Examining the Form-Function Relationship of Convective Organization and the Larger Scale with Observations and Models

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EXAMINING THE FORM-FUNCTION RELATIONSHIP OF CONVECTIVE ORGANIZATION AND THE LARGER SCALE WITH OBSERVATIONS AND MODELS

By

Emily Marie Riley

A DISSERTATION

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Doctor of Philosophy

EXAMINING THE FORM-FUNCTION RELATIONSHIP OF CONVECTIVE
ORGANIZATION AND THE LARGER SCALE WITH OBSERVATIONS AND
MODELS

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This work uses a two-pronged approach to study the form-function relationship of convective organization and the larger scale. Form is simply the visual shape of convection and function is how the convection and larger scale interact. First, CloudSat observations are used to study cloud modulation during the Madden-Julian Oscillation (MJO). Second, a cloud systems resolving model (CSRM) with parameterized large-scale dynamics is used to examine how convective organization affects the interdependence of convection and the larger-scale.

Using CloudSat observations, cloud type, total cloud cover, and temperature and moisture evolution are documented across MJO phases. Deep cloud types were classified as wide or narrow as a proxy for designating organized and unorganized convective systems. For locally defined phases, the MJO exhibits a familiar progression of cloud types from shallow clouds mixed with deep, isolated convection in the growing stages of the MJO, to deep, widespread, organized convection during the mature stages, to more anvil-dominated conditions during the decay stages. Comparison to the convectively coupled Kelvin wave reveals both wave types exhibit similar cloud type evolution,
though, the MJO was found to be modulated more by moisture variation, while the Kelvin wave was modulated more by temperature variations.

In terms of globally defined MJO phases, the wide deep precipitating systems were modulated more than other cloud types by MJO phases, with the well-known progression of cloud cover from the Indian Ocean to the central Pacific. The narrow deep precipitating systems only propagated from the Indian Ocean to the Maritime Continent.

The modeling component of this work involved periodic domains, where convective organization was controlled by adding shear to a three-dimensional (3D) isotropic CSRM domain or by altering the 3D domain to be longer and narrower, until eventually becoming a 2D domain. Snapshots of convective activity in various domains reveal the 3D isotropic domain has scattered, unorganized convection, while the addition of shear leads to squall lines, and the increasingly 2D domains lead to widely spaced infinite lines (in the periodic sense) of convective blobs. When the CSRM domain mean state is coupled to a large-scale wave equation, the coupled CSRM-wave system responds to the change in organization. The shear simulations are very similar to the no-shear simulations, suggesting the convectively coupled wave is indifferent to organized convection. However, wave amplitude at equilibrium increases as the domains are stretched from 3D to 2D, though only up to a certain xy aspect. Why a particular xy aspect gives the most responsive domain remains a mystery despite many sensitivity and forcing experiments (described in appendices here). Nonetheless, this framework for studying the convective, large-scale system offers uniquely tractable and controllable ways forward to understanding the multi-scale nature of atmospheric convection. The interdependence of convective organization and the larger-scale should help guide the
future of cumulus parameterization and/or other options for representing the convective scale in general circulation models (GCMs; e.g. superparameterization techniques).
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CHAPTER 1: Introduction

The goal of this dissertation is to better understand the interaction between convective organization and the larger-scale, within the framework of the form-function relationship of the two. Form is defined as “the visual shape or configuration of something” and function as “a thing dependent on another factor or factors” (New Oxford American Dictionary). The form of convection can be described as blobs, clumps, lines, scattered, organized, etc., while function relates to the interdependence of convection and its environment. While convection is often framed as a function of or response to the larger-scale (i.e. QE thinking), the two are sometimes so tightly coupled in observations (e.g. CCWs; Kiladis et al., 2009) it is difficult, if not impossible, to assess cause and effect with observations. The working hypothesis is that the form of convection affects the convective large-scale relationship – specifically, that more organized convection couples more strongly to the larger-scale. The first aim of this work is to describe convection within larger-scale variability using observation. The second aim is to isolate the effects convective organization has on convective, large-scale interactions using a cloud systems resolving model.

A two-pronged approach, with observations and models, is used to reach the goal stated above. The core motivation for this work is that moist atmospheric convection is observed to vary over a wide range of time and space scales – from short lived individual deep convective cells, to long lived mesoscale convective systems, to planetary scale intraseasonal oscillations – (e.g. Houze 2004), yet general circulation models (GCMs) fail to represent all those scales (e.g. Arakawa 2004, Del Genio 2012,). Namely, convection
organized on the mesoscale is neither present in cumulus parameterization schemes, nor resolved in most GCMs. Therefore, this dissertation seeks to understand how the degree of convective organization affects the interdependence of convection and the larger-scale.

This dissertation begins with an elaboration of the statement of the problem mentioned above and is followed by a background on the fundamentals of moist convection, a brief overview of the history of cumulus parameterization, and discussion of factors that help organize convection. (Throughout the dissertation “convection” will refer to moist atmospheric convection.)

1.1 Statement of problem

Moist convection is important for the redistribution of heat, moisture, mass, momentum, and energy, as well as, precipitation and boundary layer modification. Observations show that moist convection has a propensity to organize into coherent cloud field structures on the mesoscale (O(100s) kms). Individual convective cells are generally short lived and tend to lose their identity as they merge upscale to form mesoscale convective systems (MCSs; Chen et al. 1996). MCS are inherent in almost all tropical climatology including the Intertropical Convergence Zone (ITCZ), El Nino Southern Oscillation (ENSO), monsoon systems, the Madden-Julian Oscillation (MJO), convectively coupled waves (CCWs), and tropical cyclones. MCSs provide over 50% of the total precipitation to most tropical areas (Nesbitt et al. 2006) and their abundance is readily recognizable from satellite images by their extensive cirrostratus cloud shield.
Mapes et al. (2006) argue that the MCS lifecycle/structure is the fundamental unit of convective organization. They hypothesize that the MCS’s evolution (in space or time) – from shallow and congestus convection, to deep convection, followed by stratiform rain and anvil cloud shields – is mimicked across longer time and space scales forming three “building blocks” of convective organization. The building blocks are not simply aliased across longer time and space scales, rather their relative frequency of occurrence shifts with time and space. For example, Riley et al. (2011) showed that shallow convection was relatively more prevalent before active phases of the MJO. Likewise, deep convection and anvil clouds were relatively more prevalent during active phases and after active phases, respectively.

MCSs serve as a critical link between convection and the larger-scale atmospheric circulation. Distinct diabatic heating profiles associated with MCSs sculpt how the large-scale environment responds to convection (Houze 2004). Like throwing a rock into a pond, a spectrum of gravity waves emanate from the latent heat release of condensation during an MCSs’ life cycle. Two dominant modes dictate the environmental response to convective forcing: The deep convective component of MCSs produces a first baroclinic mode (vertical wavelength twice the depth of the troposphere) with heating that peaks in the mid-troposphere, while the stratiform component produces a second baroclinic mode (vertical wavelength the depth of the troposphere) with peak heating in the upper troposphere and cooling below. By mass continuity, the deep baroclinic mode requires net convergence at low levels and divergence aloft, while the second baroclinic mode requires a convergence maximum in the middle troposphere with divergence below and above (Fig. 3 in Houze 1997). Horizontal variations in the convective vs. stratiform
rainfall fraction within MCSs across the tropics lead to large-scale horizontal variability in vertical heating and motion profiles across the tropics (Schumacher et al. 2004).

Despite the wealth of MCS observed in nature and their influence in convective, large-scale interactions, the precise sensitivity of convection to large-scale variations and visa versa are still poorly understood. In most general circulation models (GCMs), MCSs are neither resolved nor present in the formulation of convective parameterization schemes (Del Genio 2011). This assumed scale gap artificially severs the continuum of convective organization observed in reality and remains a major challenge to meteorology. To move forward, convective organization must be permitted or parameterized in model formulations (Del Genio 2011).

The aim of this dissertation is to better understand the role convective organization has in the coupling of convection and the larger-scale and what that means for future cumulus parameterization schemes. Both observations and models will be utilized to address this overarching goal; observations to emphasize the importance of convective organization and models to tease apart the links between convection and the larger-scale.

1.2 Background

1.2.1 Fundamentals of moist convection

Moist convection is fundamentally based on buoyancy, and hence deeply rooted in the stability of the troposphere. Under strict stability analysis, an infinitesimal
perturbation that grows exponentially with time is considered unstable (Schultz et al. 2000). For the troposphere, vertical profiles of temperature (T) and moisture (q) dictate the stability of perturbations.

*Absolute instability* occurs when the environmental lapse rate is greater than the dry adiabatic lapse rate (or $d\theta/dz < 0$), but is rarely observed in nature since the instability is eliminated almost as quickly as it occurs. *Absolute stability* occurs when the environmental lapse rate is less than the moist adiabatic lapse rate (or $d\theta_{es}/dz > 0$).

*Potential instability* (or convective-instability) occurs when $d\theta_{es}/dz < 0$ and describes an environmental profile that is currently stable, but near an unstable state. For instability to be realized a layer of the troposphere must be lifted adiabatically such that the bottom of the layer saturates first and cools at the moist adiabatic lapse rate, while the top continues to cool dry adiabatically.

Textbook definitions of *conditional instability* (CI) state the troposphere is conditionally unstable when the environmental lapse rate is less than the dry adiabatic lapse rate, yet greater than the moist adiabatic lapse rate (or $d\theta_{es}/dz < 0$; Wallace and Hobbs 2005). Sherwood (2000) and Schultz et al. (2000) have scrutinized this textbook definition. Both papers recognize that $d\theta_{es}/dz < 0$ does not necessarily guarantee an instability, since the instability is conditioned on the parcel saturating and theta-es does not account for moisture variations. CI is perhaps better viewed in terms of parcel instability. The AMS glossary provides one straightforward definition: “A parcel of air at environmental temperature is unstable to vertical displacements if it is saturated, but stable to small vertical displacements if unsaturated.” By this AMS definition, CI cannot be considered a statement about a real instability. Rather, CI should be considered a
statement of stability uncertainty, since a finite amplitude displacement is needed for a
parcel to saturate (Sherwood 2000 and Schultz et al. 2000).

Two useful indices when considering parcel instability are CAPE (convective
available potential energy) and CIN (convective inhibition). CAPE is a measure of how
much kinetic energy a parcel could gain if the buoyancy force were unopposed, and can
be defined as:

\[ CAPE = \int Bdz = \int_{LFC}^{EL} g \frac{T_v - T_e}{T_w} dz, \]  

(1.1)

where LFC and EL represent the level of free convection and the equilibrium level,
respectively, and \( T_v \) and \( T_e \) are the virtual temperature of the parcel and environment,
respectively. CIN is the energy barrier that a parcel must overcome to become positively
buoyant, and is defined as:

\[ CIN = \int Bdz = \int_{Surface}^{LFC} g \frac{T_v - T_e}{T_w} dz, \]  

(1.2)

which is the same as CAPE, except the limits of integration are from the surface to the
LFC. On a skew-T, log-p diagram CAPE is the positive area between the environmental
temperature and the parcel path from the LFC to the EL, while CIN is the negative area
between the two profiles. Determining convective initiation is then dependent on
mechanisms that alter CAPE or CIN such that the environmental or parcel path profile is
adjusted for positive buoyancy to be realized and convection “triggered”.
Mechanisms by which CAPE or CIN are altered generally fall under two lines of thinking: 1) Quasi-equilibrium (QE) thinking (Arakawa and Schubert 1974, Emanuel et al. 1994), which focuses on alterations to the thermodynamic profiles through the entire depth of the troposphere, so could be considered more CAPE-centric and 2) Stratiform instability (Mapes 1997 and 2000), which focuses on lower-troposphere adjustments to the thermodynamic profiles, so could be considered more CIN-centric. Each view offers a unique perspective on convective organization and will be discussed in full below. The former view, QE thinking, posits convection as a slave to large-scale forcing, while the later view, stratiform instability, views convection itself as large-scale and an organizing agent for more convection. QE thinking has been the dominant influence on the formulation of cumulus parameterization schemes and has therefore had consequences on how models represent convective organization (or lack thereof).

1.2.2 Brief history of cumulus parameterization

The core of cumulus parameterization is representing the statistical behavior of unresolved cumulus convection in terms of resolved model processes. The principal closure problem is linking the existence and intensity of cumulus convection to various large-scale conditions. A complimentary problem is formulating the effects of convection on the large-scale (Arakawa 2004). The original guiding views to cumulus parameterization stem from CISK (conditional instability of a second kind) theory and QE theory, discussed below. The former is based on the principle of conditional
instability that a parcel must somehow be lifted to reach saturation and become positively buoyant (discussed above), while the later focuses on adjustments of thermodynamic profiles by deep adiabatic lifting.

1.2.2.1 CISK

Charney and Eliassen (1964) and Ooyama (1964) introduced CISK to explain the growth of tropical cyclones. They stated the growth of deep cumulus convection is cooperation between cumulus heating and large-scale moisture convergence, where the large-scale moisture convergence was provided by cyclone-scale frictionally induced boundary layer (BL) convergence (i.e. Ekman pumping). Later, wave-CISK was introduced as the synergy between cumulus heating and moisture convergence by large-scale waves (Lindzen 1974, Chang and Lim 1988). In both CISK and wave-CISK, a positive feedback loop is established wherein BL convergence increases BL $\theta_e$ lifts air parcels to their LCL, which results in cumulus development and release of latent heating that in turn amplifies the tropical vortex (for CISK) or wave (for wave-CISK). This led to parameterizations where cumulus mass flux is determined by large-scale moisture convergence (e.g. Kuo scheme, 1974 and Tiedtke scheme, 1989) (Lindzen 2003 and Bretherton 2004).

Subsequent studies, however, have questioned the causal link between large-scale moisture convergence and cumulus mass flux (e.g. Emanuel et al. 1994). While CISK is correct in that $\theta_e$ increase is key to cumulus growth, CISK fundamentally disregards convection as an instability by saying moisture convergence leads to $\theta_e$ increase. In fact,
convergence may bring lower $\theta_e$ air into the convecting region (Arakawa 2004). Additionally, tropical BL trade winds continuously lift BL air to the LCL, nullifying the need for a large-scale disturbance to do so. Rather, processes that overcome CIN are more fruitful indicators of deep convection (discussed below, 2.2.3).

### 1.2.2.2 Quasi-equilibrium

QE states that convection maintains a state of statistical equilibrium with the large-scale flow, such that large-scale destabilization of the troposphere by radiative and advective cooling and BL moistening is almost instantaneously balanced by convective heating and BL drying (Arakawa and Schubert 1974, Emanuel et al. 1994). QE presumes a time and space scale gap between convection and the larger scale. The adjustment time scale of cumulus convection must be much faster than the destabilization time scale of the large-scale and the scale over which convection is “felt” is considered local. In this regard, CAPE is relatively time invariant, such that the build up of available potential energy or CIN and processes to overcome CIN and “trigger” convection are disregarded. Instead, QE focuses on processes that alter the buoyancy of an entraining ascending air parcel through the entire depth of the troposphere. These alterations may be achieved via changes to the large-scale density (temperature) profile or the parcel’s density as a function of height.

In Arakawa and Schubert’s formulation of QE a cloud work function is defined, $A(\lambda)$, where $A(\lambda)$ is a measure of the parcels available energy – a generalization of CAPE for entraining parcels. Lambda represents an ensemble of cloud types, each with a
specified constant fractional entrainment rate. $A(\lambda)$ is separated into a large-scale component and a convective component:

$$\frac{dA(\lambda)}{dt} = \frac{dA(\lambda)}{dt}_c + \frac{dA(\lambda)}{dt}_{LS} \approx 0,$$

(1.3)

where C and LS denote the terms for the clouds (convective) and large-scale (non-convective) parts, respectively. The large-scale term in (1.3) includes surface flux, which increase a parcel’s $\theta_e$, radiative cooling, and large-scale adiabatic lifting. The cloud term includes environmental warming by latent heat release and eddy heat flux, which can be expressed in terms of mass flux for an ensemble of clouds. Under QE, large-scale adiabatic lifting, cooling, occurs through the entire depth of the troposphere. In terms of CAPE, this would be equivalent to decreasing $T_{ve}$ in (1.1) leading to an increase in CAPE (consider shifting the entire temperature profile on a skew-T left). To maintain balance, convection, and therefore the release of latent heat, would ensue to nearly cancel out the cooling. Evidence for such deep tropospheric cooling is given by Betts (1974) from soundings over Venezuela during “disturbed” (rainy) conditions. However, Mapes (1997) later argues that such soundings are the exception in the tropics rather than the rule (discussed more below).

For strict QE to hold, Emanuel et al. (1994) determined that convective heating must be in phase with the rising branch of large-scale waves and exactly in quadrature with a waves temperature perturbation. This implies there is no time lag between large-scale forcing (adiabatic ascent) and convective generation (or heating). As a result, convection associated with wave-dynamics neither amplifies nor damps the wave. In light of this, some studies view the role of waves as simply reorganizers of convection
that would occur anyway under strict QE and not producers of additional precipitation (e.g. Stevens et al. 1977 and Lindzen 2003).

If, however, QE is relaxed, such that a finite time lag is allowed between large-scale forcing and generation of convection, the convective heating shifts to the cool phase of the wave. Now, heating and temperature are negatively correlated such that convection leads to wave damping, which Emanuel et al. (1994) refer to as moist convective damping (MCD). Amplification of waves, though, may occur if enhanced surface fluxes lead convection, such that convective heating is shifted to the warm phase of the wave. This process is generally referred to as WISHE (wind-induced surface heat exchange), which describes a positive feedback between increased surface winds and increased BL \( \theta_c \), which leads to enhanced convective disturbances.

For mass continuity to hold, the mean upward motion inside convection must somehow be balanced by sinking motion outside the cloud. In QE thinking, “compensating subsidence” occurs within the clouds local environment to warm and stabilize the surrounding environment, thus preventing further convection. However, the notion of near cloud compensating subsidence such that convection is isolated between cloud free regions of sinking air is at odds with observations, which show a propensity of clouds to form in close proximity to each other.

On the global scale, convective processes must balance radiative cooling, so QE holds. However, ignoring the build up of CIN and release of CAPE seems invalid for shorter time and smaller space scales.
1.2.2.3 Stratiform Instability

As mentioned above, Mapes (1997) point out that deep adiabatic cooling is the exception to tropical soundings rather than the rule. More typical tropical soundings exhibit cooling only through the lower half of the troposphere (cf. 8 and 10 Mapes 1997). Mapes further stresses separating the tropical atmosphere by a true physical scale is flawed, as the effects of convection, namely heating, are felt at various scales beyond a convective element. Instead, convection can be considered “gregarious,” in that convection begets more convection via gravity waves (Mapes 1993).

Consider the separation of the MCS into its convective and stratiform components as discussed in the introduction. Heating from the convective component produces a fast (50 m/s) gravity wave (twice the depth of the troposphere) that provides subsidence and warming over the depth of the troposphere to the far field. This is contrary to QE thinking that subsidence occurs locally near the convective element. In fact, the slower moving (24 m/s) stratiform generated gravity provides low-level cooling and rising motion near the convective element with warming and sinking and warming to the upper half. The cooling in the near field via stratiform heating reduces CIN resulting in more favorable conditions for convection, hence the term “stratiform-instability.”

1.2.2.4 Recent advances

More recent advances in cumulus parameterization include attempts to account for the functional effects of sub-grid scale processes on convection. Xu et al. (1992) and
recently revisited by Jones and Randall (2011) showed in model simulations that cumulus activity was not completely modulated by large-scale forcing. Rather, mesoscale organization itself played a role in convective activity. This means cumulus parameterization needs to go beyond deterministic functions of large-scale variables. Adding a stochastic element to the parameterization scheme adds some higher order variability to a traditional parameterization scheme. Lin and Neelin (2003) proposed a stochastic deep-convective scheme based on random perturbations to either CAPE or the vertical heating profile.

Mapes and Neale (2011) introduce a prognostic variable, $org$, into CAM5 (Community Atmosphere Model) to account for convective organization. $Org$ acts as the controlling variable on convective plume structure and strength. As $org$ increases, plume entrainment rate of dry air decreases and precipitation strengthens. $Org$, then, acts as a positive feedback on deep convection.

Superparameterization attempts to skirt the parameterization issue entirely by placing two-dimensional (2D) CSRMs in GCM model columns (Randall et al. 2003). The net effect of the explicitly resolved convection from the CSRM is linked to the resolved large-scale processes of the parent GCM.

1.2.3 Parameterizing the “large-scale”

Recent efforts have flipped the convective, large-scale interaction problem around by explicitly resolving convection through the use of single column models (SCMs) or cloud system resolving models (CSRM) and parameterizing the large-scale. In the
context of tropical dynamics, two approaches are generally used for parameterizing the large-scale: the weak temperature gradient (WTG) approximation (Sobel and Bretherton 2000; Raymond and Zeng 2005) and the gravity wave approach (e.g. Kuang 2008).

WTG recognizes that gravity waves effectively reduce free tropospheric buoyancy gradients in the tropics. As a result, the free tropospheric temperature profile inside a SCM or CSRM column may be specified. Vertical motion, then, is that which will force the temperature inside each column to the specified reference profile; where vertical motion is parameterized as a function of the horizontal mean heating simulated by the model physics. The primitive temperature equation becomes:

$$\omega S = Q_c + Q_R + Q_{diff}, \quad (1.4; \text{Eq. (3) of Sobel and Bretherton 2000})$$

where omega is vertical velocity in pressure coordinates, S is the static stability, and the Q terms collectively represent the total heating. Under strict WTG, temperature profiles are instantaneously relaxed to the specified temperature profile. Relaxed WTG specifies a constant time scale over which the temperature profile is relaxed to the reference profile (Raymond and Zeng 2005). The moisture equation remains fully prognostic with interactive moisture convergence:

$$\frac{\partial q}{\partial t} + \mathbf{u}_h \cdot \nabla q + \omega \frac{\partial q}{\partial p} = Q_q + Q_{diff}^q, \quad (1.5, \text{Eq. (2) Sobel and Bretherton 2000})$$

where q is specific humidity, p pressure, uh horizontal velocity, and Q_q the convective moisture source and Q_{diff}^q the convergence of turbulent moisture fluxes.

Known weaknesses to the WTG approximation include: exclusion of momentum, requiring specification of omega in the boundary layer, and assuming horizontally small scales. In the boundary layer the temperature profile is not specified, rather vertical velocity is. In Sobel and Bretherton (2000) the vertical velocity is simply linearly
interpolated to zero from the lowest free troposphere level to the surface. On sufficiently large-scales temperature perturbations become important and WTG fails.

The second approach determines large-scale vertical velocity from the two-dimensional (2D) linear gravity wave equation (Kuang 2008). In this gravity wave approach, a specified horizontal wavelength is chosen to be much larger than the SCM or CSRM domain. Briefly, the wave approach maintains a balance between domain averaged temperature perturbations from explicitly resolved convection inside a CSRM and the large-scale vertical velocity derived from the 2D linear gravity wave equation. This approach is described in detail in section 4.2.1 below.

Predecessors to the gravity wave approach were introduced by Mapes (2004) and Bergman and Sardeshmukh (2004). In both papers, large-scale vertical velocity was parameterized as a time lagged function of SCM or CSRM domain averaged convective heating; similar to Kuang’s later approach. However, in Mapes (2000) only a single vertical wave structure and gravity wave speed was assumed. While Bergman and Sardeshmukh (2000) allowed for a spectrum of vertical wave structures and speeds, the approach developed by Kuang is more straightforward and elegant.

1.2.4 Factors that help organize deep tropical moist convection

The introduction of Tompkins (2001a) nicely lists factors that help organize tropical convection. They are, in no particular order: sea surface temperature (SST) gradients, localized convective moistening, vertical wind shear, cold pools, gravity waves (e.g. stratiform instability), and dry air intrusion layers (e.g. Mapes and Zuidema 1996).
Other factors include: large-scale waves (e.g. CCWs, Kiladis et al. 2009), cloud-radiation interactions (e.g. QE), and wind-surface flux interactions (i.e. WISHE).

