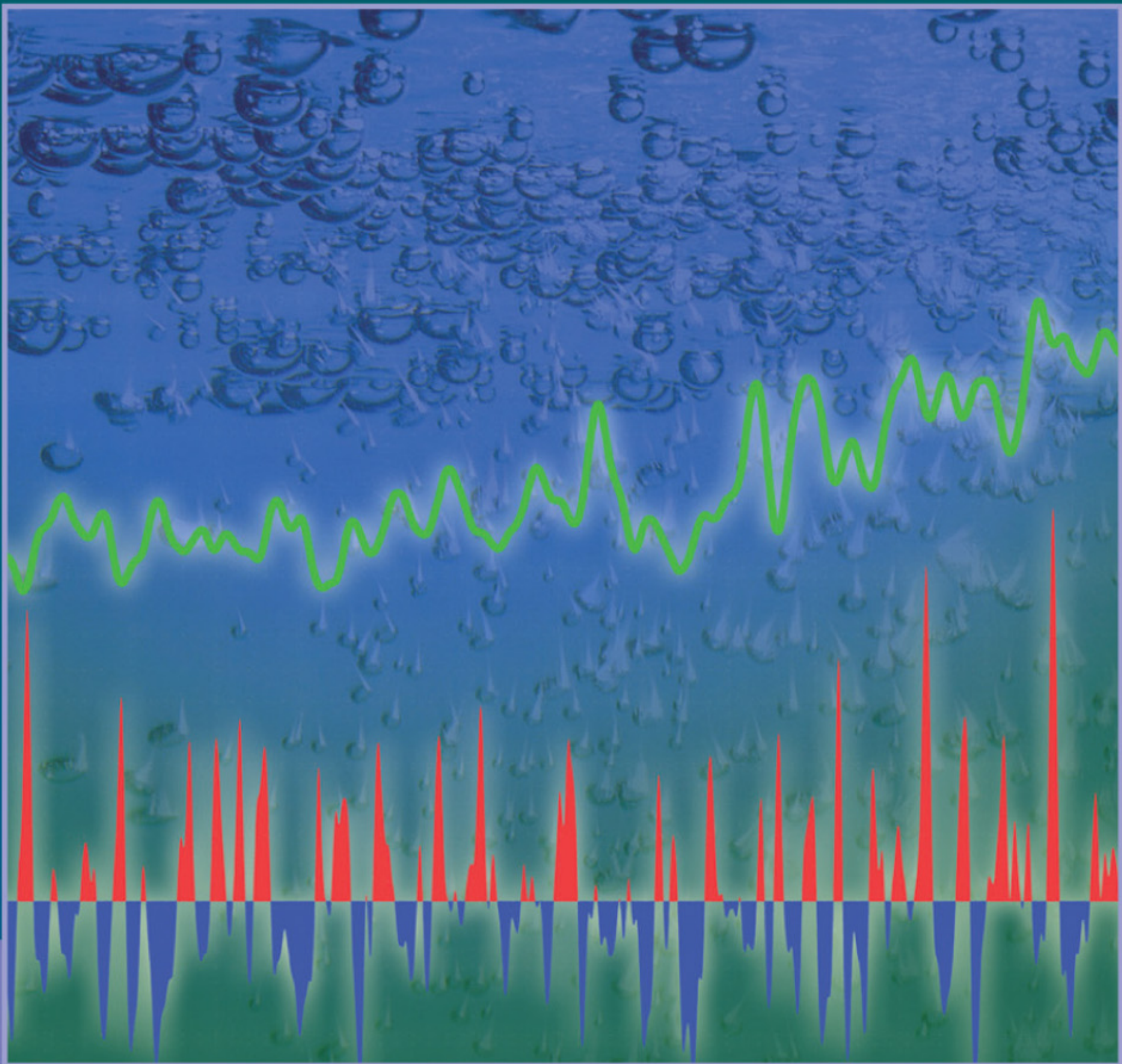
Climate Dynamics: Why Does Climate Vary?



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De-Zheng Sun
Frank Bryan
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Cover Image: Time series of the tropical maximum sea surface temperature (SST) (top curve) and the interannual anomalies of Niño 3 (90°W–150°W, 5°S–5°N) SST (bottom curve). The tropical maximum SST is obtained here by finding the maximum value of SST within the western Pacific warm pool (120°E–160°E, 5°S–5°N). Monthly SST data from the Hadley Centre for Climate Prediction and Research for the period January 1871 to January 2010 are used for the calculations. The resulting data for the tropical maximum SST and the Niño 3 SST anomalies are then smoothed by a cosine bell window with a width of 49 and 13 months, respectively.

