fig. 3.8a and b are interpolated in 10 layers. Figure 4.18 shows the 10-layer version of composites of T’ and q’ profiles in Fig. 3.8. The major characteristics of composite T’ and q’ profiles – cold tongue ahead of the event time (t = 0) at 900-800 hPa in Fig. 4.18a and 3-layer humidity pattern with two time jumps in Fig. 4.8b – are still clear in this version.

Figure 4.18 Composites of (a) T’ and (b) q’ profile of the heaviest eight rainrate cases that are area-averaged in 250-km area in GEOS-5 data. They are same as Fig. 3.8 (a) and (b) respectively, but vertical layers are interpolated in 10 layers.
To see the participation of $T'$ and $q'$ only, 3 examinations are performed. “$T \text{and} q$” is the examination that both $T'$ and $q'$ are multiplied by $M$ to calculate convective responses. “$T \text{only}$” and “$q \text{only}$” are the examinations that only $T'$ or $q'$ is used respectively.

Table 4.1 Description of experiments

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<tr>
<td>$T \text{and} q$</td>
<td>Both $T'$ and $q'$ profiles are multiplied by $M$</td>
</tr>
<tr>
<td>$T \text{only}$</td>
<td>Only $T'$ profiles are multiplied by $M$</td>
</tr>
<tr>
<td>$q \text{only}$</td>
<td>Only $q'$ profiles are multiplied by $M$</td>
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Figure 4.19 (a) and (b) are the heating and moistening responses to both $T'$ and $q'$ in “$T \text{and} q$” case. In Fig. 4.19a, both heating and moistening tendencies are sharply changed from 3 hours after the event time ($t = 0$), i.e., deep heating to cooling and drying to moistening. One noticeable feature is that the change of heating tendencies is quite neutral in the surface layer (1000-900 hPa) in Fig. 4.19a. Those tendencies are basically similar to the convective tendencies by “$q \text{only}$” in Fig. 4.19e and f. In other words, $q'$ is the important factor to decide the basic convective characteristics than $T'$. 
Figure 4.19 Convective heating and moistening tendencies by composite $T'$ and $q'$ profiles of GEOS-5 multiplied by $M$. (a) and (b) is heating and moistening tendencies by “Tandq” respectively. (c) and (d) are same as (a) and (b), but by “Tonly”. (e) and (f) are same as (a) and (b), but by “qonly”. Horizontal axis is time from -18 to +18 hours to the event time ($t = 0$), and vertical axis is vertical layer [hPa].

Heating and moistening tendencies by “Tonly” show quite different patterns than “Tandq” or “qonly”. The most deep heating happens at $t = +3$ hours, which is the
transient time in “Tandq” and “qonly” examinations. Consequently, the reconstructed rainrate also shows the heaviest rainrate at \( t = +3 \) hours in Fig. 4.20b. It may show that the cold pool effect at the surface is more than compensated by cold tongue at 900-800 hPa in Fig. 4.18a, which represents the slight rainrate dip at \( t = -3 \) hours in Fig. 4.20b. However, the magnitude of convective tendencies by \( T' \) (Fig. 4.19c and d) is only half than “Tandq” or “qonly”, and those convective tendencies by \( T' \) are not kept any more if \( q' \) effect is added to \( T' \) in “Tandq”.
Reconstructed rainrate based on the heating tendencies in Fig. 4.19. (a) is calculated based on heating tendency by “\text{Tandq}”, (b) is by “\text{Tonly}” and (c) is by “\text{qonly}”. Horizontal axis is time from -18 to +18 hours to the event time \((t = 0)\). Rainrate unit is [mm/hr].

4.6 Summary

This chapter has described and analyzed a linear response function or matrix algebra approach to characterizing how convection interacts with its larger-scale averaged thermodynamic profiles (dry static energy \(c_pT\) and latent energy \(Lq\)). The linear responses of a cloud resolving model, when perturbed around a convecting equilibrium state associated with a 7 mm/d rainrate, have been mathematically encapsulated in a matrix \(M\).
by Z. Kuang, who shared his matrix. The chapter described integrating $M$’s effects over a finite time (1 hour), regridding to 10 equally spaced pressure layers, and my work (not published by Kuang or anywhere else) interpreting and characterizing $M$ in terms of its eigenspace as well as its physical-space interpretation.

$M$ is also applied to the GEOS-5 data, which is simulated like nature. Even though the exact interpretation is not available, comparison between experiments according to the different environmental perturbations like “T and q”, “T only”, or “q only” provided the role of each factor to the convection.

In the next chapter, $M$ is used as an anomaly convection parameterization in a 10-layer spherical primitive equation solver.
Chapter 5: Using M as an anomaly convection scheme in an otherwise ‘dry’ GCM

The linear response function $M$ and its subcomponents can be used as an anomalous convection parameterization within a coarse-mesh dynamical computation, to see what large-scale convectively-coupled variability may result (and how those waves might in turn interact with global flow). Kuang (2010) showed that $M$ coupled to single-wavelength large-scale linear gravity dynamics is unstable to convectively-coupled waves, even though $M$ itself is stable with all real parts of eigenvalues negative. $M$ behaves just like a CRM in this respect – damping all anomalies toward the basic convecting equilibrium state locally, but with sufficiently slow timescales and oscillatory modes (the complex eigenmodes 6th and 7th in section 4.4.3). Those eigenmodes can couple with neutral or even mildly damped modes of the larger scale dynamics to produce an instability in the coupled system called “stratiform instability” by Mapes (2000), but with moisture field aspects as well (Kuang, 2008b).

In this section, we check for such instability in a spherical and multiple wavelength context within realistic background flows, by using a model called “Jalopy with matrix” for reasons explained below.

In a full physics GCM, convection parameterizations affect both the mean state (as described in Chapter 2) and the simulated weather variability. It is not easy to separate the direct effects of convection parameterization on tropical weather, from the indirect effects where convection changes the background global mean flow, which then in turn changes the weather dynamics. This ambiguity is especially dangerous for studies of the Madden Julian oscillation (MJO), as illustrated by Maloney and Hartmann (2001). To separate the mean flow effects from large-scale weather variability we use a spectral
primitive equation solver on the sphere (Sela 1980; due to its age we call this dynamical core a “Jalopy”) to solve large-scale dynamics. The equations being solved include the $T$ and $q$ equations (Eqs. 4.3 and 4.4). The “convective” terms on the right, such as the heat budget term $Q_{hc} = Q_i - Q_R = L(c - e) - \sigma \left( \tilde{s} \omega' \right) / \partial p$ can be divided into a time invariant forcing that maintains the background, plus anomalous tendencies due to anomalous convection. The anomaly approach allows us to separate the direct effects of convective response function on weather, within a fixed basic state flow.

Using time-invariant forcing, cleverly created as in Hall and Manabe (2000), the dry Jalopy develops a realistic climatology with jet streams and synoptic instabilities. A tracer called $q$ is also included, which advects like water vapor and also has a 3D time-invariant forcing $Fq(x,y,p)$ making its climatology realistic. The ‘anomalous’ ‘convective’ tendencies of $T$ and $q$, added to the Jalopy’s budget equations for grid-scale $T$ and $q$, are given by $Mx$ (Eq. 4.5) where $x$ is the state vector ($[T, q]$ minus its long time mean at that location, as measured from a long no-matrix run). 1 hour is roughly the time step of the GCM time integration scheme, which is one of the reasons for choosing $M$ with 1-hour average in chapter 4 (section 4.2.1).

Since mean state is already contained in the calibrated forcing on the baseline in this model, we can isolate the direct effects of convection on simulated weather features from the indirect effects through mean state changes.

5.1 Model and methodology

In Model Jalopy, “calibration run” is performed at first to define climatology as a realistic mean state. The dry primitive equations are solved by a spectral transform method with R30 horizontal resolution and 10 vertical layers centered at 950, 850, … , 50
hPa in sigma coordinates on a spherical earth. Horizontal resolution is rhomboidal truncation R60 (about 1.8 lat, 2.3 lon) or R30 in more recent runs. There is no topography in this simulation. As mentioned above, time-independent source terms are necessary to give the model climatology like the observed earth atmosphere as in Hall and Manabe (2000). Source terms (time independent forcing of temperature, specific humidity, surface pressure, divergence, and vorticity) were derived as the negative of the average of all the first-timestep changes the model produced when initialized from a large set of instantaneous reanalysis states in the JJA season. As a result, the mean of these initial states becomes the climatology for all future simulations with the source terms prescribed. In this run, dates and times are arbitrary except that JJA season is the baseline. Therefore, there is no diurnal cycle, no annual cycle, and no observed initial condition.