1.2.4.1 SST

The general spatial distribution of deep tropical moist convection is highly correlated with SSTs (e.g. deep convection’s prevalence over the W. Pacific warm pool). An observational study by Zhang (1993) found that deep convection was relatively weak and rare over SST < 26°C, but became increasingly more frequent and intense for SSTs between 26°C – 30°C. Bony et al. (1997), later emphasized that SST gradients, rather than the absolute SST value, are key to spatially distributing convection, since the gradients induce a large-scale circulation, which preferentially favors upward motion over high SSTs.

1.2.4.2 Localized convective moistening

Zhang (1993) noted a surprising reduction in convection over SSTs > 30°C. Tompkins (2001b) explained these “hot-spots” as a self-aggregation feedback between convection and water vapor. Under the subsiding branch of a large-scale circulation areas of relatively high SST can occur due to increased incident solar radiation. The dry free troposphere associated with the descending branch of the circulation prevents deep convection from breaking out. Rather, convection slowly propagates toward the higher SST values at a rate which it can moisten its surrounding environment. This moisture-
convective feedback is expounded on in Grabowski and Moncreiff (2004). Basically, convection preferentially forms in moist environments, where environmental moisture can increase from detrainment and evaporation from prior convection. The importance of this feedback was shown in Kuang (2008) when convectively coupled waves ceased after moisture advection by the large-scale wave was turned off in his CSRM simulations.

1.2.4.3 Shear

Vertical wind shear has long been recognized to organize deep convection into arcs or lines both in the tropics and extra-tropics (e.g. Houze and Betts 1981, Moncrieff 2006). LeMone et al. (1998) went so far to say that shear “has the dominant effect on organization of deep convection on the mesoscale.” Low-level shear tends to promote organization along arcs oriented nearly normal to the shear vector, which propagate in the direction of the shear, while strong shear at middle and upper levels with weak low-level shear tends to favor shear-parallel lines (e.g. LeMone et al. 1998, Robe and Emanuel 2001). Beyond organizing convection, Schumacher and Houze (2006) found that shear can be beneficial or a deterrent to MCS sustainability. For MCS over Africa during the active monsoon season they found weak upper-level shear important for maintaining leading-line, trailing-stratiform systems. However, when the upper-level shear was too strong, hydrometeors were spread beyond the stratiform rain area and the MCS decayed more rapidly than under weak upper shear conditions.
1.2.4.4 Cold Pools

Cold pools form when convective downdrafts reach the surface and spread horizontally. The spreading air acts as a density current or gust front, mechanically lifting air near the parent convection to its LCL spawning new cells (Bluestein 1993). More recently, Tompkins (2001) highlighted cold pool’s thermodynamic role in convective organization. As the cold pool spreads, dynamical uplift diminishes while air just inside the gust front maintains a higher $\theta_e$ than its surrounding environment, thus promoting a reduction of CIN and increasing the likelihood of convection reaching their LCL.

1.3 Dissertation layout

The next chapter briefly discusses the data sets used for the observational component of this work and the CSRM used for the modeling component. Following Chapter 2, Chapter 3 and 4 discuss the findings from CloudSat observations and the CSRM modeling work, respectively. Chapter 5 summarizes and concludes the dissertation, while Chapter 6 outlines possible future work related to this study.
CHAPTER 2: Data Sets and Model

2.1 CloudSat

The CloudSat satellite flies as one of the polar orbiting satellites a part of the A-Train constellation. CloudSat was launched April 2006 and remains active today (with a blip in battery power from 17 April 2011 to 15 May 2012). Onboard CloudSat is a nadir pointing 94 GHz (W-band) Cloud Profiling Radar (CPR) (Im et al. 2005). The CPR has a 1.4 km across track by 1.8 km along track nominal footprint, but oversampling results in profiles 1.1 km apart. Vertical resolution is 480 m, oversampled to give 240 m bins. CloudSat’s minimal detectable signal is -30 to -31 dBZ (Haynes and Stephens 2007).

CloudSat makes roughly 14 orbits per day with an equator passing time of 1:30 and 13:30 local time. Approximately, 6.5 years (16 June 2006 to 17 January 2013) of CloudSat data has been processed. However, only four years (16 June 2006 to 15 May 2010 of CloudSat data are used in Chapter 3. CloudSat hydrometeor echo objects were

In Riley and Mapes (2009; hereafter RM09) hydrometeor echoes were analyzed on two tiers: (1) as echo objects (EOs) and (2) as pixels comprising EOs. This dissertation uses the same approach. Briefly, EOs are identified using version 5 of the 2B-GEOPROF CloudSat product. The 2B-GEOPROF product contains 2D arrays of the radar reflectivity factor measured by the 94 GHz cloud profiling radar (CPR), as well as a “cloud mask” value ranging between 0 – 40. Higher cloud mask values indicate an increased likelihood of hydrometeor detection (Mace 2004). EOs are defined as a contiguous region of cloud mask ≥ 20, consisting of at least three pixels with their edges
(not merely corners) touching. For each EO top and base height, width, and geographical information, as well as several other attributes are saved (Table 2.1 in Riley 2009).

Following RM09, vertical temperature and moisture profiles were also available for each EO, constructed by averaging temperature and moisture at each altitude in the bounding box region of each EO from CloudSat’s European Centre for Medium-range Weather Forecasting auxiliary product (ECMWF-AUX), which has been interpolated to the CloudSat grid on a pixel-by-pixel basis.

There are 15,181,193 EOs in the full 6.5 years of data and 10,979,780 EOs in the four years of data used in Chapter 3. A joint histogram of tropical (latitude < 15°) EO base height versus EO top height reveals natural separations in cloud types (RM09’s Fig. 1a, reproduced here as Fig. 2.1a). Low-based EOs lie along the left vertical axis, while layer type clouds occur along the diagonal. By parsing the histogram into the indicated boxes, seven familiar cloud types are distinguished: (1) deep precipitating, (2) detached anvil (thick cirrus), (3) cirrus, (4) cumulus congestus, (5) alto clouds (including altostratus and altocumulus that are referred to as altocumulus throughout the text), and low clouds: (6) stratocumulus (width > 10 pixels) and (7) cumulus (width ≤ 10 pixels). Each EO is assigned one of these cloud types. Random examples of each EO type are given in Fig. 2.1b. These EOs are unlike the CloudSat cloud classification product (Wang and Sassen 2007) in that an entire continuous object is assigned a single type. Also, the cirrus definition does not include very thin cirrus, as those clouds contain relatively small particles that go undetected by CloudSat’s millimeter wavelength (Marchand et al. 2008). CloudSat also misses some shallow clouds due to surface contamination (Sassen and Wang 2008).
Figure 2.1 – (a) Distribution of horizontal cloud cover by EOs in the tropics (15° S–15° N) accounted for by clouds with tops and bases in the indicated bins for 16 Jun 2006–15 May 2010. Contour values are in units of 10^3 horizontal echo pixels per bin. Bin size is 240 m × 240 m. Lines and letters delineate EO types, roughly associated with cloud types: A, deep precipitation; B, detached anvil (thick cirrus); C, cirrus; D, cumulus congestus; E, altostratus and altocumulus; and F, low clouds, both stratocumulus (pixel width > 10) and cumulus (pixel width ≤ 10) (adapted from RM09) (b) Random examples of EO types. A subjective number of EOs were chosen to plot for each type, so there is no relation between the different elements of the figure. Type is indicated by transparent color background.
2.2 Year of Tropical Convection (YOTC) CloudSat collocated A-Train and ECMWF Data

YOTC was a joint WCRP (World Climate Research Programme) and WWRP/THORPEX (World Weather Research Programme) virtual field campaign from May 2008 to April 2010. The aim of YOTC is to synergize observations, modeling, and forecasting to intensely study tropical convection (Waliser et al. 2012, Moncrieff et al. 2012). To that end, a CloudSat-centric data set was developed that co-locates all A-Train satellite (i.e. Aqua, CloudSat, CALIPSO, and Aura) data and two ECMWF analysis outputs to the same grid.

CloudSat data geo-location and time were used to establish the common location and time for all data products. A simple nearest in time and space criteria was used to match individual products to the CloudSat grid (details found at http://csyotc.cira.colostate.edu/data_documentation.php). Fig. 2.2 is a schematic description of A-Train and ECMWF co-located data products and the satellite associated with each.

To date, AMSR, ECMWF pressure level data, MLS water vapor mixing ratio, and CERES variables available from the YOTC campaign have been processed to EO level statistics. A detailed list of all variables are in the collection parameters excel file at the website referenced above. Each EO has been assigned the mean variables from the aforementioned satellites and/or the complete vertical profile of variable information has
been assigned to each EO. Examples of preliminary work done using the YOTC data are given in Chapter 6.

Figure 2.2 – Schematic of A-Train variables available along the CloudSat track for the two YOTC years. (Fig. 7 of Moncrieff et al. 2012)

2.3 System for Atmospheric Modeling (SAM)

SAM 6.8.2 is a cloud system resolving model (CSRM) developed by Marat Khairoutdinov. Details of the model are found in Khairoutdinov and Randall (2003). Briefly, the model uses the anelastic equations of motion with three prognostic thermodynamic variables: liquid water static energy, total non-precipitating water, and total precipitating water. The model uses bulk microphysics with five types of
hydrometeors: cloud water, cloud ice, rain, snow, and graupel. A Smagorinsky-type closure is used for sub-grid scale turbulence. Periodic boundary conditions are used with a rigid lid at the domain top.

For this dissertation, 2 km horizontal grid spacing was used with 64 stretched vertical levels (~100 m near the surface to 500 m in the mid and upper-troposphere to 1 km near the top). The ocean surface was fixed at 29.5°C. A background radiation profile and vertical velocity profile were prescribed (Fig. 2.3). The velocity profile is the averaged vertical velocity over the large-scale array (LSA) during the intensive operating period (IOP) of the Tropical Ocean Global Atmosphere Coupled Ocean-Atmosphere Response Experiment (TOGA-COARE; Webster and Lukas 1992) from 1 November 1992 – 28 February 1993. A 15-second time step was used for all runs and output was saved every 3-hours, unless otherwise noted. Surface fluxes were computed with $|V|$ in equations A.1 and A.2 fixed at 5 ms$^{-1}$ to prevent WISHE. Domain size and dimensionality will vary depending on the experiment (described in Chapter 5.2).

Figure 2.3 – (a) Background vertical velocity and (b) radiation profile prescribed for all SAM CSRM runs. The velocity profile is the averaged vertical velocity over the LSA during TOGA-CORE IOP (see text for more details)
CHAPTER 3: Clouds Associated with the Madden-Julian Oscillation: A new perspective from CloudSat

3.1 Background

The Madden Julian Oscillation (MJO) is the dominant coherent mode of intraseasonal variability in the tropics (Madden and Julian 1972, Zhang 2005). At its most basic level, the MJO can be described as a multiscale envelope of organized convection and cloudiness that moves slowly eastward over the Indo-Pacific at approximately 5 m/s. Motivation for studying the MJO comes from its profound influence on tropical weather over the Indo-Pacific, monsoon rains over the nearby continents of Asia and Australia, and tropical cyclone throughout the globe, as well as, mid-latitude teleconnections and influences on the evolution of the El Nino – Southern Oscillation (ENSO) (see reviews by Madden and Julian 1994 and Zhang 2005). Despite the MJO’s importance to society, complete understanding of the phenomenon remains elusive as illustrated by the failure of most global circulation models (GCMs) to accurately simulate the MJO (Zhang 2005, and references therein). Furthermore, a comprehensive theory explaining the MJO’s initiation, propagation, and time period is lacking (see Wang 2005 for a theoretical review).

Data from various sources have been used to describe the large-scale dynamical structure of the MJO. These data include model reanalysis products (e.g. Lin et al. 2004, Kiladis et al. 2005, Benedict and Randall 2007), radiosondes (e.g. Lin and Johnson 1996, Lin et al. 2004, Kiladis et al. 2005), and satellite sounders and precipitation retrievals
(e.g. Myers and Waliser 2003, Tian et al. 2006, Benedict and Randall 2007, Morita et al. 2006). Results of these studies show that during the suppressed (dry) phases easterly winds predominate, with descending motion and anomalous dry conditions throughout the troposphere. As convection and ascending motion start to build, surface winds become westerly and low-level moistening occurs. During peak convective activity, low-level temperature anomalies are generally cool with warm anomalies through the majority of the upper troposphere and anomalous moisture in the middle troposphere. Following the wet phase, low-level drying ensues and builds upward and descending motion and easterlies return (see Fig. 3 of Benedict and Randall 2007, and Fig. 7 of Zhang 2005).

With the above knowledge, schematics have been published to visually express convective activity and clouds through the MJO (Fig. 3.1). All three MJO schematics (Fig. 3.1a,b,c) generally show a progression of cloud types from shallow, to middle topped, to deep clouds, followed by high topped somewhat thick clouds, with an eventual return to shallow clouds. Interestingly, this progression of cloud types is similar to cloud evolution observed over the life cycle of mesoscale convective systems (MCSs) (Fig. 3.1d).

While schematics are helpful at summarizing results, they often lack objectivity. With the advent of CloudSat in spring 2006, direct observations of vertical cloud structure became possible (Stephens et al. 2002). The aim of this chapter is to provide an objective view of clouds across the MJO, using CloudSat’s high sensitivity to cloud particles. Documenting cloud structure and its change across varying phases of the MJO will hopefully lead to a better understanding of convective organization within the MJO.
Furthermore, we hope to understand the relative importance of different cloud types to different MJO phases.

The novelty of CloudSat is its ability to see mid-level and low-level clouds that are optically obscured by high clouds. Myers and Waliser (2003) and Troumeur and Rossow (2010) used the ISCCP (International Satellite Cloud Climatology Project) data set to examine cloud structure across the MJO, but neither study was able to capture the vertical structure of clouds or the variation of cloud types in fine detail. Lau and Wu (2010) also did an analysis of cloud type variations across the MJO, but using TRMM (Tropical Rainfall Measuring Mission) data products, so focused on precipitating cloud types. CloudSat helps fill the gaps of these studies. For example, Masunaga et al. (2008) compared the vertical structure of reflectivity measured by CloudSat to simulated reflectivity during an MJO event from December 2006 to January 2007. Jiang et al. (2010) used CloudSat to evaluate total cloud fraction and cloud fraction by cloud type.
during boreal summer intraseasonal variability events. Also, Zuidema and Mapes (2009) compared cloud vertical structure results between the CloudSat CPR and field experiments in the Bay of Bengal (i.e. JASMINE) and the eastern Pacific intertropical convergence zone (i.e. EPIC) that used shipped based cloud radars.

The next section discusses the methods used. Sections 3.3 and 3.4 present results of cloud variations across different phases of the MJO. Section 3.5 provides a discussion of the results, while section 6 summarize the study.

3.2 Methodology: Defining MJO indices

Two approaches are used to define MJO events and phases. The first defines MJO phases relative to minima, maxima, and rate-of-change (i.e. slope) conditions in time series of space-time filtered outgoing longwave radiation (OLR) time series. Filtered OLR is used because it is a good proxy for tropical deep convection and familiar in MJO studies (e.g. Nakazawa 1988 and 1995, Lau and Chan 1985, Salby and Hendon 1994, Matthews 2008). The OLR data are twice daily on a 2.5° by 2.5° grid from 16 June 2006 – 15 May 2010. The MJO signal was obtained by space-time filtering raw OLR to isolate zonal wavenumbers 0 – 9 and periods between 30 – 96 days. We averaged the filtered OLR from 15°S – 15°N to obtain a longitude-time array (Fig. 3.3). This filtered OLR and its local time derivative were both standardized, by dividing by their respective standard deviations within the entire array. For reference, the standard deviation of this filtered OLR is 5.82 W m$^{-2}$. 
From these standardized quantities, eight phases were defined based on regions in the scatter plot of Fig. 3.2a. A pure sine wave would be a unit circle on such a diagram, but because many frequencies are present the points fall along spiral forms. To isolate strong MJO events, we only consider locations and times when the amplitude (distance from the origin in Fig. 3.2a) is greater than two (colored numerals in Fig. 3.2a). Phase 1 (black numeral 1 symbols) represents the most suppressed phase of the MJO, with the highest filtered OLR values and a time derivative near zero. Opposite phase 1, phase 5 (green) represents the most active phase of the MJO because it has the lowest filtered OLR values and a time derivative near zero. Phases 2 – 4 are the building phases and phases 6 – 8 the decaying phases of the MJO’s convective activity. Perhaps a clearer rendering of the phases is in time-longitude space (Fig. 3.3). Each color band represents the same phase as the colored numbers in Fig. 3.2a. Phase 1 (black stripes) runs through maxima in the filtered OLR, phase 5 runs through minima, and phases 2 – 4 and 6 – 8 fill in the periods between the maxima and minima. We will refer to the phases defined this way as “pinwheel” phases, as the spiral form taken in Fig. 3.2a resembles a pinwheel.

Figure 3.2 – a) Pinwheel phases. (b) RMM phases. Each colored diamond represents a day in our CloudSat data set that qualified as part of an MJO event. There is no relation between the colors in the two figures. (Fig. 3 of Riley et al. 2011)
Each EO in the latitudinal belt 15°S – 15°N was assigned to an MJO phase, by looking up the phase value at its time and mean longitude in the longitude-time array of pinwheel phase. For EOs falling in longitude-time regions with standardized amplitude <2, a phase of -1 was assigned, and such EOs are excluded from all further analysis.

The second approach defines eight MJO phases using Wheeler and Hendon’s (2004, hereafter WH04) Real-time Multivariate MJO (RMM) index (http://cawcr.gov.au/staff/mwheeler/maproom/RMM/). These phases are a function of time only, based on time series of the whole tropical belt’s resemblance to a MJO “mode” structure derived from empirical orthogonal function (EOF) analysis. The RMM1 and RMM2 indices measure the projection of daily 15°S – 15°N averaged longitude sections of OLR, 850-hPa zonal winds, and 200-hPa zonal winds onto the leading pair of EOFs of these same combined fields (with their annual and interannual variability removed) found from historical data. WH04 found that this spatial screening for large-scale structures in this combination of the three variables acts as an effective filter for intraseasonal frequencies associated with the MJO, without any time filtering. The first EOF describes enhanced convection over the Maritime Continent, and low-level easterlies across the Pacific, with upper-level westerlies. The second EOF has enhanced convection over the Pacific Ocean and wind patterns, approximately in quadrature to those of the first EOF (Fig. 1 of WH04).

The standardized principal components (PCs) of the leading pair of EOFs are referred to as RMM1 and RMM2. Because RMM1 leads RMM2 by 10-15 days, the MJO appears as a point moving counterclockwise in their joint phase space (Fig. 3.2b). Labels in each wedge in the diagram roughly indicate the geographical location of
enhanced convection associated with the MJO. Each diamond in Fig. 3.2b corresponds to a day within our CloudSat data set. Colored diamonds are days on which \((RMM1^2 + RMM2^2) > 1\), qualifying them as part of an MJO event. Each EO is assigned a WH04 RMM phase, based only on the day in which it occurred. Again, a phase flag of -1 was used to label EOs on days on which the RMM-defined MJO was weak, and such EOs were excluded from further calculations.

Figure 3.3 – Filtered MJO OLR in time-longitude space averaged over 15°S – 15°N for 15 June 2006 to 15 May 2010. Contour interval is 10 W/m² starting at 5 W/m² (solid) and -5 W/m² (dashed). Colored shading indicates MJO phase for MJO events. Horizontal bars indicate data gaps in CloudSat. (Fig. 4 of Riley et al. 2011)
3.3 Results as a function of pinwheel phase

3.3.1 EO statistics

Fig. 3.4a shows horizontal pixel coverage for each EO type by MJO pinwheel phase. Horizontal pixel cover is the number of cloudy (i.e. cloud mask ≥ 20) CloudSat profiles per EO, or simply the pixel-width per EO. Panel b shows the same information, but normalized by total horizontal pixel cover in each phase bin. Deep precipitating (Dp) EOs have been subdivided into narrow (width < 200 km, red) and wide (width > 200 km, orange), to test the conclusion of Morita et al. (2006) and Tromeur and Rossow (2010) that deep narrow (or isolated) convection is more prevalent during suppressed MJO conditions, while wide convection is more prevalent during active MJO conditions (Fig. 3.1a).

The most noticeable feature of Fig. 3.4a is the 1.5 fold increase of total EO coverage from suppressed to active phases. Though the wide Dp contributes most to the EO coverage increase, all types of EOs are more prevalent in the active phase. Fig. 3.4c,d explicitly shows this increase. Plotted in Fig. 3.4c is the absolute minimum to maximum change in horizontal pixel cover for each EO type. For example, the cirrus EO type has maximum cover in phase 6 (321,069 horizontal pixels) and minimum cover in phase 1 (208,977 horizontal pixels), giving a difference of 112,092 horizontal pixels (i.e. Fig. 3.4c green bar). Since the wide deep precipitating type dominates the absolute change in horizontal pixel cover, Fig. 3.4d shows the fractional increase in horizontal pixel cover from minimum to maximum phase for each EO type. The wide deep precipitating, anvil, and altocumulus types increase about twofold or more, while narrow
deep precipitating and anvil types increase 1.5 fold, and cumulus congestus, stratocumulus, and cumulus increase less than 1.5 fold from their minimum to maximum horizontal pixel cover. These fractional changes indicate the degree to which the MJO modulates each EO type. In that regard, wide deep precipitating, anvil, and altocumulus types are modulated most, while cumulus congestus, stratocumulus, and cumulus are modulated least (i.e. fractional change < 1.5).

Figure 3.4 – a) Total horizontal pixel cover and (b) normalized horizontal pixel cover percentage for each EO type per MJO pinwheel phase. (c) Absolute difference between the phase with maximum (max) horizontal pixel cover and the phase with minimum (min) horizontal pixel cover for each EO type. The phase with the max horizontal pixels and min horizontal pixels per EO type is as follows: narrow deep precipitating (ndp) max phase is 4 and min phase is 7; wide deep precipitating (wdp) max phase is 5 and min phase is 1; anvil (an) max phase is 6 and min phase is 3; cirrus (ci) max phase is 6 and min phase is 1; cumulus congestus (cg) max phase is 5 and min phase is 8; altocumulus (ac) max phase is 5 and min phase is 2; stratocumulus (sc) max phase is 6 and min phase is 4; and cumulus (cu) max phase is 6 and min phase is 4. (d) The fractional increase from the minimum horizontal pixels to the maximum horizontal pixels for each EO type. (Fig. 5 of Riley et al. 2011)

All EO types have maximum horizontal pixel cover in either phase 4, 5, or 6 (see Fig. 3.4 caption). The wide deep precipitating, cumulus congestus, and altocumulus EO
types have their maximum horizontal pixel cover in phase 5 (coincident with the minimum in OLR), while the anvil, cirrus, stratocumulus, and cumulus EO type maxima trail the OLR minimum by one phase (phase 6) and the narrow deep precipitating type maximum leads the OLR minimum by one phase (phase 4). Each EO type’s minimum horizontal pixel cover occurs either before or after the active phases (phases 4-6), except for the shallow types (i.e. stratocumulus and cumulus). That anvil and cirrus maxima trail the wide deep precipitating maximum by one MJO phase is consistent with the well established picture of convective cloud type evolution across various time and space scales: shallow to deep to high clouds (e.g. Zipser 1981, Lin and Johnson 1996, Mapes et al. 2006, Morita et al. 2006, Benedict and Randall 2007, and Kiladis et al. 2009).

The above results are mostly consistent with Lau and Wu (2010) who showed increased occurrence in all their precipitating and non-precipitating cloud types during active phases, except for warm-low clouds (their figs. 6 and 7). Since their cloud types are defined by brightness temperature from TRMM’s visible and infrared scanner (VIRS) and TRMM’s precipitation radar (PR), their low clouds during active phases could have been obscured to the VIRS by the abundance of higher level clouds during active phases and/or gone undetected by the TRMM PR threshold (dBZ ≥ 17, Kummerow et al. 1998) since the low clouds may not be raining or just drizzling.

In normalized terms (Fig. 3.4b), narrow Dp EOs gradually increase their percent contribution from phases 1 to 4, then slowly decrease through phase 7, suggesting a somewhat different cloud behavior than in Morita eta al. (2006) and Troumeur and Rossow (2010). Here the narrow deep convection is modulated according to building vs. trailing phases compared to active vs. suppressed phases. Anvil EOs have their greatest
(smallest) fractional contribution after (during) the active phases. Low clouds (Sc and Cu) have their greatest contribution during the suppressed phases (1, 2, 6, 7 and 8).