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CONTENTS

PREFACE

As an old saying goes, we are living in an interesting age. For the first time in the Earth's history, the most intelligent dwellers on this planet are perturbing the global energy balance of the climate system to a degree that the sustainability of the planet may be threatened. Global warming has become a household phrase and has entered the realm of economical and political debate. Consequently, the scientific community is increasingly asked to provide in a timely manner, to the public and policy makers, explanations for changes observed in the state of the climate system and predictions of how it will evolve in the coming decades and centuries. With anthropogenic forcing being traditionally introduced to the public as a perturbation to the radiation balance of the climate system, the tendency to underestimate the complexity of dynamics can be high. Indeed, concepts such as bifurcations, scale interactions, reemergence of ocean temperature anomalies, and oceans' role in integrating stochastic forcing of weather events have not become as popularly known as the greenhouse effect. But natural variability arising from the complexity of the internal dynamics of the climate system has been, and will remain, a dominant driver of changes in climate. Dynamics may also make the natural variability and anthropogenic effect more intermingled than in the linear fashion that we have often assumed.

If anything has been constant with regard to the state of the climate system, it is that it has always been changing. The change has never been monotonic either. Dynamics underlies or underpins these characteristics of Earth's climate change. With our state-of-the-art models typically underestimating natural climate variability on almost all scales, the chance that our projected global warming may be too monotonic in its pace exists. Such a chance is probably not even small. Equally likely is the risk for underestimating the potency of warming arising from dynamics. With these considerations, it is not a far-fetched idea that global temperature may cease to rise as fast as it did in the last 3 decades, or it may even start to decline over some interval in the coming decades. One thing we have learned from El Niño warming, a natural warming in the climate system, is that rapid warming is made possible by the positive feedbacks. But in the presence of dynamics, the system also tends to overshoot its equilibrium and become oscillatory because of these positive feedbacks. The other point that has often been overlooked is that the behavior of the climate system under anthropogenic forcing may not be a linear supposition of a thermodynamically forced trend on the natural variability. Such an assumption may offer convenience in many situations but has little scientific basis and may even be misleading in light of the recent findings about the diabatic and nonlinear aspects of the El Niño–Southern Oscillation (ENSO). These findings suggest that the very existence of ENSO, a natural mode of climate variability, may be linked to the intensity of the radiative heating in the tropics and that ENSO events (El Niño and La Niña events) collectively play a fundamental role in the long-term heat balance in the tropical Pacific region. To be sure, the global climate system is not the same as the ENSO system. The chance for a dramatic reversal in the global temperature trend may be small, particularly with the continuing buildup of CO_2 in the atmosphere. However small the chance for this scenario to materialize, we do not want to wait until that has happened to remind us of the importance of climate dynamics. What is at stake is the credibility of climate science. If nature indeed surprises us that way and the rate of increase in global temperature slows down (or even the temperature itself starts to decline) in the coming decades, when the next wave of warming arrives with more severity, our warning of it will not be heeded by the public. The situation may not turn out to be a modern version of "the boy who cried wolf," but the lesson learned in that story may be worth recalling given the gravity of the matter. This may sound overly alarming for the sake of illustration, but natural variability is likely to dominate the decadal-scale predictions, particularly on regional scales. To advance our understanding of climate change, we need to continue the quest to understand basic climate dynamics across a range of time and space scales.

The volume provides a collection of articles on climate dynamics, aiming to underscore the potency of dynamics in giving rise to climate change and variability. These chapters originate from the lectures given in a graduate-level class at the University of Colorado at Boulder on climate dynamics. Climate experts from NOAA and the National Center for Atmospheric Research participated in teaching this class. The class was designed to expose the students to the major climate phenomena within the climate system, in particular those that