Once the forcing is determined as above, a dry “control run” is done. There is no time-dependent forcing except mild nudging to temperature climatology with 20-day timescale for preventing drift of absolute temperature values. Since no physics scheme for moist convection is not used in this run, and $q$ is just a passive tracer, it is called “dry run”.

“Experiment runs” are similar to the dry “control run”, but anomaly physics ($\mathbf{M}$) is applied to the fluctuations of $T'$ and $q'$ (defined as temporal anomalies from the climatology of the dry run). Fluctuations occur even in the dry run due to hydrodynamic instabilities and advection, and in the experiment those are allowed to couple with $\mathbf{M}$. For this anomaly physics, $\mathbf{Mx}$ gives physical tendencies, as described in section 4.4. Column deviations $T'(p)$ and $q'(p)$ from the time-independent climatology of the dry run are used
as the input of the matrix $\mathbf{M}$, and the tendencies computed as $\mathbf{M}$’s output are applied to the model’s governing equations for $T(p)$ and $q(p)$.

In first runs analyzed here, $\mathbf{M}$ was applied globally as well as uniformly. However, $\mathbf{M}$ is originally calculated to act like a convecting column, so applying it at the poles is not realistic. Further runs will mask it to apply only in the tropics. Furthermore, in these initial runs, the model was initialized with a state far from its attractor. As a result, large-amplitude pole-to-pole oscillations, involving initial condition shock coupled with $\mathbf{M}$ everywhere, are the phenomenon examined here. Future runs with tropically localized $\mathbf{M}$ and initial conditions from the model itself will be used, and those results will be much more relevant to the Earth.

Still, these early runs allow us to see how $\mathbf{M}$ couples with dynamics to produce convectively coupled waves compared with “dry run” in the global model. Even though some strange and unrealistic kinds of disturbance patterns are expected, we start with this simplest setup and focus on differences between “NoM” and other experiments with various $\mathbf{Ms}$ (Table 5.1).

In this section, the model’s spectrum of tropical variability change in response to this use of $\mathbf{M}$ is examined. In addition, modified $\mathbf{Ms}$ using the results in section 4.4 – modes included or excluded – are used in this model to see the effects of noticeable eigenmodes. This examination is designed with 5 experiments including “control run”.

Table 5.1 Description of experiments

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<th>$\mathbf{M}$</th>
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<tr>
<td>NoM</td>
<td>Without $\mathbf{M}$</td>
<td>Dry run (Control run)</td>
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<tr>
<td>FullM</td>
<td>Original $\mathbf{M}$</td>
<td>Run for examining the $\mathbf{M}$ with all modes</td>
</tr>
<tr>
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<td>$\mathbf{M}$ without mid-level $q'$</td>
<td>Run for examining the absence of mid-level $q'$</td>
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<tr>
<td><strong>No1stmode</strong></td>
<td><strong>M without 1\textsuperscript{st} mode</strong></td>
<td>Run for examining the absence of ‘1\textsuperscript{st} mode’</td>
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<tr>
<td><strong>No6&amp;7thmodes</strong></td>
<td><strong>M without 6\textsuperscript{th} and 7\textsuperscript{th} modes</strong></td>
<td>Run for examining the absence of ‘6th and 7\textsuperscript{th} modes’</td>
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</tbody>
</table>

“\textbf{NoM}” is exactly same as the “control run”, which is a run without \textbf{M}. Comparison of “\textbf{FullM}” with “\textbf{NoM}” will show how \textbf{M} works in the GCM and its functions to the hydrodynamical weather variability. “\textbf{NoMidlevMoist}” is designed for examining the role of mid-level moisture perturbation to the convection. Mid-level moisture is known as having a role of generating and enhancing convection in reality (Grabowski and Moncrieff 2004, Khouider and Majda 2006, Kuang 2008a, b). \textbf{M} for “\textbf{NoMidlevMoist}” is obtained by removing all convective tendencies (= responses) to the $q’$ in 700-400 hPa (Fig. 5.1).

\textbf{M} without $q’$ in mid-levels (700-400 hPa)

![Diagram](image)

Figure 5.1 Same as Fig. 4.3e-h, but no $q$ perturbations in mid-levels (700-400 hPa). This is \textbf{M}(NoMidlevMoist).

Comparison of “\textbf{No1stmode}” (or “\textbf{No6&7thmodes}”) with “\textbf{FullM}” will give the role of 1\textsuperscript{st} mode (or 6\textsuperscript{th} and 7\textsuperscript{th} modes) via absence of these modes. \textbf{M} for “\textbf{No1stmode}” (or
“No6&7thmodes”) is calculated by subtracting reconstructed $M$ with only 1\textsuperscript{st} mode (or 6\textsuperscript{th} and 7\textsuperscript{th} modes) to the original $M$ like below,

\[ M(\text{No1stmode}) = M - M_1, \]

\[ M(\text{No6&7thmodes}) = M - (M_6 + M_7), \]

where $M_1$, $M_6$, and $M_7$ are the reconstructed $M$ by 1\textsuperscript{st}, 6\textsuperscript{th}, and 7\textsuperscript{th} eigenmodes. All GCM integrations are done by collaborator Dr. P. Kelly. Experiments were run during 60 days (labeled arbitrarily as running from June 1\textsuperscript{st} to July 30\textsuperscript{th}, although the only relationship to the calendar is that JJA season data were used in the calibration of the mean forcing).

5.2 Results

Figure 5.2 shows the longitude-time patterns of meridional mean omega ($\omega$) between 5\textdegree{}S to 5\textdegree{}N at 550 hPa in all experiments. Since model’s resolution is about 2 degree, 5\textdegree{}S to 5\textdegree{}N is chosen to average over the unrealistic near-gridscale noise. Compared with several areas for meridional mean (i.e., 2.5\textdegree{}S to 0, 10\textdegree{}S to 10\textdegree{}N, or 15\textdegree{}S to 15\textdegree{}N), the typical characteristics of meridional mean patterns in each experiment were not very different (not shown).

Compared with “NoM” (Fig. 5.2a), “FullM” creates tropical waves via an instability (Fig. 5.2b), even though all the real parts of the eigenvalues of $M$ are negative (section 4.3.2). Convection interacting with dynamics across multiple columns (in GCM) is a different system than convection alone in a closed periodic domain (in CSRM), therefore the former may be unstable even thought the latter is stable. One noticeable but strange feature is the zonally uniform oscillations (horizontal stripes), especially between 60\textdegree{} to 180\textdegree{}, which animations show are bizarre pole-to-pole zonal oscillations from the initialization shock. Of course these are unrealistic weather disturbances, but they are still
convectively coupled waves since they do not appear in the “NoM”. At this point, showing significant differences among model runs with and without various M's is sufficient to the purpose of this chapter, even though the wave types are unrealistic.

“NoMidlevMoist” (Fig. 5.2c) shows similar patterns as “FullM”, but more horizontally parallel propagation with a little bit weaker magnitude. Considering that “NoMidlevMoist” excludes the convective response from any moisture forcing at 700 – 400 hPa, mid-level moisture may not have a critical role for organizing convectively coupled waves. This result appears to disagree with the “moisture-stratiform instability”, found by Kuang (2008) for linear gravity waves. “No1stmode” (Fig. 5.2d) did not show any significant convective coupled wave patterns while “No6&7thmodes” (Fig. 5.2e) produces well-organized eastward propagating patterns with very strong magnitudes. This is a kind of opposite result than we expected. We expected that the 1st mode would act as simple damping since it is simply a real, negative eigenvalue. If that were the case, the “No1stmode” experiment might be expected to exhibit enhanced activity. The model results are opposite to that expectation. Likewise, we expected that the 6th & 7th modes (with its complex eigenvalue) might be the essential mechanism for M-coupled waves. If that were true, “No6&7thmodes” would lack waves entirely. Again, the model results are opposite to that expectation. It is clear we do not fully understand how the eigenmodes of M are conspiring to produce the model’s convectively coupled disturbances.
Figure 5.2 Hovmöller diagrams of averaged omega (ω) between 5°S to 5°N at 550-hPa. Time evolves down to up. (a) is “NoM”, (b) is “FullM”, (c) is “NoMidlevMoist”, (d) is “No1stmode”, and (e) is “No6&7thmodes” run.
Even if their horizontal structures are very unrealistic, comparison of vertical structures between experiments provides a general sense of how convective tendencies in the convectively coupled variability of the GCM depending on the Ms. Figure 5.3 shows a longitude-vertical section of omega, convective heating, and moistening tendency at a random moment in each experiment.