Because our EO analysis retains links to pixels, we can generate pictorial realizations of “actual” (i.e. CloudSat observed) clouds (EOs) across the MJO (e.g. Fig. 3.5). The aim of our pictorial realizations is to mimic schematics from previous works (e.g. Fig. 3.1 references), but with objective methods and CloudSat observed clouds. To make a pictorial realization, random samples are drawn from a random number generator from the set of MJO EOs, and rendered as a set of pixels centered over the EO’s continuous pinwheel phase value (see Fig. 3.5 caption for how continuous phase relates to discrete phase). The conversion factor between EO width (in pixels) and width on this diagram (in phase units) is an arbitrary choice, adjusted manually for graphical clarity. Also for clarity, pictorial realizations sample only from EOs with amplitude >3 (as opposed to a threshold of >2 for all other statistics). The random selection and plotting of EOs is repeated until a desired amount of total horizontal echo coverage is reached. One final nod to clarity: wide Dp EOs are rendered in a gray-to-red reflectivity color scheme, while the pixels making up all other types are in a gray-to-blue scheme. The faintest discernable shade of red or blue occurs near 0 dBZ.

Fig. 3.5 is one example of a pictorial realization, other examples can be found online (website reference). The example here highlights some of the statistics in the bar graphs (Fig. 3.5): There is more cloudiness during the active phases (phases 4-6), with the wide Dp EOs dominating these phases and completely absent in the most suppressed continuous phases (phases 0-2 and 7-8). Narrow Dp EOs are seen during both suppressed and active conditions (i.e. the tall blue towers near continuous phase 0.3, 1.7, 3.1, 4.8,
7.3, and 8). Shallow EOs are prevalent during all phases. The statistical variation of the other EO types is difficult to discern in this pictorial realization. However, in terms of the main features, Fig. 3.5 is quite comparable to Morita et al.’s (2006) hand-drawn Fig. 3.14 (reproduced here as Fig. 3.1c), but has the virtue of objectivity.

Figure 3.5 – One example of a pictorial realization of CloudSat clouds across continuous pinwheel phases of the MJO. Each EO is plotted centered over its continuous phase, which is multiplied by 500 (the bottom x-axis) to maximize viewing clarity. The top x-axis corresponds to the actual EO’s continuous phase. Continuous phase relates to discrete phase as follows: EOs with $0.5 \leq \text{continuous phase} < 1.5$ are in discrete phase 1, $1.5 \leq \text{continuous phase} < 2.5$ are in discrete phase 2, etc, and continuous phase $< 0.5$ and continuous phase $\geq 7.5$ are in discrete phase 8. Wide dp EOs are grey-to-red, while all other EO types are grey-to-blue. Colored shading starts at dBZ > 0. The seed is the number inputted to the random number generator to recreate this example. (Fig. 6 of Riley et al. 2011)

### 3.3.2 Pixel-level statistics

The echoes in each pinwheel phase of the MJO can also be characterized with 2D histograms of pixel dBZ vs. height (also known as contoured frequency by altitude diagrams, CFADS; Yuter and Houze 1995). It is clearest to establish the mean or background cloudiness of MJO-affected regions, then examine anomalies relative to that. Fig. 3.6’s first panel shows the normalized CFAD (NCFAD) for all EOs in all pinwheel
phases of the MJO, wherever pinwheel amplitude > 2. The most pixels occur around -25 dBZ between 11 km and 14 km. Pixels are also common near the rainfall attenuation line, with high reflectivities below 5km (white line in panel a). A broad minimum in the NCFAD occurs at reflectivities below 0 dBZ at mid-levels (between 3 km and 8 km).

Fig. 3.6’s other panels show each pinwheel phase’s NCFAD anomalies, overlaid by the background contours from Fig. 3.7a for reference.

Figure 3.6 – a) Normalized CFAD (Contoured Frequency by Altitude Diagram, or 2D histogram of pixel dBZ vs. pixel height) for all eight pinwheel phases. White line indicates the approximate rainfall attenuation line. (b – i) Shading is the percent difference between the normalized CFAD of the indicated phase and the normalized CFAD for all eight pinwheel phases (panel a). The black contours in each panel are the same as the black contours in panel (a). (Fig. 7 of Riley et al. 2011)

During suppressed phases 1 and 2, the most enhanced echoes (relative to all-phase mean echo) have low-reflectivity values, both at low levels (z < 5 km) (the signature of
nonprecipitating cumulus clouds), and mid-upper levels (9 km < z < 13 km) (the signature of cirrus, as detailed in Fig. 3.21 of Riley 2009). Additionally, in phase 2 high reflectivity upper levels enhancements emerge. By phase 3, high reflectivities above the freezing level, suggesting intense updrafts, are well established, along with enhanced echo frequencies along the high-reflectivity and high-altitude margins of the background NCFAD. Low-level, high reflectivity enhancements also appear in phase 3, indicating rainfall. In phases 4 and 5 (Figs. 3.6e,f), high reflectivity enhancements increase, indicating very deep convection and rainfall. Mid-level enhancements also appear in phase 4 and 5. The results during the active phases 4 and 5 are consistent with Fig. 4 of Masunaga et al.’s (2008) CloudSat CFAD differences from the wet and dry phases of the December 2006 – January 2007 MJO event (where they defined wet and dry by rainfall measured from the TRMM PR).

In phase 6, rain-related enhancements are nearly gone. Mid-level enhancements remain over the breadth of the reflectivities, while upper-level enhancements below -10 dBZ shift down ~ 1 km. In phase 7 (Fig. 3.5h), rain-related enhancements disappear completely. What remains a wedge of enhanced frequency from 5 km to 12 km with moderate reflectivities, along the base of the upper-level mode of the climatological distribution. These enhancements are a signature of anvil tropical NCFAD enhancements, as in fig. 3.21 of Riley (2009). Finally, enhancements in phase 8 (Fig. 3.6i) are similar to those in phase 1, indicating a return to suppressed conditions.

Looking more closely at reflectivity enhancements, Fig. 3.7 shows the mean NCFAD and anomalous CFADs during phases 3 and 7 for just wide deep precipitating EOs. The most common pixels in the all-phase mean NCFAD are at low levels (z < 5
km) and high reflectivities (dBZ > 0) (Fig. 3.7a). During phase 3, the wide deep precipitating EOs have reflectivity enhancements at low levels and high reflectivities, as well as along the upper margins of the background NCFAD (Fig. 3.7b). Phase 7 contrasts almost perfectly with phase 3: positive enhancements are in a wedge from 5 to 14 km and 225 to 5 dBZ with negative anomalies elsewhere (Fig. 3.7c). This contrast suggests that wide deep precipitating EOs prior to the most active phase (i.e., phase 5) are more bottom heavy and may contain attached precipitating midlevel EOs, while after phase 5 the wide deep precipitating EOs contain more attached anvil.

Anomalous phase CFADs for narrow deep precipitating EOs are noisy and not shown. However, the all-phase mean NCFAD is useful to compare with the wide deep precipitating all-phase mean NCFAD. Summing each NCFAD over its reflectivity dimension yields the vertical normalized percent profile of echo cover (Fig. 3.7d). The total of each profile is 100%. The narrow deep precipitating profile (dashed line) is more bottom heavy than the wide deep precipitating profile (solid line), suggesting that the greater fractional contribution by narrow deep precipitating EOs during the building phases (phases 1–3) (Fig. 3.4b) is weighted more by bottom-heavy deep cloud systems.

Summing a CFAD over its reflectivity dimension yields a profile of echo coverage. Repeating this for each MJO phase, and dividing by the total number of CloudSat-sampled profiles during that phase of the MJO, yields Fig. 3. 8a. Echo cover at all heights roughly doubles from suppressed (phases 1 – 3) to active (phases 4 – 6) conditions, while the height of the highest echoes rises about 1 km. These changes are consistent with Figs. 3.3 and 3.4 indicating the prevalence of wide intense deep convection in active phases. Each phase has two prominent echo cover peaks, one
centered around 1.5 km and the other near 12 km. During phases 1 and 2, echo cover for the lower and upper peaks is comparable. Subsequent phases show the upper peak growing relatively more than the lower peak. Fig. 3.6 also shows that upper-level cloudiness is slightly greater after active phases than before, again illustrating the prevalence of anvil type EOs there, consistent with Fig. 3.6i and Fig. 3.4b.

Figure 3.7 – (a)–(c) As in Figs. 3.6a, 3.6d, and 3.6h, respectively, except for wide deep precipitating (wd) EOs only. (d) Vertical normalized percentage profile of echo cover for wide deep precipitating (wdp; solid line) and narrow deep precipitating (ndp; dashed line) EOs. The total of each profile is 100%. The profiles are made by summing over the reflectivity dimension of the ndp and wdp NCFAD. (Fig. 8 of Riley et al. 2011)
3.3.3 ECMWF Temperature and Moisture

Fig. 3.8b shows temperature anomalies during each pinwheel phase of the MJO. This figure was constructed by simple averaging of the T(z) profiles associated with all the EOs in each phase bin, then subtracting from each phase bin the average profile over all 8 phase bins. In other words, the horizontal average of Figs. 3.8b,c at each altitude is zero by construction. The coherence of anomalies in adjacent phase bins indicates their robustness, and the main features here are consistent with previous studies (discussed below), so we have not made a careful estimation of statistical significance, and will not attempt to attach interpretation to subtle features.
In the active phases (4-5-6), the lowest kilometer of the troposphere is cool. This simultaneous relationship apparently reflects the effects of rain evaporation and convective downdrafts (e.g. Lin and Johnson 1996, Kemball-Cook and Weare 2001), since sea surface temperature tends to lag convective activity (Sperber 2003). The 1-5km layer has much weaker variations, lagging behind the planetary boundary layer (PBL) by perhaps one phase category. The 5-14 km layer varies coherently, and clearly leads convection by a whole phase bin, with opposite signed fluctuations above that up to ~18km, tilting forward (toward earlier phase) with height, consistent with upward propagating gravity waves excited by a moving heat source (Kiladis et al. 2001, 2005, Tian et al. 2006). Schwartz et al. (2008) found a similar eastward tilt in water vapor and temperature anomalies from MLS (Aura’s Microwave Limb Sounder) and GOES (Goddard Earth Observing System Data Assimilation System) satellites, respectively. Virts and Wallace (2010) also detected an eastward tilt in CALIPSO (Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations) detected tropical tropopause transition layer (TTL) cirrus. The range of variations is about 0.5 K for this unweighted composite of all fluctuations with pinwheel amplitude >2 standard deviations (i.e. twice the 5.82 W m⁻² standard deviation in filtered OLR).

The temperature structure of Fig. 3.8b is consistent with previous studies (e.g. Lin and Johnson 1996, Kiladis et al. 2005, Tian et al. 2006, and Benedict and Randall 2007). In each of those studies, a cool-warm-cool vertical tripole structure was found during the active periods of the MJO. Before the active period both Kiladis et al. (2005) and Benedict and Randall (2007) showed anomalous low-level tropospheric warmth. After the active period these studies generally showed cooling through most of the troposphere.
Fig. 3.8c shows specific humidity anomalies during each pinwheel phase of the MJO, constructed in the same manner as the temperature profiles. Generally, the troposphere is anomalously dry through all depths during the suppressed phases (i.e. phases 1 and 8) and moist during the active phases (i.e. phases 4 and 5), but there is systematic rearward tilt (toward later phases with height) to the anomalies. In phase 2, anomalously moist conditions appear in the lowest kilometer of the troposphere, while conditions remain dry aloft. The moisture anomaly deepens to 6 km at phase 3 and phase 4 and 5 are anomalously moist through the entire depth of the troposphere. Phases 6-7-8-1 are almost precisely the opposite of 2-3-4-5. Again the anomalies shown here are consistent with previous works (e.g. Kemball-Cook and Weare 2001, Sperber 2003, Kiladis et al. 2005, Tian et al. 2006, and Benedict and Randall 2007). A comparison with Kelvin wave anomalies (Fig. 3.12) is done in section 3.5 to help put these MJO anomalies in the context of other wave types.

3.4 Results: WH04 RMM Phases

A limitation of the above analysis is that it does not convey how changes in the geographic distribution of cloudiness are associated with the MJO. This section therefore describes results using the more geographically discriminating RMM index of WH04. Fig. 3.9 shows cloud cover in eighteen 20° longitude bins across the tropics during each of the 8 RMM phases when amplitude is greater than one. The line plot in each panel of Fig. 3.9 shows the mean total horizontal pixel cover over all 8 phases for each longitude bin. Here the background longitudinal structure pervades all RMM phases: generally,
cloud cover is higher over the Maritime Continent and western Pacific versus the eastern Pacific and Atlantic Oceans, though the extent to which this is true is modulated by MJO phase. Echo cover is greater over Central and South America (longitude bin 300°) versus the nearby eastern Pacific and Atlantic Ocean. Low clouds (i.e. stratocumulus plus cumulus EO types, black and purple) are more common in the east Pacific Ocean (longitude bins 220°-280°) than all other longitude bins. Deep precipitating EO types (both wide and narrow, orange and red) make their greatest percent contribution to total cloud cover over the Maritime Continent, western Pacific Ocean, and South American longitude bins. One persistent feature through all RMM phases is the presence of the stratocumulus decks off the coast of South America (purple slab in longitude bin 280°, showing about 20-25% coverage).

Perhaps more interesting than these cloud cover commonalities are the differences from phase to phase. Going from phase 2 – 8, the longitude of maximum cloud cover propagates from the Indian Ocean, over the Maritime Continent, into the western Pacific, and finally to South America. In phases 2 and 3, cloud cover is greatest over the eastern Indian Ocean (longitude bins 80° and 100°) and steadily increases over the Maritime Continent (longitude bin 120°). Phase 4 shows a broadening of the maximum cloud cover envelope to the Maritime Continent and west Pacific (longitude bins 120°-160°). These longitude bins remain above average through phases 5 and 6. In phase 6, the max cloud cover envelope shifts farther eastward into the Pacific, extending from longitude bins 120° to 180°. Concurrent with this increase in cloud cover over the Maritime Continent and western Pacific, is a decrease in cloud cover over the Indian Ocean (longitude bins 60° and 80°). Phase 6 is the cloudiest phase tropics-wide, with above
average cloud cover from the Maritime Continent eastward to South America (longitude bins 120° to 340°, excluding longitude bin 280°). By phase 7 and 8, cloud cover over the Maritime Continent and western Pacific wanes. Phase 8 shows increased cloudiness over the east Atlantic, Africa, and the Indian Ocean (longitude bins 360°, and 20° - 100°).

Western Hemisphere cloudiness also appears to be modulated by the MJO, at least by this EOF based definition. Cloud cover is largest over the east Atlantic and west Africa in phases 1 – 2 and phase 8, when the cloud cover is relatively large over the Indian Ocean but small over the western Pacific compared to other phases. These results are consistent with previous studies that found a Western Hemisphere response to the MJO in temperature, wind, and precipitation (Salby and Hendon, 1994; Bantzer and Wallace, 1996). More recently, Martin and Schumacher (2010) found positive precipitation anomalies over the Caribbean during WH04 RMM phase 1 and 2 and negative anomalies during phase 5 and 6, consistent with our Fig. 3.9.
Figure 3.9 – Horizontal pixel (hp) cover for each EO type in 20° longitude bins across the tropics during each of the 8 WH04 RMM phases. The longitude values represent the center value of each 20° longitude bin. Total hp cover is number of cloudy pixels divided by total (both clear and cloudy) pixels. The mean total hp cover over all 8 phases for each longitude bin is represented by the line-plot in each panel. For reference the bottom panel is a geographical map of the tropics. (Fig. 10 of Riley et al. 2011)
Another way to display the same information is as contour plots of cloud cover anomalies (relative to the all-phase mean) for each EO type separately (Fig. 3.10. The wide deep precipitating EO type anomalies account for much of the eastward propagation of the MJO (Fig. 3.10a, note the color scale is different for panel a vs. all others). Positive anomalies start over Africa, the Indian Ocean, and Indonesia in phase 1 – 3, move over the west Pacific in phase 4 – 6, then finally extend into the central Pacific in phases 7 and 8. In phase 7 and 8 the positive anomalies also sprawl the South American continent and Atlantic Ocean where they linger into phase 1 and 2. There is a noticeable break in the eastward propagating positive anomalies near 100° that is also present in the anvil, cirrus, cumulus congestus, and altocumulus types (panels c-f). This break or “jump” in cloud cover is seen in previous observational work (e.g. Knutson and Weickmann 1987, Maloney and Hartmann 1998) and a modeling study by Newman et al. (2009) and is likely associated with topographical influences on convection over the Maritime Continent (Wu and Hsu 2009).

The narrow deep precipitating type also shows eastward propagating positive anomalies, which lead the wide deep precipitating anomalies by one to two phases. Unlike the other EO types (except for cumulus), though, the positive anomalies only traverse the Indian Ocean and Maritime Continent, petering out over the western Pacific during phase 5. Also unique to the wide deep precipitating type (and cumulus type Fig. 3.10h) the positive anomalies sprawl the entire tropics when enhancements are positive over the Indian Ocean and Maritime Continent, though the pattern becomes weaker and noisier in the Western Hemisphere. A possible explanation for the tropic wide positive enhancements is given below.
Figure 3.10 – Anomalous horizontal pixel cover for each EO type (labeled in panels a – h). The color scale for panel (a) goes from -10% to 10% while panels (b-h) go from -5% to 5%. In a (b-h) each contour level is 1% (0.5%) with positive anomalies in solid lines, and negative anomalies dashed. The thick solid line is the zero difference contour. The longitude values represent the middle value of individual 20° longitude bins. Panel (i) is the 1-4 km temperature anomaly associated with all EO types. Contour interval is 0.1°C. A map of the tropics from 15°S – 15°N is provided in the bottom panel as a reference. (Fig. 11 of Riley et al. 2011)
The anvil EO type anomaly pattern resembles the wide deep precipitating EO cloud cover anomaly pattern (cf 3.10 a and c), but shifted westward (or later in time), by mean tropical easterlies (or a long lifetime after convection, as seen in section 3). Unlike the wide deep precipitating type the anvil type has a secondary band that propagates eastward from the central Pacific to South America. The Atlantic Ocean (~320° - 360°) anomaly pattern is opposite the Indian Ocean and Maritime Continent with generally negative anomalies during phases 1 – 5 and positive during phases 6 – 8.

The cirrus EO type (Fig. 3.10d) shows eastward propagation of positive anomalies similar to the wide deep precipitating and anvil types. Positive cirrus cloud cover anomalies start over Africa and the Indian Ocean in phases 1 and 2, spread into the western and central Pacific Ocean during phases 3 and 4, and extend from the Maritime Continent to the eastern Pacific Ocean during phase 5 and 6. By phase 7 and 8, positive anomalies are mainly over the Western Hemisphere, in agreement with Virts and Wallace (2010) who found CALIPSO detected TTL cirrus over South America and Africa in the late WH04 RMM phases. Comparison with CALIPSO results should be taken cautiously, though, as the cirrus Virts and Wallace (2010) discuss has base heights above 15 km, which go mainly undetected by CloudSat (see Fig. 2.1a).

The cumulus congestus EO type (Fig. 3.10e) cloud cover anomalies are similar to the wide deep precipitating and anvil types, though the congestus stops propagating eastward around phase 6, where the positive anomalies extend into the central Pacific. Similar to the anvil and cirrus types, there are generally negative anomalies over South America and the Atlantic in phases 1 – 5 and positive afterwards. The altocumulus (ac) EO type cloud cover anomalies (Fig. 3.10f) propagate approximately in phase with the
wide deep precipitating bands and have a similar Western Hemisphere signal as the other EO types.

The stratocumulus have a much weaker eastward propagating positive signal from the Indian to central Pacific Ocean that slightly lags the wide deep precipitating type. Also, anomalies over the East Pacific and Atlantic Oceans are positive during phases 2 – 6 and negative otherwise. The cumulus EO type exhibits a zonal anomaly structure (Fig. 3.10h) very much like the narrow deep precipitating type (Fig. 3.9b), suggesting that tropics-wide vertical instability for cumulus convection may be modulated by the MJO’s zonal mean temperature signal (Fig. 3.9i).

Fig. 3.10i shows the 1 - 4 km temperature anomaly associated with all EO types across the RMM phases. The depth 1 – 4 km is chosen since Tulich and Mapes (2010) showed that convection was most sensitive to temperature and moisture variations below 4 km. During phase 1 and 2, temperatures tropics-wide are anomalously cool, in line with increased instability and anomalously positive cumulus convection (Figs. 10b, h). From phase 3 to 8, warm temperature anomalies extend from the Maritime Continent eastward, consistent with suppressed cumulus during these phases.

The information from Fig. 3.10 is used to make pictorial realizations of the MJO’s anomalous cloudiness across the tropics as a function of RMM phase. Akin to Fig. 3.5’s purpose, the aim now is to mimic Madden and Julian’s (1972) famous schematic (their Fig. 16, reproduced here as Fig. 3.11b), but again with objective methods and actual CloudSat observations of cloud structure. To make the pictorial realizations, EOs are randomly selected from RMM phase EOs and plotted centered over their true mean longitude (where longitude has been multiplied by 40 to maximize viewing clarity). Here
the random selection is done as follows: wherever anomalous cloud cover in Fig. 3.9a is positive, random EOs are selected from the set of EOs belonging to that longitude bin, RMM phase, and EO type. The amount of cloudiness plotted per EO type is proportional to the anomaly strength in Fig. 3.9 (actually slightly more, since EOs are drawn until the desired amount of echo cover is reached and in practice exceeded).

Fig. 3.11a is an example of one pictorial realization of clouds for RMM phases. With a different seed to the random number generator, another realization can be made. We have made several (see online at http://www.rsmas.miami.edu/users/eriley/Emily/JAS2011_Supplementary_Figs12.html), and selected one based on aesthetic criteria, and consistency with the statistical results. As in Fig. 3.5, wide deep precipitating types are gray-to-red, while all other EO types are gray-to-blue, with color appearing at 0 dBZ. The eastward propagating band of wide deep precipitating is readily seen (red EOs), and the associated anvil type can be noticed. The tropics-wide enhancement of narrow deep precipitating types (tall blue columns) in phases 1-3 is also discernable. Other EO types are certainly present, but their small fractional contribution to total cloudiness makes them harder to evaluate in the pictorial realization.

Comparing this modern diagram to the Madden-Julian schematic of 1972 (Fig 5.11b) is instructive. First it must be acknowledged that they did a remarkable job of inferring cloudiness anomalies based only on spectral analysis of surface data. Their cumulus cartoon icons correspond best to our wide deep precipitating mesoscale convective systems, which carry the bulk of the MJO signal. The slightly lagging anvil clouds, related to the MJO’s tilted structure of moisture (Figs. 3.8b, c and Fig. 3.10) and
heating (Lin et al. 2004), along with the zonal structure of shallow and narrow deep cumulus convection, are, however, more than their methods could have detected.

Figure 3.11 – (a) One example of a pictorial realization of EOs across the tropics during each WH04 RMM phase. Phases run downward from 1 to 8. Each stamp is centered over its true mean longitude 3 40. The longitude scale (x axis) has been multiplied by 40 to maximize viewing clarity. (bottom) A map of the tropics from 158S to 158N is provided for reference to an EO’s true mean longitude. (b) Schematic of the tropics-wide cloud evolution and circulation associated with the MJO (Madden and Julian 1972). (Fig. 12 of Riley et al. 2011)

3.5 Discussion

In terms of pinwheel phases, the figures in this paper provide evidence to the validity of the schematics in Fig. 3.1. In those schematics, cloud structure and evolution were inferred from thermodynamic, dynamic, and precipitation data. Here, the ability of
CloudSat to detect cloud particles augments previous work by providing a direct view of the vertical structure of cloud types across the MJO.