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owe their existence to the dynamical processes and may showcase why climate varies. The class was also designed to introduce to the students some basic material on climate dynamics as well as to expose them to the forefront of research. The lecturers were instructed to make the forefront research material accessible to the minds of graduate students or young researchers. The positive feedback from the students suggests that the lecturers succeeded in doing so. Lecturers generally balanced the amounts of basic material and cutting-edge research. To have more complete coverage and to replace those lecturers who were not able to convert their lectures in time to a chapter, additional climate experts from around the world were invited to contribute to the book. By covering climate phenomena over a broad spectrum of known climate variability, we hope that the book not only adequately underscores the complexity of climate dynamics but also helps readers to have a deeper appreciation of the delicate balance and complex interaction among the various forces that maintain the stability of the climate system. Such an appreciation can only help the development of a sense of urgency in advancing our understanding of the anthropogenic effect on the state of the climate system. Underscoring the importance of climate dynamics is not the same as downplaying the effect of the anthropogenic forcing. Correspondingly, the debate on the origin of the recent observed warming should not overshadow the fact that the delicate balance among the various natural forces within the climate system is being perturbed in a significant way by human activities.

We would like to thank Brian Toon and Jeff Weiss of the Department of Atmospheric and Oceanic Sciences at the University of Colorado (CU), who helped to set up this climate dynamics course. We would also like to thank Randy Dole of NOAA/ESRL, who encouraged the lead editor to take on this exercise. We would also like to thank all the NOAA and NCAR researchers and our colleagues at other institutions, University of Hawai‘i, City University of Hong Kong, and University of Tokyo, who contributed to the teaching and provided the written summaries of their lectures. Interacting with our colleagues through such an exercise has proven to be a unique opportunity for us to learn more about the dynamics of the climate system and why climate changes. We also owe gratitude to all the students who participated in this course. Their feedback has made this exercise a wonderful learning experience for us as well. Teaching at CU is fun and intellectually rewarding. We also would like to use this opportunity to thank the National Science Foundation, in particular the Climate and Large-Scale Dynamics Program (ATM 055311 and ATM 0852329), for its generous support for our research and education activities.

De-Zheng Sun
University of Colorado
National Oceanic and Atmospheric Administration

Frank Bryan
National Center for Atmospheric Research

Introduction

De-Zheng Sun

Cooperative Institute for Research in Environmental Sciences, University of Colorado, Boulder, Colorado, USA
Earth System Research Laboratory, NOAA, Boulder, Colorado, USA

Frank Bryan

Climate and Global Dynamics Division, National Center for Atmospheric Research, Boulder, Colorado, USA

Largely following the order in which the lectures were given in the graduate class on climate dynamics at the University of Colorado, the book starts with the topic of moist convection in the tropics. Summarizing decades-long research into a succinct article, *Moncrieff* [this volume] reviews the state of the art of understanding of organized precipitating convective systems with an eye to improving the representation of such systems in global weather and climate models. Moncrieff also addresses in this chapter the multi-scale convective organization in the Madden-Julian Oscillation, a major source of intraseasonal variability in the tropics. The second chapter proceeds to a prominent phenomenon on the seasonal time scale: monsoons. In covering this topic, *Li* [this volume] focuses his analysis on the Asian monsoon and dissects the physical processes that are responsible for its intraseasonal and interannual variability. All three subcomponents of the Asian monsoon are covered here: the Indian monsoon, the East Asian monsoon, and the Western North Pacific monsoon.

Chapters 3 through 6 move on to cover the El Niño–Southern Oscillation (ENSO) phenomenon: the dominant source for interannual variability in the climate system. Chapter 3 [*Penland et al.*, this volume] provides an introduction to this coupled ocean-atmosphere phenomenon, mostly from the observational perspective. In chapter 4, *Penland* [this volume] shows how this apparently complex phenomenon can be well simulated by stable linear stochastic equations. With much of the groundbreaking work on the linear inverse modeling of ENSO being done by herself and her collaborators, Penland et al. provide a complete and in-depth view of ENSO dynamics as seen within this conceptual framework. In chapter 5, *Sun* [this volume] reviews the research on the diabatic and nonlinear aspects of ENSO, in particular the efforts leading to the "heat mixer" view of ENSO. The chapter underscores the intimate connection of ENSO with radiative heating; the existence of ENSO is not only due to the dynamical coupling between the atmosphere and ocean but is also due to the fact that the warm pool sea surface temperature (SST) is sufficiently high relative to the temperature of the subsurface thermocline water. The chapter presents new evidence showing that the collective effect of ENSO events is to cool the warm pool and heat the subsurface thermocline water, reinforcing the notion that El Niño may act as a regulator of the tropical maximum SST. The chapter also provides a theoretical framework to understand how ENSO responds to global warming. In chapter 6, *Capotondi* [this volume] examines the simulation of ENSO in the state-of-the-art models and reminds readers of the continuing challenges to realistically capture the processes that are responsible for this natural model of climate variability.