Longitude 60 to 180 is the horizontal axis and pressure is the vertical axis. Since time is arbitrary and detailed patterns are not realistic for this simplest setup with globally and uniformly applied M, I just picked up cases at a random moment rather than composites for catching a glimpse of the characteristics. Therefore the detailed features are not analyzed here, but comparison of differences between experiments is done.

Basically, the strength of vertical structures of \( Q_1' \) and \( Q_2' \) is roughly proportional to \( \omega \) as seen in Fig. 5.2. Compared with “FullM”, “No1stmode” shows very weak and “No6&7thmodes” shows very strong heating and moistening profiles in Fig. 5.3. Although the vertical structures of \( Q_1' \) and \( Q_2' \) are relatively narrow (“No6&7thmodes”) or wide (“No1stmode”), they are mostly combined with monopole and dipole.
Figure 5.3 Longitude-vertical sections of \( \omega \), convective heating \( (Q'_{1c}) \), and convective moistening tendency \( (Q'_{2c}) \) at a random moment. The horizontal axis is longitude from 60 to 180, and the vertical axis is the vertical levels [hPa]. Color scales are same in each variable. (a-c) are \( \omega \), \( Q'_{1c} \), and \( Q'_{2c} \) in “FullM” run respectively. (d-f) are same as (a-c), but in “NoMidlevMoist”.
5.3 Discussion and summary

Emanuel et al. (1994) argued that “the direct effect of convection in large-scale circulations is to reduce by roughly an order of magnitude the effective static stability felt by such circulation, and to damp all of them.” They named it as “moist convective damping (MCD)”. This wave destroying process by deep convection is also consistent with the “CAPE controlled” process in Mapes (2000). We expected that the 1st
eigenmode (purely real and negative eigenvalue) with deep heating monopole can be interpreted in this manner, but it was not according to the “No6&7thmodes” experiment. Since congestus oscillator mode (6th and 7th modes) is absent in “No6&7thmodes”, strong damping by dominant 1st mode was expected, but it showed even stronger oscillating and persistent convective signals in Fig. 5.2e and Fig. 5.3j-l.

In basic fluid energetics, $Q_{IC}$ contributes to disturbance kinetic energy (KE) via available potential energy (APE) through the correlation of heating with temperature. Heating in warm places adds APE which is quickly converted to KE, while heating in relatively cold places destroys APE. If convective heating is governed by lifted-parcel buoyancy, then it should damp APE as Emanuel et al. (1994) argued about MCD.

However, the 1st mode in our experiment seems not like acting as damping convection, but enhancing it. This is the key mechanism of stratiform instability for convectively coupled waves showed as the schematic picture in Fig. 5.4 (adopted from Fig.5 in Mapes 2000). According to Mapes (2000), if “the lower troposphere is anomalously cool, this implies decreased CIN, which enhances deep-convection frequency, which leads, after a time delay of $T_{meso} = 3$ h, to anomalous stratiform heating in the upper troposphere and cooling in the lower troposphere”. In other words, instability to enhance convection is possible because of the in phase between anomalously cool and the cooling in the lower troposphere. Therefore, it may suggest that the essential role of 1st mode is not the CAPE damping, but deep layer heating to lower level inhibition.
In this chapter, several modified $M$s are applied in the global model as the convective scheme, and the roles of major convective modes according to the analyses in chapter 4 are compared each other. The expectation for the role of major eigenmode of $M$ based on their characteristics in chapter 4 does not quite happen when they are applied in the global model. However, differently modified $M$s showed clear distinction among those examinations. Using the linear relationship by elegantly calculated $M$ would be a challenge, but really a promising way to provide some insight for interactive mechanism between convection and its environment.
Chapter 6: Conclusion and summary

This dissertation is about how to use abundant data sets wisely for better understanding about convection in the seamless and interactive atmosphere. A subtitle of this dissertation might be “Three views of convection-global interaction”. These are distinctive, they are not the only views, and the thing they have in common is their overall subject. Three main efforts were performed by uniquely different modeling works: (1) evaluating a classical GCM showing its limits caused by conventional convective scheme, (2) a direct new approach of examining global high-resolution model data like more nature, and (3) an indirect new approach with matrix linking inputs (environmental thermodynamic profiles) and outputs (convective tendencies) based on more idealized concept.

First of all, the limitation of classical GCM related to the conventional convective scheme is examined in chapter 2, which is published in Song and Mapes (2012). This chapter examines how unique climatology forecast data could be used and interpreted through the diagnosis of systematic error growth. Using the climatology forecast based on 20-year hindcast set from version 1 of the NCEP CFS, all of the animations of bias growth in all variables in all verification seasons are examined in http://www.rsmas.miami.edu/users/bmapes/CFS1biases.html. In chapter 2, one of the robust aspects involved in the convective error initiation ("CTIC" error) is analyzed. Main conclusions in chapter 2 are below:

1) The “Cold Tongue – ITCZ Complex (CTIC) error” is initiated by excessive convection in the low-latitude edge of the trade winds just upstream of the ITCZ at day 2 (Fig. 2.2). This rapid development can be interpreted as the
weak moisture sensitivity by the simplified Arakawa-Schubert scheme in CFS that vertically integrated buoyancy of weakly entraining air parcels has little sensitivity to free tropospheric dry air and shallow inversions.

2) After this initiated convective error, the CTIC error feedback is followed by interaction between errors of different components (e.g., convection, wind, heat content in the ocean, etc.) as differently saturated error rates. Atmospheric moist physics biases grow strong within a day (Fig. 2.2a) while stratocumulus cloud fields are inadequate even initially (Huang et al. 2007), and ocean model biases take months to fully saturate (Fig. 2.2c and d).

3) Through the interpretation of those error feedbacks in the atmosphere-land-ocean system, some interactive mechanisms between different atmospheric components in more nature-like framework could be suggested. It is not only shown in CTIC error feedback initiated by convective error in chapter 2, but also in other cases by different error time scale, slightly shifted thermal wind balance relationship, and errors that decay with lead time instead of increase or are saturated in appendix A.

Secondly, explicitly simulated convection in the high-resolution global model is analyzed in chapter 3. As a direct approach to treat this model data like observation, this analysis provides fairly realistic behavior of convection to evaluate and study. Using the NASA GEOS-5 simulation with a 5-km horizontal resolution by Putman and Suarez (2011), the eight heaviest rain events at 2.5 degree scale were selected and analyzed in both composite as well as individual case studies. 2.5 degree scale is selected for focusing
on the MCS scale as the linkage between convective activities and many significant larger-scale phenomena. There are two main conclusions below.

1) Composite profiles of temperature and humidity in GEOS-5 showed good agreement with observational composite studies (section 3.3.1), but not every case has all the composite characteristics (Fig. 3.24) because of diverse forcing mechanisms (section 3.3.2). Therefore, composite method could be used to get general characteristics – it is a good method, especially for observational data sets with the inevitable limitation of incomplete time or space sampling, but it should be careful if those characteristics will be applied to the individual cases.

2) Although full information about causality between convection and its environment could not be derived through this analysis, the spatial developments around the event time in individual case studies show fairly realistic behavior of convection according to the diverse forcing in the nature-like framework.

More detailed results of composite and individual case studies in GEOS-5 are summarized below.

1) Composite temperature, humidity, and vertical velocity profiles of eight heaviest rain events in 2.5 degree scale show general characteristics of MCS consistently compared with previous studies (Frank 1978, Sherwood and Wahrlich 1999, Mapes et al. 2006). Those general characteristics include cold tongue at 850 hPa ahead of the event time and three-layer patterns with two time jumps in both temperature and humidity (Johnson et al. 1999).
Individual case studies show realistically simulated MCSs according to diverse large-scale forcing mechanisms. These mechanisms include: lower boundary forcings (Amazon river and Panama Bight, case 1 and 4), the ITCZ and its mean confluence (concentrated along lines of longitude, case 7), intrusion of midlatitude troughs and their quasi-geostrophic lifting into the latitudes near 15°N (case 2 and 5), gravity waves lifting the “cap” of inhibition near 850 hPa (case 8), and coastal mountain diurnal convection (case 6). Such diverse mechanisms for MCS development are realistic, in that all are observed in nature.