The likely importance of low-clouds followed by mid-topped convection at pre-conditioning the environment prior to deep convection in the most active phase of the MJO has been highlighted in several studies (e.g. Kemball-Cook and Weare, 2001, Kiladis et al. 2005, Tromeur and Rossow 2009, Lau and Wu 2010, and Jiang et al. 2010). Figs 5 and 7 indicate that when MJO phases are defined locally (i.e. the pinwheel phases) low clouds (both stratocumulus and cumulus) have their biggest influence (i.e. largest fractional contribution) on cloud fields prior to MJO active phases, although total low-cloud amount is greater in the active phases. Perhaps, in locally defined phases, they are indeed critical to the low-level warming and moistening (seen in Fig. 3.8) ahead of peak MJO activity. However, mid-topped convection (i.e. cumulus congestus EO type) does not show much variation prior to active MJO phases as defined locally. Cumulus congestus’s fractional contribution peaks during phase 1, then decreases through phase 5 (Fig. 3.4b), leaving their role in moistening the environment prior to MJO active phases unclear from these results. Some cumulus congestus cells are part of wide deep precipitating systems, but then it is hard to ascribe them as precursors.

Contrary to the low-level cloud signal in the pinwheel phases, the stratocumulus EO type positive anomalies trail the wide deep precipitating signal in the WH04 RMM phases (Fig. 3.10g). Why there is a discrepancy between the role low-level clouds play in locally vs. globally defined phases is unclear, but is intriguing and perhaps warrants future study.
Changes in the type of deep convection across MJO events, as seen in Figs. 3.4 and 3.6, suggest the organization of convection is being modulated by MJO phase. The MJO transitions from deep, narrow, less organized convection in the suppressed phases to widespread, more organized convection during active phases, as indicated by fractional contribution of narrow deep precipitating and wide deep precipitating EO types to total cloudiness in Fig. 3.4. High reflectivities above 0°C during phase 2 imply strong updrafts capable of lofting large particles, while a preponderance of high reflectivities below 0°C in phases 3 – 4 indicate widespread rainfall (Fig. 3.6). Tromeur and Rossow (2009) offer an explanation for this change in convective organization: The large scale wave moistens the lower troposphere allowing a transition from smaller-scale, less organized convection to larger-scale, more organized convection. The effects of this convective reorganization could then feedback on the larger scale (Moncrieff 2004). A change in convective scale may also indicate changes in updraft entrainment across the MJO, as Bacmeister and Stephens (2011) pointed out that “it seems intuitively reasonable that as the size of a convective cloud increases the bulk fractional entrainment of environmental air into the cloud will decrease.” A method developed by Luo et al. (2010) to estimate buoyancy and entrainment from A-Train observations may help identify such entrainment changes across the MJO, but is not done here.

To help put MJO cloud and thermodynamic variations in the context of other tropical waves, MJO pinwheel phase results are compared to convectively coupled Kelvin pinwheel phase analysis. EOs are assigned a Kelvin pinwheel phase using the same method as the MJO pinwheel phases (Section 2.2), except using filtered Kelvin OLR (i.e. Kelvin box in Fig. 6b of Wheeler and Kiladis, 1999) between 10°S – 10° N.
For strict comparison, Fig. 3.12 is based on MJO and Kelvin events with amplitude between 2 – 3, which have a standard deviation of OLR of 10.0 Wm$^{-2}$ and 11.6 Wm$^{-2}$, respectively. (MJO results are similar to Fig. 3.6 with amplitude > 2.)

Figure 3.12 – (a-c) same as figure 8 except for the Kelvin wave with amplitudes between 2 – 3. (d) MJO and Kelvin wave difference in EO cover, where the MJO events selected for this plot had amplitude between 2 – 3. Solid (dashed) contours are positive (negative). The zero difference contour is bold. See text for more details. (Fig. 13 of Riley et al. 2011)

In Fig. 3.8, the MJO temperature anomaly structure tilts weakly backward (toward later phases) through the troposphere, while the moisture anomalies show a
comparatively stronger backward tilt. Rather, the cloud field is quite symmetric around peak cover in phase 4, though as mentioned in Section 3 upper-level cloudiness is slightly greater after vs. before phase 4. By comparison the Kelvin wave shows asymmetry around phase 4 in both lower and upper level cloudiness (Fig. 3.12a). Cloudiness at both levels is greater before phase 4 than after. Fig. 3.12d shows the difference between MJO and Kelvin wave echo cover. In phase 0, the MJO has greater cloudiness from about 2 – 15 km. Cool colors in phase 1 – 4 show that the MJO has less cloudiness compared to the Kelvin wave. Warm colors during phases 0, and 5 – 7 show the converse cloud signal. Additionally, for a comparable temperature change across phases, the MJO has a greater change in moisture (q) at low-levels (z < 5 km) than the Kelvin wave (cf 3.12b,c with 3.6b,c), in line with results of Mapes et al. (2006) and Tulich and Mapes (2010) who found greater moisture to temperature ratios for intraseasonal variations compared to shorter time scales (< 2 days). In terms of absolute temperature differences, low-level temperature variations in the Kelvin wave are about 4 times greater than those in the MJO.

The structure of the Kelvin wave also differs from that of the MJO. The Kelvin wave temperature structure is more tilted in the troposphere and its moisture anomalies are about 3 km lower than the MJO. Furthermore, Kelvin anomalies lead MJO anomalies by about one phase, with an exception at low-level (z ≤ 1 km) temperature anomalies, which show a similar pattern in phases 0 – 5. Perhaps the time scale of these two waves types offers clues to the phase and height discrepancies in the moisture field. If the Kelvin phases are thought of as days, while the MJO phases as pentads then the MJO
frequency resonates more with the residence time of water in the atmosphere (~9 days), giving more time to build moisture vertically in the MJO versus Kelvin wave.

The relative importance of moisture to temperature shed light on underlying dynamical differences in the MJO and Kelvin wave. For the Kelvin wave, low-level temperature variations are relatively more important than moisture suggesting convection is more controlled through temperature effects on buoyancy. However, for the MJO, variations in moisture seem to have greater impacts on convection. Further evidence for this view is in Yasunaga and Mapes (2010, submitted), which shows that the MJO modulates small TRMM echo objects, the convective fraction of rainfall, and precipitable water more than Kelvin waves of the same amplitude in rainfall. Their results are interpreted as indicating that convection in divergent wave types (e.g. Kelvin waves) is controlled more by convective inhibition (CIN), while rotational wave types (e.g. the MJO) have more impact through moisture. Raymond and Fuchs (2007, 2009) model results also support this idea. In Raymond and Fuchs (2007) two types of unstable modes are described, a slow moving “moisture mode” and a fast moving “gravity mode” or CIN governed instability. They link the later to the Kelvin wave and, in their 2009 paper, the former to the MJO.

Despite Kelvin wave and MJO differences that may imply different convective coupling mechanisms, their similarities indicate scale invariance to organization of cloud morphology (i.e. shallow to deep to stratiform). Overall, the MJO and Kelvin wave show similar cloud evolution (i.e. Kelvin EO type evolution is very similar to Fig. 3.5). More broadly, the evolution of predominant cloud types in the life cycle of the MJO is similar to all convectively coupled equatorial waves (CCEWs; Kiladis et al. 2009) and even the
life cycle of individual MCSs (cf. 3.1d). The resemblance of the MCS life cycle to the MJO is perhaps explained as the aliasing of shallow, deep, and stratiform parts of the MCS on to larger time and space scales, as discussed in Mapes et al. (2006). Different proportions of shallow, deep, and stratiform clouds occur in individual MCSs embedded within different large-scale wave structures, so in a filtered sense there is an evolution from enhanced shallow convection, to enhanced deep convection, followed by enhanced stratiform cloud on the large-scale. Mapes et al. (2006) refer to this filtered view as the stretched building block hypothesis.

3.6 Summary

Using CloudSat observations we have documented the evolution of total cloud cover, cloud types, temperature, and moisture across the MJO. Broadly speaking, the evolution of predominant cloud types over the life cycle of the MJO is similar to the life cycle of individual MCSs (cf 3.1a, b, and c) and even more broadly to all CCEWs (Kiladis et al. 2009). Such similarity gives credence to the stretched building block hypothesis of Mapes et al. (2006).

Differences between the MJO and Kelvin wave offer clues to underlying dynamical differences in how the wave types are coupled to convection. For a given amplitude in OLR, the MJO affects low-level cloud moisture more than the Kelvin wave, while the Kelvin wave shows larger low-level changes in temperature through the phases (Fig. 3.12). We speculate that this is because the MJO modulates moisture much more than the Kelvin wave, again for comparable temperature, echo cover, and OLR
amplitudes. One interpretation, supported by the findings of Fuchs and Raymond (2007 and 2009) and Yasunaga and Mapes (2010, submitted), is that Kelvin waves are controlled more by CIN, while the MJO more so through moisture variations. Future studies could refine and test this hypothesis through experiments with cloud models on moisture-limited vs. inhibition-limited convection, and through a better understanding of moisture budget processes in the MJO.

In terms of the WH04 RMM phases, the wide deep precipitating, anvil, cirrus, cumulus congestus, and alto cumulus, and stratocumulus EO types showed eastward propagation from the Indian Ocean to the central Pacific. The anvil and stratocumulus band tend to lag the wide deep precipitating type. These propagating EO types also showed a coherent Western Hemisphere signal. Generally, negative anomalies occur over South America and the Atlantic Ocean during phases 1 – 5 or when anomalies are positive over the Indian Ocean and Maritime Continent, with an opposite signal in phases 6 – 8. The stratocumulus type EOs also showed coherent variations over the Eastern Pacific, with positive anomalies during phases 1 – 4, and negative afterward.

The narrow deep precipitating EO type also showed eastward propagation, but only from the Indian Ocean to the Maritime Continent in phases 1 – 4. Narrow deep precipitating anomalies tend to lead the wide deep precipitating anomalies by about 1-2 phases. Along with the eastward propagating feature, positive anomalies tropics wide were enhanced during phases 1 – 3 and suppressed during phases 4 – 8. A similar enhancement, suppression pattern was observed for the cumulus EO type. During phases 1 – 2 (6 – 8) the tropics is anomalously cool (warm) at low levels (Fig. 3.10i), suggesting
that the tropics-wide vertical instability of cumulus convection is being modulated by the MJO’s zonal mean temperature signal.

The pictorial mosaics, Figs. 3.5 and 3.12, offer a novel reality check on how we interpret the multi-scale nature of convectively coupled waves, in this case the MJO. Besides representing the statistics of cloud evolution across the MJO, they also preserve cloud system morphology differences across phases, as well as richness of texture that remind us of the complexity of clouds across large-scale waves. The comparison of CloudSat mosaic to the Madden and Julian’s (1972) schematic shows that, their ability to infer cloud evolution across the tropics was remarkable, but incomplete: there is more to be learned with the detail and richness of CloudSat.

As with all detailed observations, there is a question of how to apply these results to improve models or theories that do not have such details. One avenue that CloudSat offers is a cloud water content product, which provides derived measurements of liquid and ice water content. A similar analysis, as was done here, could be done for the liquid water and ice water content variables and then compared to reanalysis data or model output, similar to what Jiang et al. (2010) did for boreal summer intraseasonal variability events.

Another approach is to use the multiscale modeling framework (MMF) or “superparameterization” as a clean comparison to detailed observations. In superparameterization, 2D cloud system resolving models (CSRM s) are embedded within each general circulation model (GCM) grid cell to serve as the convective parameterization (Grabowski, 2001 and Randall et al., 2003). Several studies (Grabowski 2003, Khairoutdinov et al. 2008, Thayer-Calder and Randall 2009, and
Benedict and Randall 2009) have shown the success of superparameterization in simulating a fairly realistic MJO, yet none of these studies examined cloud morphology per se. In the future we aim to do a similar analysis as in this current observational study, but applied to CSRM output from multiscale model runs of equatorial waves.
CHAPTER 4: The effects of organization on convective and large-scale interactions using cloud resolving simulations with parameterized large-scale dynamics

4.1 Background

As stated in Chapter 1, tropical deep convection is observed to vary on a wide range of time and space scales – from small short-lived single cell systems \((O(10 \text{ kms}), O(\text{minutes} - \text{hour}))\) to larger longer-lived mesoscale convective systems \((\text{MCSs}; O(100 \text{ km}), O(\text{hours} - \text{day}))\) and convectively coupled waves \((\text{CCWs}; O(1000 \text{ km}), O(\text{days} - \text{weeks}))\). While these scales are well observed in nature (e.g. Houze 2004; Kiladis et al. 2009), the full time and space gamut of convection is missing from general circulation models \((\text{GCMs}; \text{Arakawa} 2004; \text{Tao and Moncrieff} 2009; \text{Del Genio} 2011)\). Instead, convection is traditionally represented (or parameterized) by a single or ensemble of steady state plume(s) (e.g. Arakawa and Schubert 1974). Organization on the mesoscale is missing; it is neither resolved nor parameterized, which is unfortunate given MCSs are known to help sculpt the large-scale circulation through their diabatic heating and divergence profiles (Houze 2004, Schumacher et al. 2004). Only a few studies have attempted to account for mesoscale organization in their model formulations (Donner 1993; and Moncrieff 2004), while the NASA Global Modeling and Assimilation Office NSIPP-2 model has a separate anvil parameterization scheme (Wyant et al. 2006 and Mapes et al. 2009). Thus, parameterizing the mesoscale is the “next parameterization frontier” (Del Genio 2011).
To parameterize a process in a model means we need to understand the link between said process and the variables resolved by the model. In terms of convective parameterization, this means relating the convection to environmental properties (Arakawa 2004). Several studies, both with observations (e.g. Halloway and Neelin 2009; Neelin et al. 2009; Sherwood et al. 2010 and references therein) and models (e.g. Tompkins 2001a,b,c; Derbyshire et al. 2004; Mapes 2004; Tulich and Mapes 2010), have discussed the environmental (or large-scale) sensitivities of tropical convection. These works are useful in understanding the convective, large-scale relationship and in guiding cumulus parameterization. However, few studies (e.g. Tobin et al. 2012) have attempted to quantify the relationship between the degree of organization and the environment (or large-scale). The goal of this chapter is to address the second objective of this dissertation to isolate the effects convective organization has on convective and large-scale\textsuperscript{1} interactions.

A cloud systems resolving model (CSRM) with parameterized large-scale dynamics (PLSD) is used to achieve the above goal. This setup has resolved convection, while accounting for the coupling of convection and the larger-scale. The convection and larger-scale should be thought of as a system as the two are interrelated. As mentioned in section 1.2.3, two approaches have been developed to parameterize large-scale dynamics in CSRMS: 1) the weak temperature gradient (WTG) approximation (Sobel and Bretherton 2000; Raymond and Zeng 2005) and 2) the gravity wave approach (Kuang 2008, hereafter K08). This chapter utilizes the gravity wave approach, following K08. The degree of organization will be controlled by altering the CSRM domain size and shape, and addition of shear to the background wind. The hypothesis motivating this

\textsuperscript{1} For this study, the distinction between convective and large-scale refers to the resolved processes within the cloud resolving model and the parameterized large-scale, respectively.
work is that the coupled convective, large-scale system is more unstable to organized convection versus scattered convection. What is meant by “more unstable” is that given a forcing or domain setup, the convective, large-scale system will produce a larger response (e.g. surface precipitation) when convection is organized versus unorganized. For example, using a CSRM with the WTG approximation, Wang and Sobel (2011) showed that, for a given sea surface temperature (SST) forcing, a two-dimensional (2D) CSRM gave a larger precipitation response than a three-dimensional (3D) CSRM (Fig. 4.1), where 2D is arguably more organized than 3D since the flow is more constrained in 2D and has one less dimension to be disorganized along (Bretherton and Smolarkiewicz 1989; Tompkins 2000; Moncrieff 1981, 2004).

The following section discusses the model and methods used in this chapter. An analysis of results is done in Section 4.3. Section 4.4 attempts to explain the results via sensitivity and forcing experiments. Section 4.5 briefly discusses the dependency of the results to a different model setup than that used in Sections 4.3 and 4.4. Section 4.6 discusses the implications of organization on the convective, large-scale relationship. Section 4.7 summarizes and concludes the chapter.

Figure 4.1 – Daily rain rate $P$ (mm/d, solid line) and surface fluxes (latent and sensible heat flux, $E + H$, in the unit of mm/d, dashed line) versus SST for two and three dimensions. Wang and Sobel (2011) Fig.3.
4.2 Methodology

4.2.1 Gravity wave equation test harness

Model simulations are performed using the System for Atmospheric Research (SAM) version 6.8.2, mentioned in Section 2.3 and described in detail in Khairoutdinov and Randall (2003). Section 2.3 also described the basic setup of each model run (i.e. grid spacing, surface type used, and background radiation and vertical velocity profiles (Fig. 2.3).

Following K08, large-scale dynamics are coupled to SAM with the 2D linear gravity wave equation. K08 starts with the 2D anelastic linearized perturbation equations of momentum, continuity, and hydrostatic balance for large-scale waves:

\[
\frac{\partial (\tilde{\rho}u)}{\partial t} = -\frac{\partial \tilde{p}}{\partial x} - \varepsilon \tilde{\rho}u, \quad (4.1; \text{K08 Eq. 1})
\]

\[
\frac{\partial (\tilde{\rho}u)}{\partial x} + \frac{\partial (\tilde{p}w)}{\partial z} = 0, \quad (4.2; \text{K08 Eq. 2})
\]

\[
\frac{dp}{dz} = -\rho g \frac{T'}{T_v}, \quad (4.3; \text{K08 Eq. 3})
\]

where \( \varepsilon \) is the mechanical damping coefficient and all other variables have their traditional meteorological meaning. Primes denote wave perturbation variables, while an overbar denotes background mean variables. For simplicity primes and bars were dropped from further notation. Since only a single large-scale wavenumber, \( k \), is used at a time, wave solutions of the form:

\[
[u, w, T, p](x, z, t) = \text{real}\left\{[\hat{u}, \hat{w}, \hat{T}, \hat{p}](z, t)\exp(-ikx)\right\} \quad (4.4; \text{K08 Eq. 4}),
\]
are assumed. The lower boundary condition is \( w = 0 \). K08 cites Durran (1999, his section 8.3.2) for the radiation upper boundary condition (K08’s Eq. 5). Eqs 4.1 – 4.3 are combined by eliminating \( u \) and \( p \) from each equation. Then Eq. 4.4 is applied to give:

\[
\frac{\partial}{\partial z} \left[ \frac{\partial (\rho \hat{w})}{\partial t} \right] = -k^2 \rho g \frac{T_v'}{T_{v,\text{ref}}} - \varepsilon \frac{\partial^2 (\rho \hat{w})}{\partial z^2}. \quad (4.5, \text{similar to K08 Eq. 6})
\]

For a given \( x = x_0 \), Eq. 4.5 is multiplied by \( \exp(-ikx_0) \). After taking the real component, the equation is:

\[
\frac{\partial^2}{\partial z^2} \left[ \frac{\partial \rho w(x_0,z,t)}{\partial t} \right] = -k^2 \rho g \frac{T_v'(x_0,z,t)}{T_{v,\text{ref}}} - \varepsilon \frac{\partial^2 \rho w(x_0,z,t)}{\partial z^2} \quad (4.6, \text{similar to K08 Eq. 7})
\]

The CSRM is then used to represent the atmospheric column at \( x = x_0 \). \( T_{v,\text{ref}} \) is a fixed domain-averaged reference virtual temperature profile. \( T_v' \) is the deviation of the CSRM domain-averaged virtual temperature from the reference profile. \( T_v' \) is used in Eq. 4.6, along with the boundary conditions to evolve the large-scale wave vertical velocity, \( w \).

Effects of the updated \( w \) are then used as additional source terms of temperature and moisture at each grid point in the CSRM’s thermodynamic tendency equations,

\[
\frac{\partial T}{\partial t} = \frac{\partial T_{\text{conv}}}{\partial t} + w \left( \frac{\partial \tilde{T}}{\partial z} + \frac{g}{c_p} \right) - \varepsilon_T \quad (4.7)
\]

\[
\frac{\partial q}{\partial t} = \frac{\partial q_{\text{conv}}}{\partial t} + w \left( \frac{\partial \tilde{q}}{\partial z} \right) - \varepsilon_q \quad (4.8)
\]

, respectively. Terms with subscript “conv” are the temperature and moisture tendencies resulting from explicitly resolved cumulus convection inside the CSRM. Terms multiplied by \( w \) are the additional large-scale wave terms. Overbars on \( T \) and \( q \) indicate domain averaged temperature and moisture. \( \varepsilon_T \), and \( \varepsilon_q \) represent large-scale damping of
domain-averaged T and q, applied uniformly in the horizontal. The computed
temperature and moisture tendencies are then used to get a new CSRM virtual
temperature. The deviation of the new domain averaged virtual temperature from the
reference profile is used as $T'_v$ in Eq. 4.6 to complete the interaction between the
convection and wave. The ramifications of $\varepsilon$, $\varepsilon_T$, and $\varepsilon_q$ are discussed fully in Appendix A. Simulations were done two ways with respect to Eqs. 4.7 and 4.8: 1) The large-scale
wave vertical velocity, $w$, advected the reference T and q profiles, such that $\bar{T} = T_{ref}$ and
$\bar{q} = q_{ref}$ and 2) $w$ advected the CSRM’s current domain-averaged T and q profiles, such
that $\bar{T}$ and $\bar{q}$ are functions of time. Section 4.5 discusses the differences between the two
sets of simulations.

For clarity, Fig. 4.2 conceptually explains the CSRM-wave coupling. SAM acts
as a vertical atmospheric column in various phases of the large-scale wave (Fig. 4.2 top). The CSRM virtual temperature is used to evolve the large-scale vertical velocity equation
at every time steps. The large-scale vertical velocity is then added as an additional term
to the CSRM’s advection equations (Fig. 4.2 bottom) to complete the convection, large-
scale interaction. In this set-up, the convection evolves under the influence of the large-
scale vertical velocity, which is in turn influenced by the convection.
Figure 4.2 – Schematic explaining coupling of convection inside the cloud systems resolving model (CSRM) and the linearized large-scale 2D gravity wave. The CSRM represents the atmospheric column at \( x = x_0 \) in the large-scale wave. The CSRM domain averaged \( T'_v \) updates the large-scale vertical velocity (Eq. 4.5). The new \( w' \) is then added as an additional advective term in the CSRM’s temperature and moisture tendency equations (Eq. 4.6 and 4.7).

4.2.3 Model Setup

K08 found that when PLSD was turned on, convection spontaneously developed wave-like oscillations (Fig. 4.3). The amplitude and period of the oscillations were dependent on the large-scale wavelength choice, as wave amplitude and phase speed decreased as wavelength increases (Fig. 4.3 and his Fig. 7). Since the aim here is to isolate the role convective form has on the coupled convection, large-scale system, a single large-scale wavelength, 5000 km, is chosen for all model runs to eliminate any dependence on wavelength choice in the interpretation of results.

Various CSRM domain sizes with 2D and 3D setups are being used to sculpt the form of convection. Additional details of model setup are in Chapter 2.3. As a reminder, though, domains are doubly periodic (or periodic in \( x \) for the 2D setup). Altering the
model geometry forces the convection to align within the limits of the horizontal dimensions. Domains include 3D isotropic geometry, 3D anisotropic cuboid (or what some papers refer to as 2.5D or bowling alley, Tompkins 2001) geometry, and strict 2D. Tompkins (2000) and Wang and Sobel (2011) previously explored the impact of dimensionality – 2D vs. 3D – on CSRM simulations. Both found 2D geometry encouraged mesoscale organization, while 3D promoted scattered convection.

For experiments where the domain shape is changed, the base 3D domain is 128 km x 128 km. The 128 km isotropic domain was chosen out of necessity due to initial computational limitations of the UM supercomputer and difficulty with restart files. Successive domains doubled the x-dimension and halved the y-dimension until 2D (x = 8192 km) is reached. This doubling and halving approach maintained the same number of grid points per domain set-up, made comparisons between results fair, and was consistent with previous methods (e.g. Tompkins 2000). Throughout the text, the 2.5D domains will be referred to as “stretched” domains.

All runs retain 64 stretched vertical levels. Idealized shear profiles were also used to sculpt the form of convection, as several studies have highlighted the importance of shear in organizing convection into bands or lines (see Section 2.4.3). The shear profile is unidirectional with 10 ms$^{-1}$ zonal wind at and above 800 hPa that decrease to zero at the surface. When no shear profile is prescribed, background horizontal wind at all vertical levels is set to zero.

In all simulations domain mean horizontal wind was relaxed to the prescribed background horizontal profile on a 1-hour timescale. Initially the relaxation timescale was 20 days, which allowed shear to form in the 512 x 32 km and further stretched
domains. Since we want to differentiate the organizing effect of shear versus “2D-ness,” shorter damping time scales were tested until unwanted shear production was eliminated. As mentioned in Chapter 2.3, the surface wind was fixed at 5 ms\(^{-1}\) in the bulk aerodynamic formula to prevent WISHE. Table 4.1 shows that domain averaged 2-meter u-wind is an order of magnitude greater in the highly stretched (or more 2D) domains than in the more 3D domains. Fixing |V| in Eqs. A.1 and A.2 eliminates the possibility that changes in precipitation oscillations across domain shapes could be due to wind, surface flux feedbacks. A 100-day run with no coupling to wave dynamics was done for each domain setup. The last 50-days were averaged to define the initial temperature and moisture sounding for coupled simulations of each respective domain shape. For subsequent simulations where coupling was used, SAM was run as a stand alone CSRM for 30 days to allow the temperature and moisture profiles to reach equilibrium, after which coupling to the large-scale gravity wave was turned on. Days 20 – 30 were averaged to give the reference profile in Eq (4.6).
Figure 4.3 – Domain-averaged precipitation as a function of time after coupling to gravity wave is activated for wavelengths of (top to bottom) 2000, 2857, 5000, 6667, 10000, 13333, and 20000 km. Figure 1 of K08.

<table>
<thead>
<tr>
<th>Domain Setup</th>
<th>$\mathbf{u}_{2m}$</th>
<th>$\mathbf{v}_{2m}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>128 km x 128 km</td>
<td>$3.9 \times 10^{-3}$ ms$^{-1}$</td>
<td>$3.9 \times 10^{-3}$ ms$^{-1}$</td>
</tr>
<tr>
<td>shear</td>
<td>9.9 ms$^{-1}$</td>
<td>7.6 $\times 10^{-3}$ ms$^{-1}$</td>
</tr>
<tr>
<td>256 km x 64 km</td>
<td>$6.6 \times 10^{-3}$ ms$^{-1}$</td>
<td>$4.2 \times 10^{-3}$ ms$^{-1}$</td>
</tr>
<tr>
<td>512 km x 32 km</td>
<td>$1.9 \times 10^{-2}$ ms$^{-1}$</td>
<td>$4.7 \times 10^{-3}$ ms$^{-1}$</td>
</tr>
<tr>
<td>1024 km x 16 km</td>
<td>$3.3 \times 10^{-2}$ ms$^{-1}$</td>
<td>$5.9 \times 10^{-3}$ ms$^{-1}$</td>
</tr>
<tr>
<td>2048 km x 8 km</td>
<td>$2.9 \times 10^{-2}$ ms$^{-1}$</td>
<td>$3.8 \times 10^{-3}$ ms$^{-1}$</td>
</tr>
<tr>
<td>8192 km</td>
<td>$2.0 \times 10^{-2}$ ms$^{-1}$</td>
<td>0.0 ms$^{-1}$</td>
</tr>
</tbody>
</table>

Table 4.1 – The domain and time (post-coupling days only) averaged u- and v-wind at two meters for the indicated domain setup. The values correspond to simulations shown in Fig. 4.6.
4.3 Results

All simulation results shown in this section were done with both the large-scale wave vertical velocity, $w$, and background vertical velocity (Fig. 2.3a) advecting the time independent reference $T$ and $q$ profiles. Figure 4.4 shows 15-minute averaged snapshots of surface precipitation for the 128 km x 128 km domain, the 128 km x 128 km domain with a shear, and the 512 km x 32 km domain (panels a, b, c, respectively). In each panel, a single domain snapshot is repeated four times and stitched together to emphasize the periodicity of the domain. The appropriate xy aspect ratio is maintained for each domain, such that the x-axis is four times as long as the y-axis in the 256 km x 64 km domain. The point of the figure is to emphasize the form of the convection and not necessarily the intensity, so rain rate values are not labeled, though the same contour levels are used in each panel.

In the 128 km x 128 km domain with no shear (Fig. 4.4a), convection takes the form of blobs scattered around the domain. With the addition of shear (Fig. 4.4b), linear, squall line, structures appear in the domain. These squall lines go though life cycles of growth and decay over approximately 1 – 2 hours, and in some cases shows bow echoes. Stretching the x-dimension and shrinking the y-dimension, the 512 km x 32 km domain (Fig. 4.4c), results in blobs spaced far apart from each other in the x-dimension, yet spanning the width of the y-dimension. In terms of periodicity, this results in an infinite line of connected convecting blobs. Surface divergence is noticeable in each panel, via the overlaid surface wind arrows, around the heaviest precipitating regions, indicating the occurrence of downdrafts. In the 512 km x 32 km domain, the surface wind occurs more
strongly in the x-direction than y-direction. The addition of shear or stretching of the
domain both leads to linear convective organization, though the shear produces a
coherent convecting structure, while the stretched domain produces a row of individual
blobs. In summary, comparing the panels, the addition of shear or the change in domain
shape does indeed impact the organization of the precipitation and indicates the
usefulness of our approach.

Figure 4.4 – 15-minute averaged precipitation for a) 128 x 128 km domain, b) 128 x 128 km
domain with 10 m/s uni-directional shear, and c) 256 x 64 km domain Four copies from each
simulation have been stitched together to emphasize the periodicity of the model boundaries. The
appropriate x-y-axis ratio is maintained for each domain, such that in (c) the x-axis is four times
as long as the y-axis. The contour levels in each panel are the same.
4.3.1 Characteristics of coupled simulations

Figure 4.5 shows the time series of SAM domain-averaged surface precipitation for the continuum of domain shapes – 128 km x 128 km, 256 km x 64 km, 512 km x 32 km, 1024 km x 16 km, 2048 x 8 km, and 2D 8192 km (panels a and c – g). Results for the 128 km x 128 km with shear are also shown (panel b). For all runs a damping timescale was set to 10-days for \( \varepsilon, \varepsilon_T \) and \( \varepsilon_q \) in equations 4.6, 4.7, and 4.8, respectively. A 10-day timescale was largely chosen to be consistent with K08, as well as being a reasonable choice compared to observations of momentum damping (e.g. Lin et al. 2005 and Lin et al. 2008). Other damping timescales were tested and are discussed extensively in Appendix A. The decision to apply all three dampings is also discussed in Appendix A. In brief, applying all three dampings kept precipitation values at a more reasonable level than applying just the momentum damping as K08 did. Additionally, results across domain shapes are independent of applying a momentum damping only, or a combination of momentum and temperature damping, or momentum and moisture damping.

For all domain setups the pre-coupled mean precipitation is similar, near 9.9 mm day\(^{-1}\), though the standard deviation increases as the domains are stretched, with the exception of the 2D 8192 km domain (Table B.1). The increase in pre-coupled precipitation noise does not account for the differences in post-coupled precipitation oscillations. A set of experiments was done holding the y-dimension fixed at 128 km, while doubling the x-dimension. This had the effect of decreasing the pre-coupled standard deviations in the stretched domains to remain close to the 3D isotropic domain value, while post-coupled results remained similar (details in Appendix B).
After coupling to large-scale dynamics on day 30, precipitation oscillations develop in all domain setups, though the nature of the oscillations is dependent on the aspect ratio of the domain. After day 40 in the 128 km x 128 km domain, precipitation oscillates between approximately 5-day rainy periods and 5-day dry periods, punctuated by a hiccup of precipitation about one day after primary rain events (Fig. 4.5a). The 10-day precipitation period equates to an approximate phase speed of 6 ms⁻¹. The rain events are non-symmetric with a longer build-up vs. decay of precipitation. These oscillations are different to those in K08 who showed precipitation oscillations about the
pre-coupled mean after coupling to a 5000 km large-scale wave in a 192 km x 192 km and 384 km x 384 km CRM domain (Fig. 4.3). K08’s oscillations did not show high precipitation excursions followed by dry periods. Differences are partially explained by the large-scale wave vertical velocity advecting the reference vs. time dependent T and q profiles, as Section 4.5 discusses.

Figure 4.6b shows domain-averaged precipitation with time in the 128 km x 128 km domain with uni-directional shear. Coupled precipitation oscillations are similar to those in the no shear simulation. Appendix A discusses how the difference between shear and no shear in 3D isotropic domains is dependent on damping timescale choice. For the “realistic” 10-day damping timescale, though, it appears organizing convection via shear has little or no effect on the coupled convective, large-scale system.

Precipitation events in the 256 km x 64 km domain to 1024 km x 16 km domain have similar characteristics to the 128 km x 128 km shear and no shear domains; buildup and decay about maxima in precipitation are not symmetric with a slower buildup vs. decay and a hiccup of precipitation occurs about one day after primary rain events end. Stretching the domain setup from 128 km x 128 km to 256 km x 64 km and then to 512 km x 32 km results in increased amplitude precipitation oscillations (cf. 4.5a, 4.5c, and 4.5d). The amplitude of precipitation oscillations in domains stretched beyond 512 km x 32 km progressively decrease (cf. 4.5d – 4.5g). Interestingly, the strength of the hiccup in precipitation appears to be independent of the magnitude of the primary rain event with the secondary precipitation events having the same magnitude despite stronger primary events in the 256 km x 64 km to 1024 km x 16 km domains. A possible explanation for the relationship of hiccup precipitation events to primary rain events is given in the next
For all domains, precipitation oscillations after day 40 tend to reach and maintain constant amplitude.

Figure 4.6 summarizes the dependence of post-coupled precipitation response and wave packet speed on domain setup and hence the morphology of convective organization. Maximum post-coupled precipitation increase from the 128 km x 128 km domain till the 512 km x 32 km domain, after which maximum precipitation decreases. Wave packet speed was calculated as the elapsed time between regressed precipitation cycles. The oscillations are characterized by a wave packet speed because there are effectively two wave periods in each domain – the time between primary and secondary rain events and the time between successive primary rain events – that remain regular throughout the simulations. In other words, the primary and secondary rain events could be thought of collectively as one coherent event, or wave packet.

Figure 4.6 – Indicated domain setups versus (a) the post-coupled maximum precipitation value and (b) wave packet periodicity. See text for how wave packet speed was calculated. The corresponding y-axis for each domain, listed from left to right is: 128 km, 128 km, 64 km, 32 km, 16 km, 8 km, and 2D.

4.3.2 Vertical structure
Figures 4.7 – 4.13 show the vertical structure of temperature, moisture, the apparent heat source or convective heating (Q1), apparent moisture sink (Q2), and vertical pressure velocity of the simulated wave oscillations for each domain setup. Yani et al. (1973) define Q1 and Q2 as:

\[ Q_1 = Q_R + L(c - e) - \frac{\partial (\overline{sw})}{\partial z}, \] (4.9)

\[ Q_2 = L(c - e) + \frac{\partial (\overline{qw})}{\partial z}, \] (4.10)

respectively, where \( Q_R \) is the radiative heating or cooling of the atmosphere, \( L(c-e) \) is the latent heating or cooling from phase changes of water, and \( s \) is the dry static energy (\( s = c_p T + gz \)).

Similar to K08, figures 4.7 – 4.13 were constructed by computing the regression coefficients of various fields (i.e. temperature, moisture, convective heating and moistening, and vertical velocity) onto precipitation anomalies at various lags. All the regression coefficients were scaled by the standard deviation of precipitation for a given domain setup. Commonalities across domain setups include: an anomalously cool-warm-cool temperature structure through the troposphere during peak rainfall with an anomalously dry boundary layer and moist anomalies above peaking around 600 hPa (Figs. 4.7 – 4.13b,c). Similar to K08, the moisture anomalies are more tilted than the temperature anomalies (see his Fig. 9). Anomalously cool and dry conditions peak in the boundary layer several hours after peak rainfall associated with spreading cold pools from rain evaporation and convective downdrafts (e.g. Lin and Johnson 1996). Convective heating (Q1) above the boundary layer and cooling below accompany drying through most of the troposphere (i.e. positive Q2 values) during peak precipitation (Fig.
Drying anomalies in the boundary layer are approximately in quadrature with the moisture anomalies, with Q2 anomalies leading moisture anomalies by a day or two (Figs. 4.7 – 4.13c,e). There is less tilt in positive Q2 anomalies compared to the moisture anomalies, resulting in positive Q2 anomalies leading dry anomalies by several days in the free troposphere. Negative Q2 anomalies (or moistening) exhibit a more tilted structure, following more closely the structure of positive moisture anomalies throughout the troposphere (Figs. 4.7 – 4.13c,e). Broadly, the vertical structure around peak rainfall of these simulated waves are similar to what K08 found in his simulations (his Fig. 9) and to CCWs in nature (Kiladis et al. 2009, Riley et al. 2011, cf. 3.8 and 3.12 to 4.7 – 4.13).

For Q1 and Q2, the negative phases of the wave are weaker than the positive phase, while the opposite is true for vertical pressure velocity. This wave structure is in accordance with the precipitation evolution seen in panel a of figures 4.7 – 4.13, which show larger positive vs. negative precipitation anomalies. The larger positive precipitation anomalies are because precipitation cannot go negative, so there is a lower bound on precipitation whereas an upper bound is not set (Fig. 4.5). By contrast, K08 found similar magnitude anomalies for the negative and positive phases of his wave simulations (cf. 4.5 to his 9). Corresponding to the progression of increased precipitation response, as the domains are stretched from 128 km x 128 km to 512 km x 32 km the magnitude of the wave anomalies increase (cf. 4.7, 4.9, and 4.10).

Simulations done in this study show a more complicated vertical structure in temperature anomalies compared to K08 and observationally based CCW composites. Approximately two days after peak rainfall temperature anomalies in the 128 km x 128
km to 2048 km x 8 km domains show a four level structure with cool-warm-cool-warm anomalies. Low-level warm anomalies are centered near 700 hPa and cool anomalies from 500 hPa to 200 hPa in the 128 km x 128 km and descend slightly (~50 – 100 hPa) as the domain is stretched to 2048 km x 8 km. The four level structure anomalies are followed a day or two later by weaker opposite signed anomalies. At this time, 3 – 4 days after peak rainfall, the hiccup or rebound in precipitation occurs in the 128 km x 128 km to 1024 km x 16 km domains.

Figure 4.7 – Composite wave structure for the 128 km x 128 km domain: (a) precipitation, left y-axis and latent heat flux (LHF) and sensible heat flux (SHF), right y-axis, (b) temperature, (c) specific humidity, (d) convective heating, (e) convective drying, and (f) vertical pressure velocity. Contour intervals are indicated above each plot. Negative contours are dashed and the zero contour is thick. Wave packet periodicity is shown in in panel (a). All anomalies are scaled by the standard deviation of precipitation. See text for how wave packet speed was calculated.
Figure 4.8 – Same as Fig. 4.7 except for the 128 km x 128 km domain with shear.
Figure 4.9 – Same as Fig. 4.7 except for the 256 km x 64 km domain
Figure 4.10 – Same as Fig. 4.7 except for the 512 km x 32 km domain.
Figure 4.11 – Same as Fig. 4.7 except for the 1024 km x 16 km domain.
Figure 4.12 – Same as Fig. 4.7 except for the 2048 km x 8 km domain. The wave packet speed shown corresponds to the 15-day time period, while 11.6 ms$^{-1}$ corresponds to the 5-day period (Fig. 4.6). See text for details.
Since the secondary precipitation event and details of the wave’s vertical structure are somewhat smoothed out in the regressions, figures 4.14 and 4.15 show the full time evolution of precipitation, temperature, moisture, and vertical velocity anomalies for the 128 km x 128 km domain and the 1024 km x 16 km domain, respectively. Figures 4.14 and 4.15 clearly show that immediately following the primary rain events air subsides through the entire tropospheric column and from 1 km – 6 km strongly warms and dries as rainfall goes to zero (e.g. day 64 – 65 Fig. 4.14 and day 68 – 69 Fig. 4.15). At the surface, though, moisture anomalies are slightly positive, perhaps due to the fact that positive LHF peaks at this time (Figs. 4.7 – 4.11), so the surface remains moist. The
hiccups in precipitation follows (e.g. centered on day 66 Fig. 4.14 and day 70 Fig. 4.15) corresponding to a warm boundary layer overlaid by cool anomalies with weakened low-level dryness and weak upward motion. As mentioned above, these secondary rain events tend to be the same magnitude despite the primary rain events becoming stronger as the domains are stretched from 128 km x 128 km to 512 km x 32 km (Fig. 4.5). This is perhaps due to the fact that the dry stable layer from 6 km – 10 km corresponding to the secondary rain events maintains the same magnitude across domains.

![Image](image_url)

Figure 4.14 – Time evolution for a 128 km x 128 km simulation of, from top to bottom, domain averaged precipitation, temperature anomalies, specific humidity anomalies, and vertical velocity anomalies.
Figure 4.15 – Same as Fig. 4.14 except for the 1024 km x 16 km domain. Note the color scale is folding over in the moisture anomaly plot (3rd from top). Black and green values inside brown are actually less than -3 g/kg.

Figure 4.16 shows 3D snapshots of precipitating water taken each day from days 40 – 50 of a 128 km x 128 km no shear simulation. Day 40 is the peak of a secondary rain event (Fig. 4.16i). Low to mid-level topped convection is precipitating at this point, but presumably cannot overcome the subsiding warm, dry stable layer from 6 km – 10 km to form deep convection and create another primary rain event (cf. 4.14 day 66 and 4.16 day 40). Therefore, precipitation wanes during days 41 and 42, then slowly builds back up as moisture anomalies build upwards over the next several days (cf. 4.14 day 69 – 72 and 4.16 days 43 – 46). Day 47 is the peak in the primary precipitating event and shows a deep convective system encompassing most of the domain. Afterwards,
precipitation decays over day 48, almost completely disappearing during day 49, only to recover slightly at day 50. At this time, the cycle starts over again.

Figure 4.16 – 3D snapshots of precipitating water (rain + snow) in the 128 km x 128 km domain at the indicated day. Units are g/kg. The last panel is the domain averaged precipitation from days 40 – 50.

Thus far we have shown that when a 10-day damping timescale is used, organizing convection via shear has little effect on the coupled convective, large-scale system, as the precipitation response and vertical structure of the wave oscillations are similar. However, forcing convection to organize in infinite lines of closely spaced convective regions in the y-direction, yet widely spaced regions in the x-direction (Fig.
4.4) leads to a larger precipitation response as the domains are stretched from 128 km x 128 km to 512 km x 32 km. Beyond 512 km x 32 km, precipitation response and the magnitude of the vertical structure anomalies start to decrease. Appendix A shows that the decrease in precipitation response beyond a given x, y-axis ratio occurs regardless of damping timescale choice.

We next ask: *Why do precipitation oscillations decrease in magnitude for domains stretched beyond a given x, y-axis ratio?* In other words, why is there an optimum domain shape? If the hypothesis, which states the coupled convective, large-scale system is more responsive to organized vs. scattered convection holds then the fully 2D domain, which arguably gives more organized convection vs. 3D domains (i.e. Tompkins 2001 and Wang and Sobel 2011) should give the largest precipitation oscillations. One possible explanation is that the mean state of the different domain setups are impacting the wave oscillations and causing the most stretched domains (i.e. stretched beyond 512 km x 32 km for a 10-day damping) to have lower precipitation response than the more 3D domains. Perhaps there is a competition between the degree of organization and the environmental mean state, such that a threshold exists wherein the domain environment overwhelms the effects of organization. Another possibility is that the various domains have different sensitivities to a given forcing.
4.4 Explaining the xy aspect optimum

4.4.1 Mean State Differences across domains

To test whether the difference in mean state temperature and moisture profiles of the various domains impacts the CSRM-wave coupled system, a set of simulations were done fixing the reference T and q profiles (collectively, $T_{v_{ref}}$ in Eqn. 4.5) to the 128 km x 128 km uncoupled average T and q profiles. Figure 4.17 shows the temperature and moisture differences of the average upcoupled 128 km x 128 km shear and increasingly stretched domain soundings from the 128 km x 128 km uncoupled sounding. Recall, a long 100-day uncoupled run was done for each domain setup and the average temperature and moisture profile over the last 50-days were used as the initial sounding in each respective domain (Section 4.2.3). The 128 km x 128 km shear mean state is more unstable and drier through almost the entire troposphere compared to the no shear mean state. Only near the surface is the shear setup more stable than the no shear setup. The 256 km x 64 km temperature and moisture profile shows little difference compared to the 128 km x 128 km domain. As the domains are stretched from 512 km x 32 km to fully 2D at 8192 km, the mean state of each domain is progressively more unstable below 1 km and from 5 km – 15 km and more stable from 1 km – 5 km and drier at all levels than the 128 km x 128 km domain. The exception is the 8192 km temperature sounding from 1 km – 4 km, which lies between the 512 km x 32 km, 1024 km x 16 km, and 2048 km x 8 km curves.
Because of the differences between the 128 km x 128 km domain and all other domains, the large-scale wave vertical velocity, $w$, in Eqn. 4.5, effectively advects a moister, yet more stable (in the vertically averaged sense) reference profile in the stretched domains when the T and q reference profiles are fixed at the 128 km x 128 km profiles. Fig. 4.18 compares the precipitation response for simulations using the 128 km x 128 km mean state temperature and moisture profiles as the reference profile to those that used day 20 – 30 as the reference profile. There is little, if any change in the character of the precipitation oscillations between simulations that used the 128 km x 128 km T and q profiles vs. those that calculated their own T and q profile. Appendix C describes tests in which the sensitivity of the precipitation response to an artificial T and q reference sounding is measured and how that sensitivity varies depending on damping timescale choice. For simulations done with a 10-day damping timescale, artificially moistening or drying each domain’s respective uncoupled T and q sounding and using that as the reference profile did not change the character of the precipitation oscillations. No change occurs because despite fixing the reference profile across domain setups, the
coupled system is able to adjust to maintain the same relationship between the CSRM’s $T_v$ and the large-scale wave’s vertical velocity (Fig. C.9). A change in the mean environment, therefore, does not explain the systematic change in precipitation oscillations moving from the 1024 km x 16 km domain to the 2048 km x 8 km domain.

Figure 4.18 – Domain averaged precipitation as a function of time for the (a) 256 km x 64 km domain, (b) 512 km x 32 km domain, (c) 1024 km x 16 km domain, (d) 2048 km x 8 km domain, and (e) 8192 km 2D domain. Black lines indicate simulations that used days 20 – 30 to establish the reference temperature (T) and moisture (q) profiles. Red lines used the 128 km x 128 km no coupled sounding as the reference T and q profiles.
4.4.2 Sensitivity to forcing experiments

The fact that the CSRM $T_v$ influences the large-scale wave $w$, which in turn influences the CSRM $T_v'$, and the cycle repeats makes it impossible to assert cause and effects statements concerning one scale forcing the other. That is, we cannot say the large-scale is a forcing and the convective scale a response or visa versa. Therefore, traditional type forcing experiments (e.g. Xu et al 1992; Tulich and Mapes 2010; Jones and Randall 2011) are useful to examine how one scale impacts the other. Can different convective sensitivities to a prescribed large-scale vertical velocity explain why the 512 km x 32 km xy aspect ratio gives the largest precipitation amplitude?

To test the above question, the CSRM was subject to prescribed chirps (or pulses) of large-scale vertical velocity. Therefore, we refer to this set of experiments as chirp simulations. Output was saved every 30-minutes, compared to 3-hour in Section 5.3. As before, the chirp simulations were run for 100-days with days 20 – 30 used to define the reference $T$ and $q$ profiles. The reference $T$ and $q$ profiles were still calculated to be consistent with simulations done in Section 4.3, wherein the background vertical velocity (Fig. 2.3a) advected the reference $T$ and $q$ profiles. After day 30, the CSRM was subject to short vertical velocity chirps centered at low-levels. The first chirp was centered on day 35. These chirp simulations are a crude analogy to Kuang’s (2010) “linear response functions,” wherein a response matrix is constructed by applying steady temperature and moisture perturbation forcings to a CSRM and recording the equilibrium response in the state vector. Kuang (2012) found interesting differences in the response functions for a 128 km x 128 km domain that favored unorganized convection and an elongated 2048 km
x 64 km domain that favored more organized convection. For the 128 km x 128 km domain, the convective heating/moistening sensitivity to temperature and moisture anomalies resembled a parcel view of convective overturning, whereas in the 2048 km x 64 km domain temperature and moisture anomalies resembled a layer mode of convective overturning. Instead of constructing the linear response function for each domain, we use the simple chirp simulations to differentiate the convective response of each domain, and leave the computation of the response functions for a future study.