At last, based on the linear algebra, significant modes of convection-environment interaction are analyzed, and this linear relationship is applied to diagnosis of the observation(-like) data as well as to a prognostic global model in chapters 4 and 5. Unlike chapter 3, this is an idealized and indirect approach, but provides a clever way to understand their interaction through the power of linearity for decomposability. All the examinations in chapters 4 and 5 are based on the linear convective response tendency matrix \( \mathbf{M} \) by Kuang (2012). \( \mathbf{M} \) was calculated in 128 × 128 km\(^2\) to the environmental small perturbations (\( T' \) or \( q' \)) by CRM.

In chapter 4.3 and 4.4, for examining significant modes of the convection-environment interaction, \( \mathbf{M} \) is analyzed in the eigen space. The major convective heating modes of MCS are captured, i.e., monopole (deep warming or cooling) and dipole (half warming and half cooling) modes. Therefore, linearity as a bit of idealized concept is still enough to catch the major convective heating modes of MCS that are already shown in
modeling (e.g., Mapes 2000) as well as the observation (e.g., Mapes and Houze 1995). The detailed results are followed.

1) The 1st mode of $M$ is the monopole mode, i.e., deep cooling (or deep warming) tendency to the positive (or negative) temperature perturbation (Fig. 4.6) in chapter 4.4.1. It shows that a cool troposphere and moist PBL are very rapidly damped by convection within $\sim 45$ mins (Fig. 4.5). Kuang called this the “CAPE damping” mode, but it acts oppositely when this mode is applied to the global model in chapter 5. It is discussed later.

2) The 6th & 7th mode of $M$ is the dipole mode, i.e., cooling (or warming) in upper half and mild warming (or cooling) in lower levels to the positive (or negative) temperature perturbation (Fig. 4.14) in chapter 4.4.3. They are only weakly damped ($\sim 6.7$ hours in Fig. 4.5). This pair has same real profiles with opposite-signed imaginary profiles (Fig. 4.12). Those imaginary parts bring the oscillating or descending evolution patterns with time in Fig. 4.15. These modes could be called the “congestus-deep oscillator” mode.

In chapter 4.5, this $M$ is applied to the GEOS-5 (from chapter 3) to diagnose how linearity calculated by CRM works on the observation-like data. Since causality between convection and its environment is difficult to interpret or infer in observations, it would be helpful for better understanding of causality between them if linearity will somehow work with observation. Three examinations are performed – both $T'$ and $q'$ (“$T_{\text{and}q}$”), only $T'$ (“$T_{\text{only}}$”), or only $q'$ (“$q_{\text{only}}$”) is multiplied by $M$ (Table 4.1). Even though the exact interpretation is not available, comparison between experiments provided the role
of each factor ($T'$ or $q'$) to the convection. Some diagnoses through comparison are followed.

1) Reconstructed convective tendencies in “$T_{and}q$” and “$q_{only}$” are similar in Fig. 4.19. Humidity ($q'$) seems to be the factor to decide the basic convective characteristics than temperature ($T'$).

2) According to the reconstructed rainrate, “$T_{only}$” shows that the cold pool effect at the surface (the heaviest rainrate at $t = +3$ hours in Fig. 4.20b) is more than compensated the cold tongue at 900-800 hPa (the slight rainrate dip at $t = -3$ hours in Fig. 4.20b). However, the effect of $T'$ is only half than both $T'$ and $q'$ or only $q'$ according to the magnitude of reconstructed convective tendencies in Fig. 4.19.

There was another examination to calculate the “linear response matrix” with observation (TOGA-COARE). Even though this calculation was not quite successful, some processes, results and discussions are included in appendix C. Since forcing and response is not separable in observation data, “linear regression matrix $\hat{\beta}$” is calculated in contrast to $\mathbf{M}$ by CRM. Although the method to calculate the matrix $\hat{\beta}$ is totally different from the $\mathbf{M}$, $\hat{\beta}$ and $\mathbf{M}$ occupy the same position in an equation expression an assumed linear relationship. As a fitting matter, $\hat{\beta}$ can reconstruct major features and patterns of convective factors (e.g., rainrate, heating or moistening tendency) reasonably (Fig. C.2c, Fig. C.5, and Fig. C.8), $\hat{\beta}$ itself is still not clearly interpretable. To get better relationship on $\hat{\beta}$, some efforts are done like minimizing the expected problems (e.g., overfitting) or simplifying the convective tendency patterns using the EOF analysis, but they were not quite successful. However, linearity is proved as the simple but essential
cornerstone to understand the interaction between convection its environment by many modeling works (Mapes 2004, Tulich and Mapes 2010, Kuang 2010). Therefore, these efforts to calculate $\beta$ with observation data would make a meaningful step to understand convection in the interactive large-scale framework.

In chapter 5, $\mathbf{M}$ is also applied to the global model to examine how linear convective responses work as a convective scheme in the large-scale framework. The global model used in this study (called “Jalopy with matrix”) separates the mean flow effects from large-scale weather variability, so it would be clearer than classical full physics GCMs to isolate mechanisms of convection-dynamics interaction. Five $\mathbf{Ms}$ modified by interesting modes are examined (Table 5.1) – without $\mathbf{M}$ (“$\text{NoM}$”), with original $\mathbf{M}$ (“$\text{FullM}$”), $\mathbf{M}$ without mid-level $q'$ (“$\text{NomidlevMoist}$”), $\mathbf{M}$ without 1$^{\text{st}}$ mode (“$\text{No1stmode}$”) and $\mathbf{M}$ without 6$^{\text{th}} & 7^{\text{th}}$ modes (“$\text{No6&7thmodes}$”). The main conclusion is that the expectation for the role of major eigenmodes of $\mathbf{M}$ does not quite happen when they are applied in the global model. However, differently modified $\mathbf{Ms}$ showed clear distinction among those examinations. Noticeable results are followed.

1) All experiments with $\mathbf{Ms}$ create tropical convective waves except “$\text{No1stmode}$” (Fig. 5.2). “$\text{No6&7thmodes}$” create much stronger eastward waves. Since we expected the strong damping effect by 1$^{\text{st}}$ mode, those contrast between “$\text{No1stmode}$” and “$\text{No6&7thmodes}$” are opposite than our expectation.

2) As it is mentioned in above, the 1$^{\text{st}}$ mode of $\mathbf{M}$ acts counter-intuitively when it interacts with environment like nature in the global model. It seems like acting as enhancing convection instead of damping it. This could be related to the key
mechanism of stratiform instability for convectively coupled waves to enhance convection (Mapes 2000), which is the in phase between anomalously cool and the cooling in the lower troposphere.

This is a beginning stage to attempt applying easily interpretable linearity to the nonlinear convective behaviors in the interactive framework. Although there are inconsistent behaviors of convection between individual convective modes (chapter 4) and their application in the global model (chapter 5), some rational diagnoses were available through the comparison between experiments, whether or not results were expected or unexpected. Therefore, using the linear relationship by elegantly calculated $M$ would be a challenge, but really a promising way to provide some insight for interactive mechanism between convection and its environment.
Appendix A: Interpretations of systematic errors in the NCEP CFS

After examining all of the animations of bias growth in all variables in all verification seasons, three robust aspects seemed understandable and interesting in light of our prior expectations about bias error growth with lead time. This chapter is about two of them except the “CTIC error” initiated by convective parameterization with NCEP CFS climatology forecast data in chapter 2. This appendix is based on Song and Mapes (2012).

A.1 A broader view of near surface temperature errors

This section expands the view from regional to global. Error growth with lead-time for T2m in DJF verification season is shown in Fig. A.1. Land errors have a tendency to grow quickly and saturate within a month or so, while ocean errors grow more gradually but keep growing for a longer time. This impression is quantified better in Fig. A.2, which indicates the lead time at which the systematic T2m error at each grid point first reaches 70% of its value at 256-day forecast lead. The area with final errors smaller than 0.2 K is masked out as white, and color is segmented into 3 groups: roughly a week, a month, and a season as blue, green and red, respectively. The land generally has week to month scales, while the main ocean basins have longer time scale like a season. Some tropical lands have longer time scales for error saturation (red patches), especially near coasts which may indicate the influence of adjacent oceans, but also inland which may involve deep soil moisture reservoirs.
Figure A.1 Error patterns of surface temperature (K) with lead time (a) 2 days, (b) 8 days, (c) 32 days, (d) 64 days, (e) 128 days, and (f) 256 days in the globe in DJF season.
Figure A.2 Lead-time (day) map at which error of surface temperature at each grid point reaches 70% of its value at lead time 256 days. The area with small errors from -0.2 K to 0.2 K is masked out (white) in DJF season. Color segmented into 3 groups: a week, a month, and a season as blue, green and red, respectively.