Each chirp was Gaussian shaped in time and sinusoidal in height. The full time-period of each chirp was 10-days, but positive vertical velocity lasted for only 30-minutes with compensating subsidence evenly spread over the remaining time, such that the 10-day time integrated vertical velocity is zero. Figure 4.19 shows the structure of the vertical velocity chirp. The vertical structure is a half sine wave that peaks at 2.5 km and equal to zero above 5 km. A low-level perturbation was chosen to be consistent with the stratiform instability mechanism of Mapes (2000 and 2004) in which low-level perturbations control deep convection. Note, negative values are not shown in figure 4.19 due to the large difference between positive and negative vertical velocity values. For each domain setup the large-scale vertical velocity chirp acted on the 128 km x 128 km T and q reference profile to ensure each domain setup received the same large-scale temperature and moisture tendency (i.e. the large-scale terms, second terms, in Eqs. 4.7 and 4.8 are the same across different domain setups). Since the chirps were turned on at day 30, each chirp simulation experienced 7-chirps, which can be thought of as ensemble members, since the 70-day simulation is essentially equivalent to doing seven separate 10-day simulations. Arguably, seven ensemble members is not very many, as Tulich and
Mapes (2010) had 45 ensemble members for their forcing experiments, but results are robust, as shown for example by the fact that they are smoothly varying with domain shape.

Figure 4.19 – Profile of idealized forcing pulse. The pulse is repeated every 10 days starting on day 34. The inset is a zoomed in image of the pulse. As the text states, the full width at half maximum of the Gaussian is set to 15 minutes.

Ensemble averages the 10-day chirps were computed for precipitation, temperature, moisture, Q1, and Q2. Anomalies across the temperature, moisture, Q1, and Q2 ensemble average were then computed such that the 10-day integral of each anomaly at a specified height is equal to zero. Figure 4.20 shows the ensemble average precipitation response with time for each domain. Time is relative to the peak in the vertical velocity Gaussian. Only 6-hours before and after peak vertical velocity are shown to emphasize the convective response to the positive large-scale vertical velocity. [Note, the 128 km x 128 km shear case is not shown because post-simulation analysis revealed that the simulation had different large-scale moisture and temperature tendencies (second term Eqs. 4.7 and 4.8) than the other runs] As expected by the design of the
forcing, the precipitation responses are shaped like Gaussians. For all domains there is about an hour lag between maximum vertical velocity and peak rainfall. Generally, maximum precipitation rate decreases as the domains are stretched to be more 2D. The exception is the 256 km x 64 km domain, which actually has a higher precipitation rate than the 128 km x 128 km domain. As maximum precipitation rate decreases, though, the width of the precipitation Gaussian tends to increase. For reference, a horizontal line was added at 10 mm/day. The choice for the reference line was arbitrary, but based on the eyeball average of precipitation across all domains during hours -6 – 0. The amount of time precipitation stays above 10 mm/day increases as the domains become more 2D.

For brevity, only the Q1 anomalies are shown (Fig. 4.21). All other composites are shown in Appendix D. For all domains, heating peaks 30 minutes after the vertical velocity maximum (hour zero) near 800 hPa. The 256 km x 64 km domain has the most intense, short-lived convective heating response of all domain setups. As domains are stretched beyond 256 km x 64 km the convective heating response systematically
weakens and lengthens at all levels. Consistently, top-heavy cooling starting around hour 1.5 is most intense in the 256 km x 64 km domain and weakens as domains are stretched. Convective drying profiles in Appendix D show a similar systematic change in intensity as the heating profiles.

The picture that emerges is that as the domains transition from 3D to 2D the convective heating response transitions from a relatively intense local (near 800 hPa) and non-local response (upper-level) that is immediately partially compensated for by a relatively intense upper-level cooling to a more local, weaker heating response that is balanced by cooling beyond the 6-hour time interval shown. It seems deep convection in the more 3D domains is more capped and/or has more stored instability (CAPE) than in the more 2D domains, such that when an intense external forcing is applied, the energy is released via convection extending far above the imposed vertical velocity chirp. This picture is consistent with the change in precipitation response across domains in figure 4.20.

Two important points are gained from these chirp simulations. First, the change in precipitation oscillations across various domain setups in the coupled CSRM-wave framework is not simply explained by sensitivity differences across domain shapes. Rather, the coupled system conspires to give optimum wave amplitude response in the 512 km x 32 km domain. Second, they offer clues to the response of deep convection to low-level perturbations in various environmental regimes. Recall, as the domains became increasingly 2D their mean sounding became drier, yet more unstable (in the vertically averaged sense) (Fig. 4.17).
Figure 4.21 – Composites of anomalous heating (Q1) for chirp simulation described in Section 4.4.2 for the (a) 128 km x 128 km, (b) 128 km x 128 km + shear, (c) 256 km x 64 km, (d) 512 km x 32 km, (e) 1024 km x 16 km, (f) 2048 km x 8 km and (g) 8192 km domains. Hour zero is the peak in the prescribed large-scale vertical velocity forcing profile. Black contours are overlaid every 6 K/day. The zero-contour is thickened and dashed contours represent negative values. Each panel uses the same color scale.
4.5 w advects current vs. reference T and q profiles

This section discusses the robustness of results found in section 4.3 to simulations where the large-scale vertical velocity advected the time dependent temperature and moisture profiles, as opposed to the time independent temperature and moisture reference profiles. Figure 4.22 is the same as 4.6, except the large-scale wave vertical velocity, w in Eq. 4.5, advected the CSRM’s current domain-averaged T and q profiles of each time step. The 128 km x 128 km simulation no longer produced a regular cycle of primary followed by smaller amplitude secondary rain events (Fig. 4.22a). Rather, precipitation oscillations were somewhat irregular and most rain events peaked near 40 mm/day (Fig. 4.22a). As in Fig. 4.5 rain events come in pairs, but with a less regular period and consistent relationship between the first rain event and second rain event amplitude. The vertical structure of the wave in figure 4.22a more closely resembles that of K08’s simulations and observed CCWs (Fig. 4.23). The wave packet speed increased from 6 ms$^{-1}$ to 15 ms$^{-1}$ (Fig. 4.23a). A tri-modal temperature structure with local maxima at the surface, 800 hPa – 600 hPa, and the tropopause is coherent from day -5 to +5, which compares well to K08’s figure 9. Moisture anomalies above the boundary layer descended, peaking near 800 hPa compared to 700 hPa and 400 hPa in the simulations where the wave vertical velocity advected the reference T and q profiles (cf. 4.23 and 4.7). These two anomalous moisture peaks more closely resemble K08’s 2000 km wave simulation (his Fig. 8). Analogous changes to the temperature and moisture anomalies occurred in the Q1 and Q2 structures (Fig. 4.24d,e). However, the oscillations were still not as regular as K08 showed. The vertical structure of the 128 km x 128 km domain run
with shear and constant exchange coefficients is similar to the no shear simulation (Fig. D.1).

The 256 km x 64 km, 512 km x 32 km, and 1024 km x 16 km domains also show less regular oscillations than simulations where the wave vertical velocity advected the reference T and q profiles. Precipitating event pairs consisting of a primary and subsidiary rain event were still present in the 512 km x 32 km and 1024 km x 16 km domains (Figs. 22d, e). The 2048 km x 8 km and 2D 8192 km domains are similar to simulations in Fig. 4.5. The vertical structures of the stretched domains are shown in Appendix E.

Figure 4.22 – Same as Fig. 4.5 except the large-scale vertical velocity acted on the current T and q of the CSRM as opposed to the reference T and q profile. See more details in text.
Figure 4.23 – Same as Fig. 4.7 except the large-scale vertical velocity acted on the current T and q of the CSRM as opposed to the reference T and q profile. See more details in text.

4.6 Discussion

The guiding hypothesis of this chapter was that the coupled convective, large-scale system would be more responsive to organized vs. disorganized convection (Section 4.1). Responsiveness was measured via maximum coupled precipitation value. Organization was controlled via two methods: 1) the addition of shear and 2) changing the domain shape to be increasingly more 2D. The importance of shear at organizing
convection was previously discussed in Section 1.2.4.3 and previous work from Tompkins (2000) discussed how 2D vs. 3D encouraged mesoscale organization.

When organization was increased by the addition of shear in the 3D isotropic domain, precipitation oscillations and wave vertical structure were similar to the no shear simulation (cf. 4.5a, b and 4.7 and 4.8). However, when organization was changed by altering domain shape via doubling the x-dimension and halving the y-dimension, domains produced larger precipitation amplitude up to a certain x-y-axis ratio (i.e. the 512 km x 32 km domain; Fig. 4.6). The similarity between the shear and no shear simulations was initially surprising given their different mean reference profiles (Fig. 4.17). The shear case was drier and cooler than the no shear case, closely resembling the reference profiles of the 1024 km x 16 km domain – a domain which did produce much larger precipitation oscillations than the 3D isotropic no shear case. However, experiments wherein the reference T and q profile were prescribed to be the same across all domain setups showed similar results to the original experiments, which calculated the T and q reference profile from days 20 – 30 (Fig. 4.18). The invariance to the prescribed reference profile was attributed to the ability of the coupled system to maintain the same $T'_v - w$ relationship (Fig. C.8).

The indifference to a sheared wind profile may be due to how the squall lines orient themselves relative to the direction of the prescribed shear and/or due to the smallness of the 3D domain. The 2D snapshot of precipitation in Fig. 4.4b and animations (not shown) show the squall lines are oriented approximately 45° to the shear vector or x-axis, as the shear is parallel to the x-axis. This contrasts to the stretched domains wherein convection is aligned in infinite closely spaced regions in the y-
direction perpendicular to the x-axis (Fig. 4.4c, d). The difference means that in the shear case there is intercloud spacing in both the x- and y-dimension, whereas the stretched domains generally only have inercloud spacing in the x-dimension. The spacing differences means that while both shear and stretched domains alter the form of convection, the flow between convective elements is different in the two setups. Shear perpendicular squall lines would create intercloud spacing preferentially in the x-dimension and possibly give precipitation results similar to the stretched domains. Robe and Emanuel (2001) extensively examined the dependence of convective organization, including squall line orientation, to various shear profiles. They found that “superoptimal” shear created squall lines oriented at an angle to the shear vector. To generate squall lines perpendicular to the shear vector either the magnitude of the shear could be reduced (their Fig. 2c) or the shear profile could be changed to represent a jet of shear (their Fig. 11f). A jet shear profile was tested for simulations that used a 1-day \( u \)-, \( T \)-, and \( q \)-damping timescale and gave a larger precipitation amplitude than either the no shear or uni-directional shear simulations (Fig. C.1). In the future, the jet-shear profile with the weaker 10-day damping timescale will be tested, as the 1-day damping at least glimpses the possible effects of using a jet shear. Other shear profiles, such as a sheared upper-level troposphere (i.e. \( p \geq 500 \text{ mb} \)) should be tested as (Schumacher and Houze 2006) found upper-level shear may impact stratiform rain production.

In terms of the smallness of the domain, approximately only one squall line exists in the 3D domain at a time. Robe and Emanuel (2001) had similar results in their study and stated the finite size of the domain may be artificially limiting the convection. Tompkins and Craig (1998) also found that to successfully simulate a radiative-
convective equilibrium the domain size must be large enough to contain the subsidence region from the convective scale of interest. If the domain is too small, they found the convection would become intermittent with bursts of convection separated by inactive periods. Such description is similar to the precipitation evolution seen in Fig. 4.4. Future studies will use a larger domain size to test the robustness of the results found here.

Of the stretched domains, the 512 km x 32 km had the largest coupled precipitation amplitude with the 1024 km x 16 km having a slightly smaller amplitude (Fig. 4.6). While the increase in coupled precipitation amplitude from the 3D isotropic domain to the 512 km x 32 km domain is in agreement with the hypothesis, we still need to determine the physical mechanism within the coupled framework that leads to such increases in amplitude. Is the change in convective organization influencing non-WISHE surface flux feedbacks?

Beyond the 512 km x 32 km domain, decreases in precipitation amplitude could neither be explained by mean state differences nor sensitivity experiments to a given large-scale vertical velocity forcing (Section 4.4) despite increased mesoscale organization in the 1024 km x 16 km, 2048 km x 8 km, and fully 2D domain (Fig. 4.22). Forcing experiments like Wang and Sobel (2011) where the convective, large-scale system remains coupled and the precipitation response to various SST forcing are tested (Fig. 4.1) may elucidate reasons for the decrease in equilibrium wave amplitude in the 1024 km x 16 km and more 2D domains. Such experiments would reveal if a continuum of increasing precipitation values exists for a given SST when the domain transitions from 3D isotropic to 3D isotropic with shear to fully 2D. That is, do the shear and stretched domains fall in between the blue and black solid lines on Fig. 4.1? Another
possibility is to perform perturbation type experiments as in Tulich and Mapes (2010). These experiments would be similar in spirit to the large-scale forcing experiments in Section 4.4.2, except temperature and moisture perturbations would be done about the time dependent CSRM T or q profile. In that manner, the system remains coupled and we would test responses of each domain setup to perturbation on the coupled system.

The optimum response of the CSRM-wave system in the 512 km x 32 km is an intriguing mystery that warrants further study. This study offers a useful starting point for future examination of the convective and large-scale coupled system. The approach of changing domain shape or adding shear to alter the form of convection offers a unique opportunity to explore the form function relationship of convection and the larger-scale. Also, the results from the stretched domains hint at the possibility of “convective organization feedbacks” (Tobin et al. 2012). As convective organization increases the mean state of the environment becomes drier, as seen in Fig. 4.17b and observations from Tobin et al. (2012). Therefore, a change in the degree of convective organization may have “the potential to affect the atmospheric circulation and its organization over a large range of scales” (Tobin et al. 2012, p. 6902; who sites Slingo and Slingo 1988; Randall et al. 1989; Zurovac-Jevtic et al. 2006; Kang et al. 2008). Lastly, the approach of this study has the potential to guide the usefulness of convective parameterization. Are separate parameterization schemes needed to account for the organization of convection or are traditional convective parameterization approaches (i.e. deterministic, diagnostic ensemble plume approaches) sufficient? Relying on the shear results seen in Figs. 4.5 – 4.8, the coupled system appears indifferent to organization, so would imply the need for a separate parameterization scheme for organized convection unnecessary. However, the
stretched domains do show a systematic increase in precipitation response as convection becomes more organized up to a point. This leaves the door open for the utility of somehow accounting for organized convection in models, whether by a separate parameterization scheme or superparameterization.

In superparameterization the embedded CSRM has a very limited domain, more so than the domain sizes explored here (e.g. 200 km 2D domains in Grabowski 2001). If the mesoscale is important to convective, large-scale interactions then perhaps exploring different 2D domain sizes or 3D configurations in the superparemeterization setup would provide interesting, useful results to compliment the very limited 2D domains currently used inside GCMs. M. Pritchard (personal communication) has explored such embedded CSRM domain variety and found results similar for 3D vs. 2D configurations and various 2D domain sizes. His results again hint at the indifference of mesoscale organization to the convective large-scale system. Extensions to this current work will hopefully provide more conclusive evidence to the role the mesoscale plays in convective, large-scale interaction. Those possible extensions are explored in detail in Chapter 7.
CHAPTER 5: Summary and Conclusions

The form-function relationship of tropical atmospheric convection and the larger-scale was studied with the aim of understanding how organized convection affects the symbiotic relationship of the convective scale and larger-scale. “Form,” as it relates to convection, is the visual shape and configuration of convective systems, where a system can be single-celled or multi-celled convection ranging from kms to 100s or even 1000s of kms. “Function,” in terms of the convection, relates to the interdependence of convection and its environment.

The first half of this dissertation utilized satellite observations from CloudSat to examine cloud modulation by the MJO (Chapter 3). This work was previously published in Riley et al. (2011). The MJO is one example where the large-scale circulation and the convective scale are coupled. By characterizing the evolution of cloud types and specifically examining how deep convective organization changed through various phases of the MJO, Chapter 3 addressed how the form of convection relates to large-scale circulation conditions, where the large-scale circulation conditions were defined by MJO phases. A crude width criterion was used to differentiate organized vs. non-organized deep convective systems. Along with the evolution of various cloud types and total cloud cover across locally and globally defined MJO phases, temperature and moisture evolution were documented across locally defined phases. MJO results were compared to the convectively coupled Kelvin wave to put the results in context of CCWs on different time and space scales. Main conclusions from Chapter 3 include the following:
1. In locally defined phases, the evolution of cloud (referred to as echo object, EO) types across both the MJO and Kelvin wave are similar to the life cycle of individual MCSs (Figs. 3.1d and 3.4). Such similarity gives credence to Mapes et al.’s (2006) stretched building block hypothesis.

2. The form of deep convection – wide vs. narrow, or organized vs. unorganized, deep precipitating systems – is modulated differently by the MJO. Narrow are relatively more prevalent during developing phases of the MJO, while wide are more prevalent during mature phases (Figs. 3.4b – 3.6). Additionally, in terms of fractional changes in cloud cover between suppressed vs. active phases, the wide deep systems are modulated most by the MJO (Fig. 3.4d).

3. For a given amplitude of OLR, the MJO shows larger low-level moisture variations than the Kelvin wave. The Kelvin wave, however, has larger low-level temperature variations than the MJO. Based on other works (Fuchs and Raymond 2007 and 2009; Yasunaga and Mapes 2011), we speculate that the Kelvin waves are controlled more by CIN, while the MJO more so via moisture variations (cf. 3.7 and 3.11).

4. For globally (RMM index, WH04) defined phases, wide deep precipitating systems were again modulated most by the MJO. Wide deep systems, along with anvil, cirrus, cumulus congestus, and alto cumulus, and stratocumulus EO types showed eastward propagation from the Indian Ocean to the central Pacific, while the narrow deep systems only showed propagation from the Indian Ocean to the Maritime Continent (Fig. 3.9).
5. The anvil and stratocumulus EO types lagged the wide deep precipitating systems by approximately one RMM phase, while narrow precipitating systems led the wide by as much (Fig. 3.9).

6. The role of shallow and middle topped convection at pre-moistening the environment ahead of MJO active phases remains unclear from this work. By locally defined phases, the cumulus EO type was relatively more abundant leading up to MJO active phases, despite having more total cloud cover during active phases. However, middle topped cumulus congestus showed little variation ahead of the active phases. Additionally, the stratocumulus EO type signature was opposite for locally vs. globally defined MJO phases. In local phases the prevalence of the stratocumulus type lead the active phases, while in global defined phases they trailed the wide deep precipitating systems.

7. Coherent western hemisphere signals accompany the eastward propagation of the MJO. For the wide deep precipitating, anvil, cirrus, cumulus congestus, and alto cumulus, and stratocumulus EO types negative EO cover over South America and the Atlantic coexists with positive EO cover anomalies over the Indian Ocean and Maritime Continent (Fig. 3.9).

The second half of the dissertation used a CSRM with parameterized large-scale dynamics to isolate how the degree of organization affected the coupled convective, large-scale system. The guiding hypothesis was that the coupled system would be more responsive to organized vs. scatter convection. Simulations were run wherein convective
organization was altered by either the addition of shear to 3D isotropic domains or by changing the domain shape to have an increasingly elongated x-dimension with a shorter y-dimension. The following are the main results from Chapter 4:

1. The addition of shear or change in domain shape does alter the organization of convection within the CSRM. A uni-directional shear profile produced squall lines that moved at an angle to the shear vector. Increasingly elongating the x-dimension and shortening the y-dimension until the domain was fully 2D results in infinite lines (in the periodic sense) of convective blobs that become further spaced apart as the domain is stretched (Fig. 4.4).

2. 3D simulations run with uni-directional shear give a similar precipitation response and wave vertical structure as the no shear case, suggesting the coupled system is indifferent to convection in the form scattered blobs vs. linear structures (Figs. 4.5, 4.7, and 4.8).

3. Altering organization by changing domain shape leads to larger precipitation oscillations and longer wave period in the coupled system up to a certain x-, y-axis ratio, namely the 512 km x 32 km domain (Figs. 4.5 and 4.6).

The above results from Chapter 5 were extensively tested with: 1) a variety of damping choices (i.e. momentum, thermal, or moisture) and damping timescales applied to the large-scale wave equation (Appendix A), 2) three different large-scale wavelengths (Appendix G), 3) two different formulations of the surface latent and sensible heat fluxes (Appendix A), 4) a fixed idealized radiation profile (Fig. 2.3b) and a fixed radiation
profile from a previous no coupled interactive radiation simulation (Appendix F), 5) the
large-scale vertical velocity advecting the current vs. reference T and q profiles (Chapter
4.5), 6) the prescribed background vertical velocity (Fig. 2.3a) acting on the current vs.
reference T and q profile (Appendix F), and 7) two different coupling frequencies
(Appendix F). While precipitation oscillations and wave vertical structure varied
somewhat given the various sensitivity tests, the main conclusions remained robust. The
null hypothesis that differences in noise across domain shapes led to the excitation of
damped possibly identical modes of the coupled system and thus accounted for the
change in wave amplitude across domains was tested in Appendix B and found false.

Two experiments were conducted in attempt to explain why the 512 km x 32 km
domain gave the largest wave amplitude response. The first set of experiments ran each
domain setup with the reference T and q profile set to the no coupled 128 km x 128 km
soundings to test if mean state differences across domain setups caused the different
precipitation oscillations. Recall, convection is coupled to the large-scale by passing the
CSRM domain averaged virtual temperature anomaly to the large-scale wave equation to
update the large-scale vertical velocity. The anomalous virtual temperature anomaly is
the domain averaged virtual temperature minus a reference virtual temperature profile
(Eq. 4.5 – 4.7). By fixing the reference T and q profile to the no coupled 128 km x 128
km sounding, the 3D domain with shear and the stretched domains advected a moister
reference profile than they otherwise would have (Fig. 4.17). In these experiments, the
large-scale vertical velocity acted on the fixed 128 km x 128 km sounding such that the
coupled system was continually influenced by the “foreign” reference profile.

Nevertheless, the precipitation oscillations remained similar to simulations where the T
and q reference profile was calculated as the average sounding from days 20 – 30 (i.e. 10 days before coupling was turned on). Wave oscillations remained the same, as the coupled system maintained the same relationship between virtual temperature anomalies and large-scale vertical velocity (Fig. C.9).

The second set of experiments decoupled convective processes and the large-scale wave by subjecting convection inside the CSRM to short, low-level vertical velocity perturbations or chirps (Fig. 4.19). The motivation was to test the sensitivity of each domain to the same forcing. The guiding hypothesis was that the 512 km x 32 km domain would give the most intense instantaneous responsive with the 1024 km x 16 km, 2048 km x 8 km, and 8192 km 2D domain less so for a given forcing. Results showed that the 3D isotropic domain gave a relatively intense, but short convective heating and drying response that weakened, yet lengthened as the domain was stretched. This set of experiments showed that change in wave amplitude oscillations for the coupled CSRM-wave system cannot be explained by sensitivity differences as the domain shape is changed. Rather, the coupled system conspires to give optimum wave amplitude response in the 512 km x 32 km domain. The behavior of the coupled system is uniquely different than the decoupled system and warrants further investigation.

This study offers a useful starting point for future examination of the convective and large-scale coupled system. The approach of this study has the potential to guide the usefulness of convective parameterization. Are separate parameterization schemes needed to account for the organization of convection or are traditional convective parameterization approaches (i.e. deterministic, diagnostic ensemble plume approaches) sufficient?
### 5.1 Summary of Appendices

Table 5.1 briefly describes the purpose and conclusions of each appendix.