Over both land and oceans, the geographical pattern of systematic T2m error seems to grow in magnitude with almost fixed spatial structure. To quantify this impression of fixed structure, we define reference patterns $E_{\text{ref}}(x,y)$ by standardizing the spatial patterns of error and averaging these standardized patterns over all the lead times. We defined reference patterns for land and ocean areas separately (displayed as inset panels in Fig. A.3), as well as for the whole domain (an average of standardized versions of Fig. A.1’s panels, not shown). Latitudes are clipped 60°S to 60°N to avoid dominance by polar areas with many grid cells and large anomalies but little structure (as seen in Fig. A.1). Some of the fixed structure involves fairly small geographic features (Great Lakes, inland seas, etc.), but the nearly fixed structure is also true on broader scales.
Figure A.3 Amplitude coefficient, $A$ of the reference pattern with log scale lead-time (day) in 60°S to 60°N in DJF season. Each amplitude ($A$, thick) and residual ($R$, thin) coefficient is calculated by total (land+ocean, solid), land only (dashed), and ocean only (dotted) conditions. Inset maps are reference patterns times magnitude $A$ at 16 days in land and 64 days in the ocean respectively, with color range from -2 K to 2 K.

If the fixed structure hypothesis were exactly true, the pattern of systematic error $E(x,y,\text{lead})$ could be written as $E_{\text{ref}}(x,y) A(\text{lead})$, so that the amplitude coefficient $A(\text{lead})$ captures all the lead time dependence to the reference pattern $E_{\text{ref}}(x,y)$. In reality, there is a residual part $R(x,y,\text{lead}) = E(x,y,\text{lead}) - E_{\text{ref}}(x,y)A(\text{lead})$. The curve $A(\text{lead})$ which minimizes $R(x,y,\text{lead})$ was obtained by computing the spatial correlation of $E$ with the reference pattern $E_{\text{ref}}(x,y)$ at each lead time. Figure A.3 shows $A(\text{lead})$ curves for land (dashed), ocean (dotted), and the whole domain (solid). Down near the bottom are curves showing the spatial RMS of the unexplained residual $R(x,y,\text{lead})$. It is indeed small, indicating that the fixed structure hypothesis is well satisfied.
The different time scales of land and ocean error saturation are also quantified in Fig. A.3: the land error pattern grows in magnitude until a lead time 16 or 32 days, then levels out (saturates), while the ocean error pattern keeps growing in magnitude until a lead time of 64 days. Ocean temperature errors keep growing beyond 64-day lead time at low latitudes (Fig. A.2), so the fixed structure hypothesis is less well satisfied then. Also, for lead times longer than 64 days, the errors have grown in a previous season from the verification season, so perhaps those errors are more likely to have a different spatial pattern for that reason.

The land reference pattern has 2 main aspects: cold bias in high northern latitudes, and mostly warm biases elsewhere. The time scale for growth of the warm biases in the subtropics and midlatitudes (Europe, Australia, subtropical Africa) presumably involves soil moisture as the memory variable. A drying of the soil is also indicated also by low bias of latent heat flux or high bias of Bowen ratio, which develops on a similar, 1 month time scale (not shown, see Web animations).

The cold bias in Arctic latitudes also develops over weeks (16–32 days), which seems too long for the small sensible heat capacity of a frozen surface under a winter inversion. The long time scale suggests that some larger system with more thermal or mechanical inertia may be linked to these growing T2m biases. But the timescale is weeks, not months as might be the case if it were driven by the remote oceans.

One possibility for the week timescale of Arctic error growth is that the T2m errors are part of hemispheric-scale jet stream (polar vortex) dynamics via the thermal wind balance relationship. Inertia in the balanced winter vortex provides time scales of many weeks (e.g., Hurrell 1995, Thompson and Wallace 1998, 2000). There are also several
studies that winter flow can be predicted by surface temperature on Siberia in the preceding fall (Cohen and Entekhabi 1999, Cohen et al. 2001, Cohen and Fletcher 2007). Their proposed mechanism is that cold Siberian surface conditions drive upward stationary Rossby wave propagation to affect the jet stream. If such zonal eddy (continental enhanced) temperature biases are linked to zonal mean vortex strength, and if thermal wind balance further cools high latitudes with a continental bias, it may imply interesting eddy-mean feedbacks involving the surface. But with transient eddies also active (which cannot be diagnosed from climatological outputs) the full winter jet dynamics are beyond our ability to study in detail here.

To examine the thermal wind relationships around the Arctic winter cold bias, zonal mean temperature and wind error structures with height are shown in Fig. A.4 at 8, 32, and 256 day lead times. Again the fixed structure hypothesis seems to hold, this time in the vertical plane. Panels c and f correspond to the DJF error of runs initiated in the previous spring, so in that case the cold-season physical processes over land have time to fully express themselves in the DJF errors seen here and warm-season process errors could also play a role if their influences linger through boreal fall.
Figure A.4 Zonal mean error patterns of (left) temperature (K), and (right) geostrophic wind (m/s), with height in Northern Hemisphere in DJF season. Lead-times of (a, d) 8 days, (b, e) 32 days, and (c, f) 256 days respectively.

Surface temperature (T2m) error is assigned to 1000 hPa in Fig. A.4 for contouring purposes, while the rest of the tropospheric (virtual) temperature error is hypsometrically constructed from model geopotential outputs at 1000, 850, 700, 500, and 200 hPa. Geostrophic zonal wind is also shown at right. We find that zonal mean total wind ($u$) at 200 hPa (a model output) agrees well with the computed geostrophic wind ($u_g$) at that level, supporting the use of geostrophy at the other levels.

The surface (T2m) error does indeed appear to be seamlessly connected to deeper thermal wind balanced tropospheric errors, suggesting that they are part of a whole polar vortex error, not just a thin surface layer thermal bias trapped under the winter inversion.
The vortex error has zonal wind getting stronger to the north and weaker to the south of 40°N in an equivalent barotropic manner. The high-latitude cold error thus corresponds to an erroneous enhancement of the jet stream on its northern flank (red wedge on Fig. A.5). If the pattern in Fig. A.4 is interpreted in light of the two kinds of winter jets, subtropical Hadley driven and eddy driven midlatitude one (e.g., Lee and Kim 2003), the error features in Fig. A.5 may suggest a systematic weakening of the subtropical jet and enhancement of eddy driven jet with long forecast lead-times.

Figure A.5 Zonal mean zonal wind (m/s) at 200 hPa with latitude verifying in DJF season at lead-time 1 day (solid) and 256 days (dotted). Red and blue shading represent positive and negative error respectively.

This jet stream error in Fig. A.5 is relatively small in terms of the zonal wind itself, but the dynamics of Rossby wave ducting is sensitive to small changes in the background flow, such as the difference between January and February mean flow (Newman and Sardeshmukh 1998). Figure A.6 explores the effects of the zonal wind errors at 200 hPa (u200) on the ducting of stationary Rossby waves, based on the stationary wave number...
$K_s$ is a good indicator for main waveguides and preferred global wave propagation routes (Hoskins and Ambrizzi 1993). We follow their formula exactly using zonal wind at 200 hPa as the only input variable to calculate $K_s$. Shown in Fig. A.6, mean zonal wind patterns (Fig. A.6, left) are almost the same for both 1-day and 256-day forecasts.

![Figure A.6 Zonal wind patterns at 200 hPa (m/s) in the globe and stationary wave number $K_s$ patterns in the selected region (black box in the left panels) at lead-time (a, b) 1 day and (c, d) 256 days in DJF season.](image)

The small zonal wind errors in Fig. A.6 make a few significant differences to $K_s$ (Fig. A.6, right). The focused area of $K_s$ marked as black box in the zonal wind pattern panels. White shading areas represent zero or negative values of $K_s$, so it means stationary waves cannot propagate through these regions. In the long-lead forecasts, the east Pacific and Atlantic equatorial wave ducts seen in the analyses (Fig. A.6b) are blocked at almost all longitudes (Fig. A.6d). Meanwhile the observed wave barrier across the Mediterranean region instead has a slight propagation duct in the long-range forecasts. These $u_{200}$ errors
might be causing wave propagation path errors in the CFS forecasts, producing patterns of bias that cannot simply be subtracted off for anomaly forecasts.