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<td>Appendix G</td>
<td>Sensitivity to 2500 and 10 000 km large-scale wavelength</td>
<td>Conclusions in Ch. 4 remain robust.</td>
</tr>
</tbody>
</table>

Table 5.1 – Summary of each appendix
CHAPTER 6: Future Works

6.1 More lessons from CloudSat

Chapter 3 provided one example of how our EO data set provides new insights into tropical phenomena. Keeping with the theme of the dissertation, preliminary work has examined the tropic wide distribution of wide and narrow (crudely organized and organized, respectively) deep precipitating EO types. Figure 6.1 shows the geographical location of wide (red dots) and narrow (blue dots) deep EO types over the tropical (20°S – 20°N) ocean and lowlands (z < 1 km) through the seasons. Each red or blue dot is centered over the corresponding EO’s thickness-weighted mean latitude and longitude. Specifically, the latitude and longitude of every pixel comprising an EO is included in the calculation of mean latitude and longitude as opposed to the midway point between the start and end of the EO. The thickness weighting ensures each EO is closely centered over the precipitating portion of the cloud, as opposed to a huge anvil with a precipitating core attached to the extremes of the EO.

Well-known seasonal shifts are evident – the ITCZ migrates equatorward in the boreal winter and a secondary ITCZ forms in the east Pacific during boreal spring (Zhang 2001, Masunaga and L’Ecuyer 2010), while monsoonal changes are clear in the Arabian Sea, Bay of Bengal and the Gulf of Guinea. [The converse image, wide EOs (red) on top of narrow EOs (blue), was also plotted to verify the mentioned ubiquity of the wide EOs.] Interestingly, the wide EO dots appear through the narrow dots despite being
outnumbered about 2 to 1. For example, wide EOs over the Arabian Sea during June, July, and August (JJA) occur farther west than the narrow EOs.

Figure 6.1 – Geographical location of wide (red dots) and narrow (blue dot) deep EOs over the tropical (20°S – 20°N) oceans and low-lands (z < 1 km) by (a) June, July, and August (JJA), (b) September, October, and November (SON), (c) December, January, and February (DJF), and (d) March, April, and May (MAM). Each asterisks is centered over the EO’s thickness-weighted mean latitude and longitude.

Figure 6.2a shows the longitudinal distribution of horizontal pixel coverage for each EO type during JJA for every 5° longitude bin over the Arabian Sea and India (50°E – 90°E and 5°N - 20°N). Figure 6.2b shows the same information, but normalized by total horizontal pixel cover in each longitude bin. Total horizontal pixel cover increases
from the Arabian Sea to the Indian sub-continent, as well as the percent contribution from deep EOs. Looking at panel a, the bulk of the east to west increase in horizontal pixel cover is due to an increase in the amount of deep (wide + narrow) EO coverage over India. Panel b indicates there is a shift from generally convectively suppressed conditions over the Arabian Sea to convectively active conditions over India (i.e., the contribution to horizontal coverage from deep convection increases from ~ 10% over the Arabian Sea to ~ 50% over India). The deep convection over the Arabian Sea is almost entirely from the wide deep EOs. In fact, coverage due to narrow EOs is almost non-existent in the three western most bins.

Figure 6.2c shows one pictorial realization of “actual” (CloudSat observed) clouds over the Arabian Sea and India during JJA (similar to Fig. 3.5). The width of each EO has been converted from degrees to kilometers to provide an easy size comparison between EOs. As a reminder, wide deep EOs are plotted in a grey-to-red color scale, while all other EO types are plotted in a grey-to-blue color scale. The pictorial realization nicely shows the transition from suppressed convection over the western Arabian Sea to deeper, more active convection over India. For this realization, there is almost a linear increase in narrow EO top heights from the Arabian Sea eastward (dashed line). However, just as the bar graphs showed, wide deep EOs (red dots) exist over the generally convectively suppressed Arabian Sea.
Figure 6.2 – (a) Total horizontal pixel cover and (b) normalized horizontal pixel cover percentage for each indicated EO type per 5° longitude bin. (c) Random sample of JJA EOs over indicated green box. EOs are centered over their mean longitude. The width of each EO has been converted from degrees to kilometers. (b)

Figure 6.3 (left) shows normalized histograms of precipitable water (PW) for each tropical (20°S – 20°N) narrow (black line) and wide (red line) ocean EO. The right panel is the same except weighted by EO area. PW values are from AMSR-E data that was co-located with the CloudSat track as part of the 2-year YOTC virtual campaign (Chapter 2.2). The total of each normalized histogram equals one. The mode of the wide distribution is slightly lower than that of the narrow (57 mm vs. 59 mm, respectively). The tails of the two distributions are quite intriguing. At low PW values the wide deep
histogram has a fatter, longer tail compared to the narrow histogram, while at high PW values the narrow histogram tail is longer than the wide. The fat tail of the wide EOs at low PW values is consistent with figure 6.2 that wide convective systems seem to occur in marginally moist environments where narrow convection cannot.

The YOTC ECWMF pressure level data has also been examined on a tropics-wide scale for the wide vs. narrow EO types, but meaningful discussion was hard to gain from preliminary figures. Future work will refine figures by perhaps focusing on a specified region in the tropics, like the Arabian Sea as done in figure 6.2. Also, additional YOTC fields will be compared for not only wide vs. narrow EOs, but all EO types (as defined in Chapter 3).
6.2 Further analysis in the CSRM + PSLD framework

This study lays a solid foundation for additional experiments within the coupled convection and large-scale framework. Specifically, refinements to conclusions or questions posed in Chapter 4 are perhaps possible through the following analyses or experimental designs:

- *Analyze the surface precipitation and precipitable water within the convecting region of each domain.* Domain wide averages smear out any potentially interesting relationships between precipitation and PW, SHF, or LHF. For example, figure 6.4 shows domain averaged surface precipitation vs. PW. It is difficult to say whether these domain averaged relationships are interesting given the stretched domains have more cloud free area than the more 3D domains, naturally leading to lower domain averaged PW values. If, however, 2D statistics from each domain simulation are saved and the convecting regions identified and isolated, we could potentially evaluate if different precipitation vs. PW relationships exist given the various degrees of convective organization. Would all the points in figure 7.4 lie on top of each other? Halloway and Neelin (2009, 2010), Neelin et al. (2009), and Yano et al. (2013) extensively looked at the relationship between rain rate and PW. All showed rain rate generally increases as PW values increase. Neelin et al. (2009) further showed a power-law pickup in ensemble average precipitation once PW reached a critical value. While Neelin et al. (2009) did not differentiate between organized vs. unorganized system in their
study, the framework used in this study offers an opportunity to do so. Figure 7.3 from CloudSat and AMSR-E observations already hints at unique relationships between rain rate and PW for organized vs. unorganized systems.

- **Apply multiple k’s to the large-scale wave equation (Eq. 4.6).** This would test the nonlinear interaction of convectively coupled waves of different scales. The large-scale vertical velocity could be computed for both a long and short (e.g. 10 000 km and 2500 km) wavelength. The CSRM temperature and moisture tendency equations (Eqs. 4.7 and 4.8) would then have additional terms from both the long and short wavelength vertical velocity. The single w value in the second terms of Eqs. 4.7 and 4.8 would become $w_{\text{long}} + w_{\text{short}}$.

- **Identify various cloud types within each domain setup.** Analysis would be done in a similar spirit to the CloudSat observation work. That is, cloud types would be identified and classified into various phases of the simulated convectively coupled waves. Cloud type identification could be done similar to Tulich et al. (2007) using precipitating water (e.g. Fig. 4.16) to define cloudy regions and vertical velocity to further separate the cloud types. By do so; we could definitively show how convection is changing for each domain setup. Initial identification of cloud types would admittedly be more cumbersome than the CloudSat EO identification or that done by Tulich et al. (2009) given those studies analyzed 2D statistics and the CSRM simulation output would give 3D cloud objects. However, analysis tools such as VAPOR (used here to make Fig. 4.16) and the Integrated Data
Viewer (IDV) make 3D analysis possible. Alternatively, SAM could be hardcoded to save a 2D slice of information along a specified domain point every specified number of time steps. This method would produce a collection of cloud statistics similar to CloudSat.

- **Redo simulations starting with a larger 3D isotropic domain.** Perhaps using a larger domain size would give similar 3D isotropic results as K08. Figure 6.5 shows the precipitation oscillations for a 512 km x 512 km and 768 km x 768 km domain simulation when a 1-day u-, T-, and q-damping timescale was used. Both the 512 km x 512 km and 768 km x 768 km domain have less high variability noise than the smaller 128 km x 128 km domain. Post-coupled precipitation in the 786 km x 786 km drifts to a lower precipitation rate than the pre-coupled precipitation mean, which may be caused by convective self-aggregation (e.g. Tompkins 2001; Bretherton et al. 2005; Muller and Held 2012). Given a large enough domain, convection within CSRM simulations has been shown to self-aggregate wherein the equilibrium state consists of an isolated moist convectively active portion of the domain while the remaining area is dry and convectively suppressed (e.g. Bretherton et al. 2005; Muller and Held 2012). Convective self-aggregation, though, partially relies on radiative feedbacks between cloudy vs. clear regions, so may not be a suitable explanation for the precipitation drift in the 786 km x 786 km simulations that used a fixed idealized radiation profile.

While the larger 3D simulations took considerably longer to run, re-doing the larger 3D domains with the 10-day damping timescale is certainly doable and
worth of investigation. Additionally, it could be interesting to see if self-aggregation still occurs, through simple moisture convective feedbacks, when radiation is fixed throughout the domain.

- **Redo simulations with the radiation profile fixed at -1.5 K/day instead of 1.3 K/day.** (Need to run 128 km x 128 km domain with -1.5 K/day. With 4000 km large-scale wavelength, a 128 km x 128 km domain with 4 km spacing, 5-day damping timescale, and -1.5 K/day produced oscillations that did not consist of primary and secondary rain event pairs. Though, oscillations were still not as regular as K08.)

- **Apply T and q perturbations to the coupled convective and large-scale framework.** As mentioned in Chapter 4.6, these experiments would be similar to Tulich and Mapes (2010) and would probe the sensitivity of the different domain shapes to various T and q perturbations. Such experiments would also test the validity of Kuang’s (2012) linear response function results. He showed different sensitivities for a given T or q perturbation when the convection was organized (in a 2048 km x 64 km domain) vs. unorganized (in a 128 km x 128 km domain; his Fig. 8).

- **Compute the linear response functions for each domain setup.** This is the corollary to the above bullet and would further verify the utility of the linear response functions (Kuang 2012).
• *Do SST forcing experiments.* As mentioned in Chapter 4.6, these experiments would be similar to Wang and Sobel (2011), except the wave equation test harness (K08) would be used instead of the WTG approximation. The system would remain coupled and the response to various external SST forcings could be measured.

• *Objectively define organized vs. non-organized convection.* In Chapter 4, organized and non-organized convection was subjectively defined by comparing the form of convection in the 3D isotropic domain to the stretched domains and 2D domain. Convection was more scattered in the 3D isotropic domain compared to the stretched domains and was therefore deemed non-organized. Tobin et al. (2012) used three objective measurements to classify the aggregation or organization of convection: 1) Number of convective clusters in a specified area, 2) The clumping of the clusters, which basically measures the distance between nearest neighbor convective cells, and 3) A simple convective aggregation index (SCAI), which combines the information from (1) and (2). These three measurements could easily be applied to our model output to objectively define organized vs. non-organized convection.

6.3 Beyond the CSRM + PLSD framework

• *Redo the chirp simulations but with a longer forcing period and/or different shape.* This work only tested the stratiform instability mechanism by prescribing
a low-level forcing chirp. However, some domains may be more unstable to a deep vertical velocity forcing.

- **Do large 2.5D experiments without PLSD.** In these experiments the x-dimension would be large enough to contain a large-scale overturning circulation. The spontaneous organization of convection could be measured in a similar manner to Tulich and Mapes (2008) or Posslet et al. (2012).

- **Use the superparameterization framework to test and further results found in this study.** In the superparameterization framework (Grabowski 2001, 2003; Randall et al. 2003) both the large-scale and convective scale are explicitly resolved. This allows convection to interact with the full spectrum of large-scale wavelengths, as opposed to one wavelength in the wave equation test harness approach. Studying the consequences of different embedded CSRM domain sizes and dimensionality would inform future studies on best practices of embedded CSRM domain
APPENDIX A: Damping Sensitivity

The purpose of this appendix is twofold: 1) to confirm that conclusions made in Chapter 5 are robust to various damping choices (i.e. either use of a combination of mechanical, thermal, and/or moisture damping, or damping timescale choice) and 2) to catalog various simulations run for this work.

Simulations were done calculating the surface latent heat flux (LHF) and sensible heat flux (SHF) two ways with regard to the exchange coefficients in the bulk aerodynamic formulas (Eqs. A.1 and A.2): 1) where $C_E$ and $C_H$ changed depending on the stability of the surface layer and 2) where $C_E$ and $C_H$ were fixed at $1.1 \times 10^{-3}$. In both methods, the wind speed is fixed at 5 m s$^{-1}$ to prevent wind induced evaporation feedbacks (i.e. WISHE, discussed in Chapter 1.2.2.2). Motivation for using varying vs. fixed exchange coefficients was fortuitous, as the original SAM code obtained for this work had the exchange coefficients dependent on surface stability, while a later version of the code had them fixed. Since the use of varying vs. fixed exchange coefficients impacts the damping choice, results from the two different sets of simulations are discussed in this Appendix.

\[
LHF = C_E |V| \left( q_{sat}(T_s) - q_{air} \right) \quad (A.1)
\]
\[
SHF = C_H |V| \left( T_{suf} - T_{air} \right) \quad (A.2)
\]
Section A.1 examines the sensitivity of the various domain setups to different damping combinations (i.e. the use of $\varepsilon$, $\varepsilon_T$ or $\varepsilon_q$), while section A.2 examines the sensitivity to damping timescale choice. Section A.3 briefly discusses the implication of the various damping choices.

A.1 Using $\varepsilon$, $\varepsilon_T$ and $\varepsilon_q$ damping combinations

A.1.1 Using the surface flux formula with varying exchange coefficient

In Chapter 5, $\varepsilon$, $\varepsilon_T$ and $\varepsilon_q$ damping were applied in Eqs. 4.6 – 4.8. This appendix sub-section examines the sensitivity of the results given only $\varepsilon$, $\varepsilon_T$ or $\varepsilon_q$ was prescribed or combinations of $\varepsilon$, $\varepsilon_T$ and $\varepsilon_q$ were prescribed. Figures A.1 and A.2 show the mean surface precipitation response given the panel indicated 10-day and 4-day damping combinations, respectively, for the 128 km x 128 km domain.

For a 10-day damping all combinations of $\varepsilon$, $\varepsilon_T$ and/or $\varepsilon_q$ result in regular precipitation oscillations characterized by precipitating event pairs; relatively large-amplitude precipitation events are followed by smaller secondary precipitation events (similar to results discussed in Chapter 5.3). The amplitude of the primary rain events varies depending on damping combination choice. A thermal only damping (Fig. A.1b) resulted in precipitation oscillations that continued to grow with time (at least over the simulation time period). Using $\varepsilon$ or $\varepsilon_q$ damping only, or a combination of $\varepsilon$ and $\varepsilon_T$ damping or $\varepsilon_q$ and $\varepsilon_T$ damping resulted in precipitation oscillations peaked near 80 mm/day (Fig. A.1a, c, d, and f). The combination of $\varepsilon_q$ and $\varepsilon_q$ or all three dampings kept
precipitation oscillations near 40 mm/day (Fig. A.1e, g). Wave period changed slightly too depending on damping combination choice, but is less noticeable than changes in amplitude. For example, the $\varepsilon_T$ only damping and $\varepsilon$ and $\varepsilon_T$ damping combination have a slightly longer wave period than the other simulations (cf. A.1b,d to other A.1 panels).

Figure A.1 – Domain averaged precipitation as a function of time for the 128 km x 128 km domain with 10-day damping timescale for (a) momentum, (b) temperature, (c) moisture, (d) momentum and temperature, (e) momentum and moisture, and (f) momentum, temperature, and moisture. Days 0 – 30 is the pre-coupled time period. The dashed line at 40 mm/day is repeated in each panel for ease of comparison.

Applying a thermal 4-day damping, Fig. A2.b, resulted in increasingly larger precipitation events with time, as the third peak reaches 120 mm/day, while the fourth reaches 140 mm/day (cutoff in panel b). For all other 4-day damping combinations, the precipitation events remained bounded under ~40 mm/day, though the nature of the
precipitation events varied. For example, precipitating event pairs show larger differences between primary and secondary precipitation events when only $\varepsilon$ or $\varepsilon_T$ or a combination of $\varepsilon$ and $\varepsilon_T$ damping was applied compared to other damping combinations (cf. A.2a and A.2c). Precipitation oscillations peak near 25 mm/day when $\varepsilon_q$ was used in the damping combinations (Fig. A.2c, e – g).

Figure A.2 – Same as Fig. A.1, except for a 4-day damping timescale is used.

Figure A.3 shows various 4-day damping combinations for the 512 km x 32 km and 1024 k km x 16 km domains. Similar to the 128 km x 128 km domain, variations in precipitation oscillation amplitude are the most noticeable difference across damping combinations. As in Fig. 5.5, comparison between respective damping combinations
across the 128 km x 128 km, 512 km x 32, and 1024 km x 16 km domains shows the stretched domains have a larger precipitation response than the 3D isotropic domain.

![Graph](image)

**Figure A.3** – Indicated 4-day damping combinations for the (a) 512 km x 32 km and (b) 1024 km x 16 km domain.

**A.1.2 Using surface flux formula with fixed exchange coefficient**

Figure A.4 is analogous to figure A.1, except simulations were done using \( C_E \) and \( C_H \) fixed to \( 1.1 \times 10^{-3} \) in the bulk aerodynamic formulas (Eqns. A.1 and A.2). Unlike simulations done with \( C_E \) and \( C_H \) dependent on the stability of the surface layer (i.e. Fig. A.1), all damping combinations produced precipitation oscillations bounded under 40 mm/day, except for the thermal only damping, which peaks near 90 mm/day (Fig. A.4). In terms of wave period and vertical structure, simulations run across the various domain
setups with fixed exchange coefficients are similar to those done with surface stability dependent exchange coefficients (cf. 5.5 and A.5 and 5.7 – 5.13 and A.6 – A.12).

Comparing the evolution of LHF in simulations done with varying vs. constant exchange coefficients reveals two differences: 1) LHF varies less when $C_E$ is fixed (i.e. $\sim 5$ Wm$^{-2}$ vs. $\sim 15$ Wm$^{-2}$, Fig. A.13), and 2) The lag time between peak rainfall and peak LHF anomalies approximately doubles when $C_E$ is fixed (Fig. A.14). These two effects render thermal and moisture damping unnecessary to keep oscillations at reasonable values. While there is a slight shift ($\sim 3$ hrs) in lag time between peak rainfall and peak SHF, this change was insignificant as a simulation done holding $C_E$ constant, while $C_H$ varied depending on surface stability, gave precipitation oscillations similar to Fig. 5.17a.

Figure A.4 – Same as A.1, except the bulk aerodynamic exchange coefficients are fixed at $1.1 \times 10^3$ ms$^{-1}$ for each simulation.
Figure A.5 – Same as Fig. 5.5, except the bulk aerodynamic exchange coefficients are fixed at $1.1 \times 10^{-3}$ m$s^{-1}$ for each simulation.
Figure A.6 – Same as Fig. 5.7, except the bulk aerodynamic exchange coefficients are fixed at $1.1 \times 10^{-3}$ ms$^{-1}$ for each simulation.
Figure A.7 – Same as Fig. 5.8, except the bulk aerodynamic exchange coefficients are fixed at $1.1 \times 10^{-3}$ ms$^{-1}$ for each simulation.
Figure A.8 – Same as Fig. 5.9, except the bulk aerodynamic exchange coefficients are fixed at $1.1 \times 10^{-3}$ ms$^{-1}$ for each simulation.
Figure A.9 – Same as Fig. 5.10, except the bulk aerodynamic exchange coefficients are fixed at $1.1 \times 10^3$ ms$^{-1}$ for each simulation.
Figure A.10 – Same as Fig. 5.11, except the bulk aerodynamic exchange coefficients are fixed at $1.1 \times 10^{-3}$ m s$^{-1}$ for each simulation.
Figure A.11 – Same as Fig. 5.12, except the bulk aerodynamic exchange coefficients are fixed at $1.1 \times 10^{-3}$ ms$^{-1}$ for each simulation.
Figure A.12 – Same as Fig. 5.12, except the bulk aerodynamic exchange coefficients are fixed at $1.1 \times 10^{-3}$ ms$^{-1}$ for each simulation.

Figure A.13 – Domain averaged surface latent heat flux as a function of time for a 128 km x 128 km domain simulation with the surface exchange coefficients dependent on the surface stability (black line) and surface exchange coefficients fixed at $1.1 \times 10^{-3}$. 
Figure A.14 – Domain setup vs. lag in latent heat flux (LHF) peak after peak rainfall for simulations where the exchange coefficient in the latent heat flux formula is dependent on stability (black line) and fixed at $1.1 \times 10^{-3}$. The corresponding y-dimension for each domain, listed from left to right is: 128 km, 128 km, 64 km, 32 km, 16 km, 8 km, and fully 2D.

A.2 Damping timescale choice

A.2.1 Using surface flux formula with varying exchange coefficients

Several $\varepsilon$, $\varepsilon_T$, and $\varepsilon_q$ dampings were tested, ranging from 1-day to 10-days, with 1-day being the strongest damping and 10-day the weakest, for the various domain setups discussed in Chapter 5. Figure A.15, A.16, and A.17 are synonymous with figure 5.5 except a damping timescale of 1-, 2-, and 4-days were used, respectively. The motivation for examining different damping timescales was to find a timescale that made comparisons between the various domain setups cleanest or most interpretable. Figure A.15 shows a 1-day damping timescale for the 128 km x 128 km shear and no shear domain produced precipitation oscillations about the pre-coupled mean precipitation – similar to K08’s results. However, the 1-day damping made comparisons between
domain setups difficult as the stretched domains entered exotic states, which made it
difficult to interpret model behavior (discussed more below). Generally, as the damping
weakened (i.e. damping timescale increased), precipitation amplitude increased for each
respective domain setup (cf. A.15 – A.17). The exception was the 8192 km 2D domain,
in which the precipitation oscillations changed little with weaker damping. Since
precipitation values cannot go negative, the difference is largely seen in the increase in
max surface precipitation (Fig. A.18).

Figure A.15 – Same as Fig 5.5 except a 1-day damping timescale is used. Note the y-axis range only
goes from 0 – 40 mm/day.
Figure A.16 – Same as Fig. 5.5, except at 2-day damping timescale is used.
Figure A.17 – Same as Fig. 5.5, except a 4-day damping timescale is used.

Figure A.18 – Damping time vs. maximum post-coupled precipitation for the indicated domain setup.
The transition between precipitation troughs and crests differs from the 1-day damping to the 10-day damping timescale. With a 1-day and 2-day damping timescale, the growth and decay of precipitation is rather symmetric about peak rainfall events in the 128 km x 128 km domain, both with and without shear, and in the 256 km x 64 km domain (Fig. A.15a – c). These wave oscillations are most similar to the results of K08 (his Fig. 2c). As the damping weakened, the precipitation became increasingly non-symmetric around peak rainfall with slower buildup to maximum precipitation vs. decay, especially for the stretched domains, and hiccups in precipitation became noticeable (panels d – f in A.15 – A.17).

The contrast between the 128 km x 128 km domain run with and without shear differed depending on damping timescale. With a 1-day damping timescale, there is a noticeable difference between the shear and no shear case. The shear simulations have larger precipitation oscillations than the no shear simulation (Fig. A.15b, c and A.18). Figure A.19 shows the differences more clearly. When the damping was weakened to 2-, 4-, and 10-days the no shear and shear simulations show similar, in terms of amplitude and period, post-coupled precipitation oscillations (cf. A.15 – A.17, and 5.5).

![Figure A.19 – Domain-averaged surface precipitation as a function of time for simulations done in the 128 km x 128 km domain with indicated shear profiles. See text for details on shear profiles.](image)
For a 1-, 2-, and 4-day damping timescale, precipitation oscillations in the 256 km x 64 km domain have similar amplitude and period as the 128 km x 128 km domain (Fig. A.15d – A.17d and Fig. A.4). With a 1-day damping timescale, the 1024 km x 16 km and 2048 km x 8 km (Fig. A.15e, f) precipitation response entered exotic states. The 512 km x 32 km domain shows irregular precipitation oscillations (Fig. A.15d). Two other simulations were done for the 512 km x 32 km domain, each initialized with different random noise, to test the robustness of the irregular oscillations. Figure A.20 shows there is no consistency between the time period of the “heavy” rain events and the extended lower precipitating periods. For example, the red simulation shows two heavy precipitating events, centered on day 37 and day 75, with low precipitation in between, whereas the blue simulation has about four heavy precipitating events, with shorter low precipitating periods in between compared to the red simulation (Fig A.20).

The most responsive, in terms of max precipitation, domain is dependent on the damping timescale (Fig. A.18). With a 1-day damping timescale, the 2048 km x 8 km domain produced the largest precipitation response. However, as just noted, the stretched domains with a 1-day damping timescale are hard to interpret. With a 2-day damping timescale, both the 1024 km x 16 km and 2048 km x 8 km domains have the largest precipitation response, while the 1024 km x 16 km and 512 km x 32 km domains produced the largest rainfall response for a 4-day and 10-day damping, respectively (Fig. A.18).
A.2.2 Using surface flux formula with fixed exchange coefficients

Figures A.21 – A.23 show the precipitation response for each domain setup when a 2-, 4-, and 10-day momentum damping timescale was used, respectively, and the bulk aerodynamic formulas had exchange coefficients ($C_E$ and $C_H$ in Eqns. 5.8 and 5.9) fixed at $1.1 \times 10^{-3}$. General characteristics of the precipitation oscillations using a 2-, 4-, and 10-day momentum damping are very similar to results shown in Section 5.3. Again, the 512 km x 32 km and 1024 km x 16 km domains, respectively, have the largest precipitation response (Fig. A.24).

In general, simulations run with constant vs. varying exchange coefficients in the flux formulas resulted in larger precipitation oscillations across each respective domain setup (cf. A.18 and A.24). As in the simulations run with varying exchange coefficients, for a 2-, 4-, and 10-day u-damping there is little difference between the 128 km x 128 km domain run with and without shear (cf. A.18 and A.24). The use of constant exchange coefficients with a 2-day u-damping altered the behavior of the precipitation oscillations.
in the 1024 km x 16 km, and 2048 km x 8 km domains. These two stretched domains no longer enter extended dry periods, interrupted irregularly by precipitation (cf. A.2e, f and A.7e, f). Rather, precipitation oscillations are more regular, behaving similar to the 3D isotropic domain, except for different precipitation amplitude and wave period (cf. A.7a and A.7e, f). Perhaps the extended lag between peak rainfall and peak LHF, along with more steady LHF values, prevents the 1024 km x 16 km and 2048 km x 8 km domains from entering extended dry periods.

Figure A.11 – Same as Fig. 5.5, except for a 2-day momentum damping is used and the bulk surface exchange coefficients are fixed at $1.1 \times 10^{-3}$. 
Figure A.22 – Same as Fig. A.22, except for a 4-day momentum damping.
Figure A.23 – Same as Fig. A.22, except for a 10-day momentum damping.
A.3 Discussion

Irrespective of using constant vs. varying exchange coefficients in the bulk aerodynamic formulas or the use of different damping combinations the following statements can be made:

1. All damping combinations, except for a thermal only damping, result in bounded precipitation oscillations

2. Due to (1), any damping combination, except for thermal only, produce the same conclusions stated in Chapter 5. Namely, precipitation oscillations increase as domains are stretched to a certain xy-aspect optimum, after which precipitation oscillation amplitude decreases.
3. The conclusion in Chapter 5 that convective organization via shear has little impact on the coupled convective, large-scale system remains valid for all damping timescales tested except the 1-day timescale.

4. Altering organization via change in domain shape does alter the behavior of the convective, large-scale system, in terms of max precipitation response, wave packet speed, and symmetry of precipitation around peak rainfall, independent of damping timescale choice.

5. The most responsive domain is either the 512 km x 32 km or 1024 km x 16 km domain when a 4-day or 10-day damping is applied (Figs. A.4 and A.9). Section 5.4 further discusses possible explanations for the change in precipitation response when domain shape is changed.
APPENDIX B: Holding y-dimension constant at 128 km

This appendix examines the null hypothesis that organization has no effect on the coupled convective, large-scale system, rather changes in noise, as the domain shapes are changed, lead to different precipitation responses. In geophysical fluid dynamic systems, noise is sometimes enough to excite the least damped modes of the system. For example, Jin et al. (2001) found that including mid-latitude weather variability (i.e. stochastic noise) as a forcing in their simple global coupled atmosphere-ocean climate model was able to excite the two least damped modes – about 16 year and 10 year variability – in the tropical Pacific basin. While a complete eigenmode analysis is not done here, a set of simulations wherein the noise remains similar across domain shapes is done to test the null hypothesis.

Table 1 in Chapter 5 showed that the standard deviation of pre-coupled precipitation increased as the domains were stretched from 128 km x 128 km to 2048 km x 8 km. Repeating the set of simulations, but holding the y-dimension fixed at 128 km while doubling the x-dimension successfully reduced the standard deviation of the pre-coupled precipitation in the stretched domains to values similar to the 3D isotropic domain (Table B.1). Figures B.1 and B.2 show that despite the reduction in noise when the y-dimension is held fixed, the post-coupled precipitation oscillations remain similar to simulations where the y-dimension was increasingly halved as the x-dimension was doubled. Therefore, the increase in noise as the domains are stretched does not explain the differences in post-coupled precipitation oscillations. The conclusion holds for
simulations done using varying or fixed exchange coefficients in the bulk aerodynamic formula, Figs. B.1 and B.2, respectively.

<table>
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<th>$C_E$ and $C_H$ fixed</th>
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</tbody>
</table>

Table B.1 – Standard deviation of pre-coupled precipitation for indicated simulation setup, where the leftmost column indicates x-dimension length. For the simulations where the y-dimension is halved, the respective y-dimensions going down the column are 128 km, 64 km, 32 km, 16 km, and 8 km. Units are ms$^{-1}$. 

Figure B.1– Domain-averaged surface precipitation as a function of time for indicated domain size. These simulations were run with a 1-day u, T, and q-damping timescale and the bulk exchange coefficients dependent on the surface stability.
Figure B.2 – Domain-averaged surface precipitation as a function of time for indicated domain size. These simulations were run with a 10-day $u$, $T$, and $q$-damping timescale and the bulk exchange coefficients dependent on the surface stability.
APPENDIX C: T and q mean state sensitivity

This appendix further examines the sensitivity of the coupled CSRM-wave system to a specified $T_{v,ref}$ profile, as was briefly done in Chapter 5.4.1. The motivation for running these sensitivity simulations was to see if the coupled system could be tweaked in such a way as to prevent the elongated domains from entering irregular oscillations with extended dry periods. Figure C.1 shows the precipitation evolution of each domain setup when the 128 km x 128 km T and q uncoupled profiles are used to compute $T_{v,ref}$ in Eqn. 4.6. This specified reference profile is referred to as $T_{v,128}$. These simulations, and those shown in C.2 – C.4, used a 1-day u-, T-, and q-damping timescale and LHF and SHF were computed with exchange coefficients dependent on surface stability.

Figure C.1 – Three hour domain averaged surface precipitation with time for the indicated domain size. For each simulation, the 128 km x 128 km $T_{v,ref}$ was used in Eq. 4.6.

The precipitation evolution in the 128 km x 128 km shear setups and the 256 km x 64 km domains using $T_{v,128}$ has no discernable difference from simulations that establish their own $T_{v,ref}$ profile (cf. C.1 and A.1b, c). As domains are stretched beyond 256 km x
64 km, precipitation evolution, when $T_{v,128}$ is used, no longer shows irregular oscillations or enters extended dry periods. Rather, the domains have regular precipitation oscillations that progressively increase in equilibrium amplitude, as well as, oscillate around an increasingly higher mean post-coupled rainfall value. The exception is the 2048 km x 8 km domain, which has lower amplitude oscillations than the 1024 km x 16 km domain.

The increase in oscillation amplitude as domains are stretched is not related to the increase in post-coupled mean rain rate. A set of experiments was done in the 128 km x 128 km domain wherein the background vertical velocity profile (Fig. 2.3a) was doubled and quintupled. The increased background vertical velocity increases both the pre- and post-coupled precipitation mean, while post-coupled precipitation oscillations maintain similar amplitudes (Fig. C.2). Also, if oscillation about a higher mean rain rate was correlated with higher amplitude oscillations, than the 2048 km x 8 km domain would have produced similar amplitude oscillations as the 1024 km x 16 km domain, as both domains have similar post-coupled mean rain rate. I speculate that the increase in amplitude is due to a change in the relationship between convection and the large-scale wave caused by a change in convective organization as domain shape is changed. To completely validate the above statement, simulations using the 512 km x 32 km, 1024 km x 16, and 2048 km x 8 km domains will be run with doubled and quintupled background vertical velocity profiles to ensure post-coupled precipitation amplitude in those domains are not related to mean rain rate.
Figure C.2—Three hour domain-averaged surface precipitation with time for the 128 km x 128 km domain run with a 1-day u-, T-, and q-damping timescale. LHF and SHF were computed with CH and CE dependent on surface stability. The background vertical velocity profile (Fig. 2.3a) was increased by two and five in the red and blue simulations, respectively.

For completeness figure C.3 shows that when the converse experiment is done – i.e. giving the 128 km x 128 km domain the $T_v^{ref}$ profile from other domain setups – a synonymous change in precipitation oscillations occurs. There is little to no change to precipitation evolution when the 256 km x 64 km $T_v^{ref}$ reference profile is used in the 128 km x 128 km domain, but the oscillations become smaller, if oscillating at all, and centered on a lower mean rainfall rate when the $T_v^{ref}$ profile is from the 512 km x 32 km or further stretched domains.

Figure C.3—Three hour domain-averaged surface precipitation for the 128 km x 128 km domain run with $T_v^{ref}$ fixed to the indicated uncoupled sounding average.
Figure C.4 shows that specification of the q profile, as opposed to the T profile, in the calculation of $T_{v\text{ref}}$ is the key to kicking the stretched domains into a state where precipitation oscillations occur. Figures C.5 – C.7 show the results of extensive sensitivity testing of the different domain setups to specified artificially moist or dry q-ref profiles to calculate $T_{v\text{ref}}$. For the 128 km x 128 km domain simulations were tested using a 1-, 4-, and 10-day damping timescale (Fig. C.5). Only a 1- and 10-day damping was tested for the 1024 km x 16 km and 2048 km x 8 km domains (Figs. C.6 and C.7). Given a sufficiently moister or drier q-profile to calculate $T_{v\text{ref}}$, the 1-day damping simulations produces precipitation oscillations around a higher or lower, respectively, rain rate than that of the pre-coupled precipitation mean (panels a Figs. C.5 – C.7). A “sufficiently moister q-profile” refers to the fact that a 5% vs. 1% moister q reference profile is sometimes necessary to shift the coupled system to a new state. For the 10-day damping simulations, a moister or drier q-profile does not shift the precipitation to a new mean state. Rather, oscillations remain unchanged given a 1% moister or drier q-profile to calculate $T_{v\text{ref}}$. With a 5% moister q-profile, the amplitude of the precipitation oscillations does increase (Figs. C.5c, C.6b).

The sensitivity simulations show that the coupled system responds in one of two ways when given a specified $T_{v\text{ref}}$ profile: 1) Precipitation oscillations remain unchanged as $T_v$ and $w$ adjust to the specified $T_{v\text{ref}}$ profile to maintain the same relationship as when the model finds its own $T_{v\text{ref}}$ profile (Fig. C.8), or 2) The coupled system is kicked to a new mean state wherein post-coupled precipitation oscillate around a different mean rain rate than the pre-coupled mean. The later case made it possible to compare precipitation oscillation amplitude amongst the various domains when a 1-day damping
timescale was used. The conclusion remains robust that effectively altering the organization via changing domain shape leads to an increased precipitation response up to a certain $xy$ axis ratio, after which, precipitation oscillation amplitude decreased (Fig. C.1).

Figure C.4 – Three hour domain averaged surface precipitation with time for the (a) 1024 km x 16 km domain run with the indicated 128 km x 128 km uncoupled $T$ and $q$ average profile used to calculate $T_{v, ref}$. (b) Same as (a) except for the 512 km x 32 km domain.
Figure C.5 – Three hour domain averaged surface precipitation with time for the 128 km x 128 km domain run with (a) 1-day u-, T, and q-damping, (b) 4-day u-, T, and q-damping, and (c) 10-day u-, T, and q-damping. In each panel the 128 km x 128 km uncoupled average T-profile was used to calculate $T_{v\text{ref}}$. Different colors indicate by how much the 128 km x 128 km uncoupled average q-profile was made artificially moist or dry in the calculation of $T_{v\text{ref}}$. The control (black) simulations did not alter the uncoupled average q-profile.
Figure C.6 - Three hour domain averaged surface precipitation with time for the 1024 km x 16 km domain run with (a) 1-day u-, T, and q-damping, and (b) 10-day u-, T, and q-damping. In each panel the 1024 km x 16 km uncoupled average T-profile was used to calculate $T_{v,ref}$. Different colors indicate by how much the 1024 km x 16 km uncoupled average q-profile was made artificially moist or dry in the calculation of $T_{v,ref}$. The control (black) simulations did not alter the uncoupled average q-profile.
Figure C.7 – Same as Fig. C.6 except for the 2048 km x 8 km T and q uncoupled average profiles were used in the 2048 km x 8 km domain to calculate $T_{v\text{ref}}$. 
Figure C.8 – (a and c) $T'_v$ for the 2048 km x 8 km domain when the 2048 km x 8 km uncoupled average q profile was used as is (a) and made 5% moister (c) in the calculation of $T_v^{ref}$. (b and d) The large-scale wave vertical velocity, $w'$, for the 2048 km x 8 km domain when the 2048 km x 8 km uncoupled average q profile was used as is (b) and made 5% moister (d) in the calculation of $T_v$. Note, above 15 km in panels (a) and (c) the color scale is folded over, such that red within blue actually indicated negative values lower than -2 K.
APPENDIX D: Forcing experiment composites

Section 5.4.2 discussed the precipitation response to the various domain setups when given the same large-scale vertical velocity forcing. In that section only composites for the vertical structure of Q1 were shown. This appendix shows the temperature, moisture, and Q2 composites for each domain setup.

All domains are anomalously cold and moist from the surface to ~600 hPa from hour zero to half an hour with a cold, dry boundary layer following peak precipitation for the remaining 6-hours shown. As the domains are stretched the boundary layer becomes increasingly dry, while positive moisture anomalies above strengthen (Figs. D.1b – D.6b). Meanwhile, cool low-level temperature anomalies from 0.0 hrs – 6.0 hrs strengthen and deepen as the domains become increasingly 2D (Figs. D.1a – D.6a). Opposite signed temperature and moisture anomalies occur about 600 hPa from hours 0.0 – 6.0 and weaken as the domains are stretched (Figs. D.1a,b – D.6a,b). Panel c and d in each figure show the composite Q1 and Q2 profiles. Drying anomalies are more bottom heavy than heating anomalies, but show a similar transition to weaker, yet longer lasting positive anomalies as the domains become increasingly 2D. Along with the Q2 maximum around 800 hPa, there is also a secondary maximum in the boundary layer that the Q1 composites did not show.
Figure D.1 – 30 minute composite anomalies for the 128 km x 128 km domain of (a) temperature, (b) moisture, (c) convective heating (Q1), and (d) convective drying (Q2). Anomalies are the difference from the -6.0 hrs to 6.0 hrs mean. Units are given in parentheses above each panel and panels (a) and (b) additionally give the contour values.

Figure D.2 – Same as Fig. D.1, except for with uni-directional shear (described in Chapter 5.2)
Figure D.3 – Same as Fig. D.1, except for the 256 km x 64 km domain.

Figure D.4 – Same as Fig. D.1, except for the 512 km x 32 km domain.
Figure D.5 – Same as Fig. D.1, except for the 1024 km x 16 km domain.

Figure D.6 – Same as Fig. D.1, except for the 2048 km x 8 km domain.
Figure D.7 – Same as Fig. D.1, except for the 8192 km 2D domain.
APPENDIX E: Remaining “$w$ advects current $T$ and $q$ profile” simulation composites

In Chapter 4.5 precipitation response of simulations run with the large-scale vertical velocity, $w$ (Eqn. 5.5), advecting the CRSM’s current $T$ and $q$ profiles. Only the 128 km x 128 km vertical structure regressions were shown (Fig. 4.20). This appendix documents the remaining vertical structure composites from the simulations where $w$ advected the instantaneous domain averaged $T$ and $q$ profiles in Eqs. 4.7 and 4.8.

Figure E.1 shows the vertical structure of the 128 km x 128 km domain with shear simulation. As stated in Chapter 4.5 the shear and no shear cases are similar. Figures E.2 – E.6 show the vertical structures of the stretched domains. The 256 km x 64 km has a similar vertical structure to the 128 km x 128 km domain (Fig. E.2). The 512 km x 32 km and 1024 km x 16 km temperature structures look more similar to the simulations wherein the large-scale wave vertical velocity advected the reference $T$ and $q$ profiles than the temperature structure in the more 3D domains (Figs. 5.20 and E.2) or in K08’s results. There is little difference in the 2048 km x 8 km domain and 8192 km 2D domain vertical structures compared to their original simulation counterparts (cf. 5.12 and 5.13 to E.5 and E.6).
Figure E.1 – Same as Fig. 4.7, except for the 128 km x 128 km domain with shear.
Figure E.2 – Same as Fig. 4.7, except for the 256 km x 64 km domain.
Figure E.3 – Same as Fig. 4.7, except for the 512 km x 32 km domain.
Figure E.4 – Same as Fig. 4.7, except for the 1024 km x 16 km domain.
Figure E.5 – Same as Fig. 4.7, except for the 2048 km x 8 km domain.
Figure E.6 – Same as Fig. 4.7, except for the 8102 km 2D domain.
APPENDIX F: Other sensitivity tests

As described in Chapter 2.3, each simulation in Chapter 5 used the same fixed idealized radiation profile and background vertical velocity profile (Fig. 2.3a). The prescribed background vertical velocity profile acted only on the reference T and q profile of each domain setup and hence was a constant background forcing on the CSRM. This appendix examines the sensitivity of the 128 km x 128 km simulation results with a 5000 km large-scale wavelength to: (1) prescribing a fixed radiation profile derived from a prior no coupled interactive radiation run (as was done in K08), (2) having the background vertical velocity advect the current T and q profile, and (3) changing the coupling frequency between the CSRM convection and the large-scale wave.

Figure F.1 shows the time evolution of precipitation for both the 128 km x 128 km domain run with a fixed idealized vs. fixed interactive radiation derived profile. Both simulations are run with bulk aerodynamic coefficients fixed at $1.1 \times 10^{-3}$, a 10-day momentum damping, the large-scale wave vertical velocity advecting the current T and q profiles, and the background vertical velocity advecting the reference T and q profiles. While the simulations are not exactly the same, they have the same general character – similar wave period and amplitude. A fixed idealized vs. fixed interactive radiation derived profile was also compared in a 192 km x 192 km domain. Again, the characteristics of the precipitation oscillations were similar (not shown).

The use of an idealized radiation profile for each domain saved time when doing the no coupled runs to establish an equilibrium T and q profile. For example, a 100-day no coupled interactive radiation simulation done for the 1024 km x 8 km domain took

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approximately 36 hrs using 64 processors compared to 4.5 hours with 16 processors when an idealized radiation profile was used. The extended time period to complete the interactive radiation run is partially due to the extra computations done every specified time step, but also due to limitations of the number of processors SAM6.8.2 can use for a given domain setup. The total number of processors has to either be divisible by both the x-dimension and y-dimension or the number of vertical levels. Since each domain setup has 64 vertical levels, 64 processors are the most that can be used for any given simulation.

Figure F.1 – Three hour domain averaged surface precipitation with time for the 128 km x 128 km domain run with a 10-day mechanical damping and $C_H$ and $C_E$ fixed in Eqs. A.1 and A.2. The black simulation used a fixed radiation profile derived from a previous uncoupled interactive radiation run. The red simulation used the idealized radiation profile (Fig. 2.3b)

Figure F.2 shows the precipitation evolution with time of two 128 km x 128 km simulations that are exactly the same except in one the background vertical velocity (Fig. 2.1) acts on the reference T and q profiles and hence gives a constant CSRM vertical advective tendency (black line), while in the other the background vertical velocity acts on the current T and q profiles (red line). As in figure F.1, both runs were done using fixed surface exchange coefficients, a 10-day mechanical damping, and the large-scale wave vertical velocity advected the current T and q profiles. The general characteristics of the
precipitation oscillations are the same. Assuming the other domain setups show the same insensitivity, results are not sensitive to the background vertical velocity profile acting on the reference or instantaneous CSRM domain-averaged T and q profiles.

Figure F.2 – Three hour domain averaged surface precipitation with time for the 128 km x 128 km domain run with a 10-day mechanical damping and $C_H$ and $C_E$ fixed in Eqs. A.1 and A.2. In the black simulation the background vertical velocity (Fig. 2.3a) acted on the reference T and q profiles, while in the red simulation the background vertical velocity acted on the instantaneous CSRM domain-averaged T and q profiles (see Chapter 4.2 and Eqs. 4.7 and 4.8 for details).

Figure F.3 shows the precipitation evolution with time for a simulation that used a coupling frequency of every time step (i.e. every 15 seconds) and another simulation that used a 30-minute coupling frequency. There is no significant difference between the two coupling frequencies. Both simulations used a 10-day u-, T-, and q-damping with surface exchange coefficients dependent on surface stability. Additionally, the large-scale vertical velocity acted on the reference T and q profiles. Though only two coupling frequencies were tested, it is assumed that as long as the coupling frequency is shorter than the convective adjustment time scale for each domain, the results will be similar.

Figure F.3 – 128 km x 128 km simulations run with indicated large-scale coupling frequency.
APPENDIX G: Wavelength dependence

This appendix examines the robustness of the results to different large-scale wavelength choices. K08 extensively looked at how precipitation oscillations change given a wide variety of wavelength choices (i.e. 2000 km – 20 000). Here we redo the simulations described in Appendix A.1.2 and A.2.2, but for a 2500 km and 10 000 km wavelength. Recall, simulations done in Appendix A.1.2 and A.2.2 used surface exchange coefficients fixed at $1.1 \times 10^{-3}$ and a 10-day mechanical damping. Both the background and the large-scale wave vertical velocity profile advect the reference T and q profiles.

As K08 showed the period and amplitude of the precipitation oscillations has an inverse relationship with large-scale wavelength. In the coupled CSRM and wave framework wave “speed” is actually a measure of the CSRM domain averaged $T_v$ and large-scale vertical velocity cycle. That is, specified wavelength, k, and nonlinearity intrinsic to convection may affect wave “speeds” (K08). The 2500 km wavelength produces faster, higher amplitude oscillations compared to the 5000 km simulations, while the 10000 km wavelength produces slower, smaller amplitude oscillations (Figs. G.1 and G.2). Regardless of wavelength choice, though, the comparison of precipitation response across domain setups remains nearly the same (Fig. G.3). The most obvious exception is when a 2500 km wavelength is used, as the 256 km x 64 km domain actually produced a higher post-coupled maximum rainfall than the 1024 km x 16 km domain (Fig. G.3). Conclusions stated in Chapter 4, therefore, are at least robust to the varying wavelength choices examined here.
Figure G.1 – Same as figure A.22 except a 2,500 km large-scale wavelength is used.
Figure G.2 – Same as figure A.22 except a 10,000 km large-scale wavelength is used.

Figure G.3 – Same as figure A.18, except now the comparison is done across indicated wavelengths instead of damping timescale.
REFERENCES


LeMone, M. A., E. J. Zipser, and S. B. Trier, 1998: The role of environmental shear and thermodynamic conditions in determining the structure and evolution of mesoscale