The ultimate source of the winter vortex bias remains unclear. Surface thermodynamic process errors might play a role, but it is hard to separate them from other sources of error. On the one hand, cold anomalies under a stable boundary layer are very inefficient as a forcing for the jet stream compared to tropical SST influences (Inatsu et al. 2000). On the other hand, planetary-scale effects of Eurasian snow cover, via the surface temperature, pressure, and thus stationary waves (Cohen and Fletcher 2007, Hardiman et al. 2008, Alexander et al. 2010), suggest substantial upward propagation of surface effects. However, the shaping of jets by their own transient eddies and negative viscosity (e.g., McIntyre 2008) makes jet errors hard to interpret. It could be informative to change some snow or other parameters in CFS and see how that affects hindcast bias.

A.2 Errors that decay with lead time

Among all the systematic error growth patterns examined, the velocity potential ($\chi$) error patterns seemed most counter-intuitive, especially in summer season (JJA). Unlike most variables whose errors grow with lead-time until saturating (like the time series in Fig. A.3), the magnitude of $\chi$ grows initially but then decays for very long lead times (Fig. A.7, at 850 hPa). Also the fixed pattern hypothesis fails here: error patterns are more west-east for short leads and shift to more north-south for leads beyond 16 days.
Since χ is the inverse Laplacian of divergence, these quantities are related nonlocally, although linearly. In essence, χ can be viewed as a highly smoothed view of divergence. Positive values of χ error (indicating erroneous low-level convergence) tend to prevail over the land-rich parts of the world (the Eastern and Northern hemispheres), perhaps linked to CFS’s warm land biases. The error in Fig. A.7 saturates in 16–32 days, similar to the time scale of land temperature error saturation (Fig. A.3), but the link is not simple since χ responds to condensation heating not just surface temperature or sensible heating. For example the peak positive core of χ in Figs. A.7d–f over Maritime Continent longitudes corresponds to positive monsoon precipitation errors (not shown), not just sensible heating over hot land surfaces.

The decay of χ errors with lead times beyond 64 days is presumably due to a compensating bias growth in the ocean model, based on the long time scale. The exact
nature of these ocean biases, and whether the compensation is fortuitous or a systematic negative feedback on the $\chi$ errors themselves, is impossible to diagnose in detail with the data available. Broadly speaking, ocean biases are mostly cold and increasing with lead-time just as in the DJF case (Fig. A.1). This would not compensate warm land biases in a simple sense, but again $\chi$ reflects (an inverse Laplacian of) heating rate, not the surface temperature field. For example, the core positive 32–64 day forecast $\chi$ errors in Figs. A.7e–f decay along with the monsoon precipitation errors in those longitudes, because the model’s rainy bias moves to the western Indian Ocean with emergence of an east-to-west T2m gradient bias resembling the Indian Ocean Dipole (Saji et al., 1999) in those longer-lead forecasts (not shown). In short, since $\chi$ is nonlocally related to a nonlocally controlled process (heating rate) it is hard to interpret. In any case its sign reversal for lead times beyond 64d is understandable, despite seeming counter-intuitive initially, in light of the different land vs. ocean model error growth times seen in Fig. A.3.
Appendix B: Weighted composite vs. regression coefficient

Composite $\overline{x(z,t_0)}$ is the averaging process of all cases over the time and vertical height dimensions. The composite is calculated as,

$$\overline{x(z,t_s)} = \frac{1}{N} \sum_{i=1}^{N} x_i(z,t_0 + t_s), \quad (B.1)$$

where $i$ is the case number, $N$ is the total number of events, and $x$ is the variable of interest. $z$ is the vertical array (37 layers) and $t_s$ is sampled time array from -18 hours to +18 hours at the event time ($t_0$) in each event $i$. The overbar means the event-averaged quantity. $T$ and $qv$ as the $x$ values in all composite studies are time-anomaly values at each vertical height. Composite of $\omega$ uses original values for keeping the upward and downward signals.

Weighted-composite $\overline{x_w(z,t_s)}$ is same as Eq. B.1, but weighted by the maximum precipitation value in each case,

$$\overline{x_w(z,t_s)} = \frac{\sum_{i=1}^{N} y_i(t_0)x_i(z,t_0 + t_s)}{\sum_{i=1}^{N} y_i(t_0)}, \quad (B.2)$$

where $y_i(t_0)$ is the maximum precipitation at event time ($t_0$) in each event $i$. Equation B.2 can be also expressed as below,

$$\overline{x_w(z,t_s)} = \frac{y(t_0)x(z,t_0 + t_s) + y'(t_0)x'(z,t_0 + t_s)}{y(t_0)}$$

$$= \frac{x(z,t_0 + t_s)}{y(t_0)} + \frac{y'(t_0)x'(z,t_0 + t_s)}{y(t_0)} = \frac{\text{cov}(y(t_0),x(z,t_0 + t_s))}{y(t_0)}, \quad (B.3)$$

where $\text{cov}( )$ is the covariance, and primes represent anomalies. Covariance is calculated over the all events at every time steps ($t_s$ from -18 to +18 hours) and vertical levels ($z$).
The weighted-composite maintains almost the same pattern as the non-weighted composite, but with enhanced magnitude by the second term of the right hand side in Eq. B.3.

The weighted-composite method is often comparable to the regression method. The regression method is the way to calculate regression coefficient, and its relationship to the weighted composite is the subject of this Appendix. The regression coefficient is the best slope, $b$, for the least squares fit in a linear relationship, $\bar{y} = b\bar{x} + a$. $b$ for precipitation ($y$) and another variable ($x$) is calculated as below,

$$b(z,\tau) = \frac{\text{cov}(y(t),x(z,t+\tau))}{\text{var}(y(t))} = \frac{\sum_{t} y'(t)x'(z,t+\tau)}{\sum_{t} y'(t)^2},$$  \hspace{1cm} (B.4)

where, $y(t)$ is the precipitation corresponding to time $t$ with another variable, $x$ as $T$ or $qv$. $t$ is all time period and $\tau$ is the time lag according to the purpose of examination. For the comparison with weighted-composite, $\tau$ is selected from -18 hours to +18 hours like sampled time array $t_s$ in Eqs. B1 and B.3. While selected events ($is$) with sampled time array ($t_s$) are used for weighted-composite (Eq. B.3), all time steps with time lag ($\tau$) become events for regression coefficient (Eq. B.4). Therefore, $t_0$ and $t_s$ in Eq. B.3 become all time step $t$ and time lag $\tau$ in Eq. B.4 respectively.

Although every time steps are counted as events for regression coefficient calculation, this defines “events” as so relaxed meaning that any nonzero anomaly of precipitation $y'(t)$ is selected as the “event”. Many large numbers of weak events for regression coefficient calculation don’t dilute the signal because they are assigned little weight. In addition, according to the comparison of (B.4) with (B.3), the regression coefficient can be viewed as an anomaly version of the weighted-composite owing to the absence of the
averaged quantity of the variable $\overline{x(z,t_0 + t)}$. It is confirmed by similar patterns between weighted-composite and regression coefficient ((b) and (c) in Fig. B.1 and B.2).

(a) Non-weighted $T'$

(b) Weighted $T'$

(c) Regression $T'$ (0.3 K per mm/hr)

Figure B.1 (a) Composite, (b) weighted-composite, and (c) regression coefficient of time-anomaly temperature ($T'$) at each vertical height. Unit is [K]. X-axis is time centered at the event time $t=0$ [hr], and y-axis is vertical height [hPa]. The red line in (c) is the auto-correlation of precipitation.
Figure B.2 Same as Fig. B.1, but specific humidity ($q v'$). Unit is [g/kg].

Vertical velocity $\omega$, however, shows a different pattern when the regression technique is used compared to the other two techniques (Fig. B.3c). Regression patterns away from the event time (before and after around ± 6 hours) have the opposite sign as well as different vertical patterns compared to the other techniques. This is partly caused by the
sign of precipitation anomaly for calculation that $y_i(t_0)$ for covariance in (B.3) is always positive, while $y'(t)$ in (B.4) can be either signs, especially negative away from the event time. This is also partly cause by the using either anomaly ($T$ and $qv$) or original data ($\omega$). If anomalies (i.e., $T$ and $qv$) are used for weighted-composite calculation (Eq. B.3), $x(z, t_0 + t_s)$ becomes zero, and in the consequence, Eqs. B.3 and B.4 show identical formula except different definition of events. As it is already mentioned above, compared with weighted-composite calculation, every time step events for regression coefficient calculation don’t affect to change signals very different because of little weight by many weak events. On the other hand, if original data (i.e., $\omega$) is used for weighted-composite calculation, $x(z, t_0 + t_s)$ is not zero. Therefore, the different regression patterns and signs of $\omega$ compared with weighted-composite in Fig. B.3 are affected by two facts: 1) precipitation anomalies ($y'(t)$) for regression coefficient could be either positive or negative sign, and 2) using original data ($\omega$) keeps the $x(z, t_0 + t_s)$ term that makes Eqs. B.3 and B.4 different.
In summary, composites and weighted-composites can be interpreted as the same quantities. For regression patterns to be analogous to composites, anomalies must be used in the composite calculation.
Appendix C: Linear regression matrix beta $\hat{\beta}$

The linear regression matrix $\hat{\beta}$ is calculated to show the relationship between precipitation and environmental temperature and moisture anomalies in observation instead of the cloud-resolving model. However, observation and modeling work have many distinctions, especially observation has no forcing and response because all convective (precipitation) and environmental (temperature and moisture) variables are already known. Therefore, we cannot expect to see the causality between convection and environments, but only their relationship with observation. In the consequence, the method to calculate the matrix $\hat{\beta}$ is also totally different from the linear response matrix ($M$) in Kuang (2010, 2012), but $\hat{\beta}$ and $M$ occupy the same position in an equation expression an assumed linear relationship. Therefore, $\hat{\beta}$ and $M$ should have similar values if convection in nature is like in a periodic CRM and if there is enough data to reliably estimate $\hat{\beta}$ by regression – but there are not. This is the motivation for the material in this appendix.

TOGA-COARE at Outer Sounding Array (OSA) is used for calculating $\hat{\beta}$. Spatial array of OSA is from 147°E to 167°E and 10°S to 10°N as seen in Fig. C.1. Time array is 480 values – 120 days with 6 hourly interval – from November 1992 to February 1993 (Webster and Lukas 1992), and 33 vertical levels from 1000 to 200 hPa with 25 hPa interval are used.
The linear regression coefficient matrix $\hat{\beta}$ is expressed as below,

$$Y = X\hat{\beta} + \text{residual},$$  \hspace{1cm} (C.1)

where $Y$ is predictand and $X$ is predictor matrix. Since $\hat{\beta}$ is calculated with observation, residuals are minimized by the ordinary least square method as below,

$$\hat{\beta} = (X^TX)^{-1}(X^TY).$$  \hspace{1cm} (C.2)

Unlike cloud resolving model, heating ($Q_1$) and moistening ($Q_2$) tendencies or rainrate ($R$) in observation contain other effects (e.g., large-scale advection) as well as convection. Moreover, there are no clear “forcing” and “response” terms because of the mixed dynamics and feedback in observation. Nevertheless, “forcing” and “response” terms will be used in this chapter for convenience because convection representatives like $R', Q_1'$
and $Q_2'$ are placed in predictand matrix $Y$, and environment representatives like $T'$ and $q'$ in predictor matrix $X$. In short, $\hat{\beta}$ still can show the convective responses from the small change of environments.

**C.1 $\hat{\beta}$ on raw data basis**

First of all, two $\hat{\beta}$s are simply calculated with raw data of both $T'$ and $q'$ – one is for $R'$ and the other is for $Q_1'$ and $Q_2'$. $\hat{\beta}$ for $R'$ is calculated with $T'$ and $q'$ together, but drawn separately in Fig. C.2a and 2b because of different unit range. And $R'$ is reconstructed by $\hat{\beta}$ for $R'$ (red line in Fig. C.2c), and compared with the original $R'$ (black line in Fig. C.2c). While $R'$ is reasonably reconstructed by $\hat{\beta}$ in Fig. C.2c, the linear regression profile $\hat{\beta}$ itself is too noisy to be interpreted physically (Fig. C.2a and b). The profiles in $\hat{\beta}$ are still noisy if $R'$ is replaced by $Q_1'$ and $Q_2'$ in Fig. C.2d. Although it seems to reconstruct $R'$ reasonably (Fig. C.2c), it is not evidence for the validity of that assumed linear form. Fitting is just minimizing the RMS of fitting errors, and that could be done using many arbitrary functional forms, e.g., a linear form with many tunable coefficients in a matrix. In other words, these fitting coefficients would perform very poorly on independent data. However, there are some possibilities to improve the noisy linear regression profile $\hat{\beta}$ with some adjustments, and they were performed in the rest of this appendix.
Figure C.2 Both (a) and (b) is the linear regression profile ($\hat{\beta}$) for $R'$ to the $T'$ and $q'$ respectively – they are one profile, but drawn separately because of different unit range. (d) is $\hat{\beta}$ for $Q_1'$ and $Q_2'$ to the $T'$ and $q'$. Vertical axis in (a), (b), and (d) is response layer (hPa). Since $Q_1'$ and $Q_2'$ are vertical profiles unlike $R'$ which is vertically accumulated value, horizontal axis in (d) is forcing layer (hPa). Both forcing and response layers are from 1000 to 200 hPa with 25-hPa interval. (c) is the $R'$ with 6-hourly time interval. Black line is original $R'$ and red line is reconstructed $R'$ by $\hat{\beta}$ of (a) and (b). $R'$ unit is [mm/d].

One possibility of noisy linear regression profile $\hat{\beta}$ is the overfitting problem. Extracting too many regression coefficients in limited data can cause overfitting. Predictors ($X$ with $T'$ and $q'$ in Eq. C.1) with too close consecutive vertical level (25 hPa interval) could be interdependent, or minor fluctuations of less important variables (others than $T'$ and $q'$) could be exaggerated. To reduce this problem, predictors ($T'$ and $q'$ in $X$) are rebinned for 4 vertical levels and used for calculating $\hat{\beta}$ in chapter C.2 and EOFs modes of predictands ($Q_1'$ and $Q_2'$ in $Y$) are also used for clarity of interpretation in chapter C.3.
C.2 $\beta$ on the data with reduced vertical level

33 vertical levels are rebinned as 4 layers - planetary boundary (~900 hPa), lower (~750 hPa), middle (~400 hPa), and upper (~200 hPa) levels. It is roughly separated by lag regression patterns of specific humidity with rainrate based on the Mapes et al. (2006) in Fig. C.3. Vertical thickness changes with ± 50 hPa are not sensitive for results. $T'$ and $q'$ in $X$ in equation C.1 and C.2 are also rebinned as 4 layers, but $Q_1'$ and $Q_2'$ in $Y$ still keep 33 levels because predictands are not affected by overfitting problem.

![Figure C.3 Lag regression of specific humidity with rainrate in TOGA-COARE in IFA area (Fig. 14b in Mapes et al. 2006). Layers are separated by 4 layers in planetary boundary (~900 hPa), lower (~ 750 hPa), middle (~ 400 hPa), and upper (~ 200 hPa) levels.

$\hat{\beta}$ for $Q_1'$ and $Q_2'$ to the $T'$ and $q'$ with 4 layers is in Fig. C.4. One noticeable feature captured in this matrix is the cooling tendencies ($Q_1' < 0$) above the layer of the positive $T'$ at PBL, lower and middle layers. It seems to be similar to the convective inhibition effect of unorganized convection by Kuang (2012) that “an anomalously warm layer forms a buoyancy barrier that eliminates certain convective updraft parcels, causing cooling at and above the layer.” These consist of the convective inhibition (CIN) layer in Tulich and Mapes (2010) and Kuang (Fig. 12 in 2010) depending on the forcing layer of
While Tulich and Mapes (2010) showed that the “effective inhibition layer for deep convection is about 4 km deep, far deeper than traditional convective inhibition defined for undiluted lifted parcels”, Fig. C.4 shows even far deeper and stronger cooling tendency from the higher (middle layer from 750 to 400 hPa) than 4 km (deep blue in third column of upper-left quadrant in Fig. C.4).

The strong convective tendencies to the $q'$ in the upper layer (redish last column for both $Q_1'$ and $Q_2'$ from $q'$ in Fig. C.4) are not quite realistic responses because it is already very dry in the upper layer. Therefore, the convective responses to $1\text{g/kg}$ of $q'$ may not happen in usual.

![Figure C.4 The linear regression profile matrix $\hat{\beta}$ for heating ($Q_1'$) and moistening ($Q_2'$) tendency anomalies to the temperature ($T'$) and moisture ($q'$) anomalies.](image)

Reconstructed $Q_1'$ and $Q_2'$ profiles by $\hat{\beta}$ are compared with the original one in Fig. C.5. While reconstructed $Q_1'$ profiles (Fig. C.5b) capture the upper-level heavy heating or
cooling patterns quite well, reconstructed \( Q_2' \) profiles (Fig. C.5d) cannot capture the complex moistening tendencies like original one (Fig. C.5c).

Figure C.5 (a) and (c) are original TOGA-COARE \( Q_1' \) and \( Q_2' \) profiles with time respectively. (b) and (d) are same as (a) and (c), but reconstructed profiles by \( \hat{\beta} \). Horizontal axis is time with 6-hour interval and vertical axis is vertical level [hPa].
C.3 $\hat{\beta}$ on the EOF basis with reduced vertical level

For clarity of interpretation, only major patterns of predictand ($Q_1'$ in $Y$) are selected by EOF and used for calculating $\hat{\beta}$. Predictors ($T'$ and $q'$ in $X$) still keep 4 vertical-layer values like in chapter C.2.

Predictand’s matrix $Y$ can be expressed with EOFs modes as below,

$$Y = PE,$$

where $P$ is time series variability (480 time array) and $E$ is basis functions (eigenvectors or EOFs modes here, 33 modes) matrices. 1$^{\text{st}}$ and 2$^{\text{nd}}$ EOFs modes are in Fig. C.6. 1$^{\text{st}}$ mode captured the deep convective heating mode, which is the upper-level heavy heating (or cooling) pattern with same sign through the whole vertical layer (Fig. C.6a). And 2$^{\text{nd}}$ mode captured the stratiform mode, which upper half and lower half layers have an opposite sign (Fig. C.6b).

![Figure C.6](image)

Figure C.6 (a) 1$^{\text{st}}$ and (b) 2$^{\text{nd}}$ EOF modes of $Q_1'$ in TOGA-COARE. Each percent variance is 80.7% and 7.6% respectively.
Two modes captured the major convective heating modes well and the total variance is about 90% (1\textsuperscript{st} mode represents 80.7% and 2\textsuperscript{nd} mode does 7.6%). Therefore, these two modes of $Q_1'$ are used as the predictands in $Y$ in this chapter. $Y'$ for two EOFs modes is reconstructed as below,

$$Y' = PE',$$

where $E'$ has only 2 modes (1\textsuperscript{st} and 2\textsuperscript{nd} modes) instead of 33 modes.

for major EOFs modes of $Q_1'$ to $T'$ and $q'$ is in Fig. C.7. It reflects the top-heavy (upper two quadrants in Fig. C.7) and two-sign heating (lower two quadrants in Fig. C.7) responses based on 1\textsuperscript{st} and 2\textsuperscript{nd} EOFs modes of $Q_1'$ in Fig. C.6 respectively. The deep convective heating mode (= 1\textsuperscript{st} mode of $Q_1'$) would participate the inhibition effect of $\hat{\beta}$ in Fig. C.4 as it shows the deep cooling tendency to the $T'$ in middle layer (deep blue in third column in upper-left quadrant in Fig. C.7).

![Figure C.7](image-url)  
Figure C.7 Same as Fig. C.4, but $\hat{\beta}$ for 1\textsuperscript{st} and 2\textsuperscript{nd} EOFs modes of $Q_1'$ in TOGA-COARE.
Reconstructed EOFs modes of $Q_i'$ by $\hat{\beta}$ are compared with original one in Fig. C.8. Since 1$^{st}$ EOFs mode of $Q_i'$ shows 80.7% variance, the most of the original $Q_i'$ (Fig. C.8a) patterns can be explained by 1$^{st}$ EOFs mode of $Q_i'$ (Fig. C.8b). These top heavy heating (or cooling) patterns are reconstructed reasonably (Fig. C.8c) with less magnitude and without higher-frequency variability. In the case of 2$^{nd}$ EOFs mode, the signal patterns (positive/negative two-sign vertical patterns) are roughly reconstructed, but cannot reconstruct the detailed lower frequency variability.
Figure C.8 (a) is the original TOGA-COARE $Q_1'$ profiles with time. (b) and (d) are 1st and 2nd EOFs mode of $Q_1'$ respectively. (c) and (e) are same as (b) and (d), but reconstructed profiles by \( \hat{\beta} \). Horizontal axis is time with 6-hour interval and vertical axis is vertical level [hPa]. Unit of $Q_1'$ is [J/kg/s].

C.4 Summary

Unlike controlled cloud model experiments which can isolate the convective responses from the environmental forcing, responses and forcing in observation are not separable because of the mixed feedback. Nevertheless, a linear regression matrix \( \hat{\beta} \) which plays M's role in an identical equation can be estimated. As a fitting matter, it can reconstruct major features and patterns of rainrate ($R'$, Fig. C.2c), $Q_1'$, or $Q_2'$ (Fig. C.5 and C.8) quite well, but would probably perform poorly on out-of-sample data (like the DYNAMO sounding array, or on samples from COARE withheld from the fitting in a bootstrap exercise). This is known as the overfitting problem, and can be minimized by reducing the dimensionality of the data, for example using the only major convective modes by EOFs, etc.
Besides being noisy, linear regression matrix $\hat{\beta}$ is also not clearly interpretable. Should it give the same answer as $M$? To make the question sharper: Could we infer $M$ through a regression exercise from the output data of the matrix jalopy model (Chapter 5)? In Jacobian form with partial derivative notation, $M$’s quadrants are like $\partial Q_1 / \partial T$ with $\omega$ held constant at $\omega=0$ by the periodic boundary conditions on divergent wind. Meanwhile the corresponding ‘partial derivative’ in the regression matrix $\hat{\beta}$ does not hold $\omega$ constant, unless $\omega$ is used as another independent predictor in the regression. There is surely not enough data of sufficient quality to discover slight partial dependences of convection on $T$ and $q$ in the presence of the poorly-measured but highly convection-correlated predictor $\omega$.

For the future tasks, a synthetic data exercise to try and estimate $M$ by linear regression of Jalopy-with-matrix model output (Chapter 5) would be revealing. Can $M$ be inferred from output of a coupled system of $M$ and large-scale dynamics? If so, how much data is required for a given precision in reconstructing $M$?
REFERENCES


Cohen, J., C. Fletcher, 2007: Improved skill of northern hemisphere winter surface
temperature predictions based on land–atmosphere fall anomalies. *J. Climate*,

Cohen, J., K. Saito, D. Entekhabi, 2001: The role of the Siberian high in northern

Cronin, T. W., 2012: The role of the diurnal cycle in tropical land-ocean contrasts,

Crook, N. A., M. W. Moncrieff, 1988: The effect of large-scale convergence on the
generation and maintenance of deep moist convection. *J. Atmos. Sci.*, 45,

Del Genio, A. D., 2012: Representing the sensitivity of convective cloud systems to
tropospheric humidity in general circulation models. *Surv. Geophys.*, 33,


Donner, L. J., 1993: A cumulus parameterization including mass fluxes, vertical
momentum dynamics, and mesoscale effects. *J. Atmos. Sci.*, 50,

including mass fluxes, convective vertical velocities, and mesoscale effects:
Thermodynamic and hydrological aspects in a general circulation model. *J. Climate*, 14,
3444-3463, http://dx.doi.org/10.1175/1520-

Emanuel, K. A., J. D. Neelin, C. S. Bretherton, 1994: On large-scale circulations in
convecting atmospheres. *Quart. J. Roy. Meteor. Soc.*, 120,

generated stratospheric gravity waves. *J. Atmos. Sci.*, 49,


circulation models to capture the effects of Eurasian snow cover on winter

Herman, M. J., Z. Kuang, 2013: Linear response functions of two convective

Hoskins, B. J., T. Ambrizzi, 1993: Rossby wave propagation on a realistic longitudinally

Rev.*, **105**, 1540-1567, http://doi.org/10.1175/1520-


Atlantic Basin from coupled climate hindcasts. *Clim. Dynam.*, **28**, 661-682,

P. Bowman, E. F. Stocker, 2007: The TRMM Multisatellite Precipitation Analysis
(TMPA): Quasi-global, multiyear, combined-sensor precipitation estimates at fine

temperatures and precipitation. *Science*, **269**, 676-679,
http://doi.org/10.1126/science.269.5224.676.

schemes on the ENSO-like phenomena simulated in a CGCM. *J. Meteor. Soc.

Inatsu, M., H. Mukougawa, S.-P. Xie, 2000: Formation of subtropical westerly jet core in

Johansson, Å., C. Thiaw, S. Saha, 2007: CFS retrospective forecast daily climatology in

Johnson, R. H., T. M. Rickenbach, S. A. Rutledge, P. E. Ciesielski, W. H. Schubert,
1999: Trimodal characteristics of tropical convection. *J. Climate*, **12**, 2397-2418,