Chapters 7, 8, and 9 bring the focus to the extratropics. In chapter 7, *Alexander* [this volume] examines processes that influence North Pacific sea surface temperature including the Pacific Decadal Oscillation (PDO). The role of the surface ocean in integrating the stochastic forcing from weather events, the reemergence mechanism associated with the seasonality, and the "atmospheric bridge" that connects the tropical Pacific with the extratropical regions are all covered

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in this chapter. The chapter ends with a comprehensive analysis of the causes of PDO. In chapter 8, *Nakamura et al.* [this volume] address the low-frequency variability of the extratropical planetary waves. It is these waves that cause the geographically fixed longitudinal variations in the climate of the extratropics. The three-dimensional structure and dynamical characteristics of the Northern Hemisphere climatological planetary waves are described and explained in this chapter, including their seasonal and geographical dependence. Nakamura et al. also contrast the differences in these waves between the western and eastern hemispheres and between the midlatitude ocean basins and continental regions. They also discuss the long-term changes in the planetary waves and the consequences of these changes on the predictability of the dominant modes of variability.

Chapter 9 moves to the polar regions. It deals with a polar climate phenomenon that has caused great concern to both scientists and the public: the melting of sea ice as revealed by the satellite observations. Find out in this chapter *Holland*'s [this volume] answer to the question whether the observed changes in sea ice are indicative of a tipping point behavior, leading to abrupt and irreversible changes.

The final chapter of the book addresses another topic that is of concern to both climate scientists and the public at large: Are the tropical cyclones becoming stronger because of global warming? *Chan* [this volume], a veteran watcher of tropical cyclones, carefully reviews recent studies on the change of tropical cyclone activity in the western North Pacific and discusses how it might or might not be related to global warming. He reminds the readers here that the often emphasized thermodynamic conditions are just one of the factors that influence the intensity of the tropical cyclones. The dynamical conditions cannot be overlooked in understanding the observed changes in the statistics of the tropical cyclones and in predicting future changes.

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F. Bryan, Climate and Global Dynamics Division, National Center for Atmospheric Research, P.O. Box 3000, Boulder, CO 80307, USA.

D.-Z. Sun, Physical Science Division, Earth System Research Laboratory, NOAA, Boulder, CO 80303, USA. (dezheng.sun@noaa.gov)

The Multiscale Organization of Moist Convection and the Intersection of Weather and Climate

Mitchell W. Moncrieff

National Center for Atmospheric Research, Boulder, Colorado, USA

Moist convection organizes into cloud systems of various sizes and kinds, a process with a dynamical basis and upscale connotations. Although organized precipitation systems have been extensively observed, numerically simulated, and dynamically modeled, our knowledge of their effects on weather and climate is far from complete. Convective organization is absent de facto from contemporary climate models because the salient dynamics are not represented by parameterizations and the model resolution is insufficient to represent them explicitly. High-resolution weather prediction models, fine-resolution cloud system models, and dynamical models address moist convective organization explicitly. As a key element in the seamless prediction of weather and climate on timescales up to seasonal, organized convection is the focus of the Year of Tropical Convection, an international collaborative project coordinated by the World Meteorological Organisation. This paper reviews the scientific basis of convective organization and progress toward comprehending its large-scale effects and representing them in global models.

1. INTRODUCTION

Numerical weather prediction and climate modeling are on convergent paths with respect to climate variability and change. Weather prediction has historically put extraordinary demands on numerical computation in order to advance forecast skill through improved resolution, data assimilation, and parameterization. Moving forward from their research heritage, climate models must now address the complex problem of "climate prediction," where computer power is ever more necessary. As the primary vertical transport process for thermodynamic quantities (heat and moisture), dynamical quantities (mass, momentum, kinetic energy, and vorticity), and chemical constituents in the Earth's atmosphere, moist convection is a long-standing uncertainty that compromises the fidelity of all numerical prediction systems.

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The structural complexity of moist convection is compounded by nonlinearities involving microphysics (e.g., phase changes of water) and macrophysics (e.g., latent heating, convective transport, cloud-radiation interaction, and convective organization). Atmospheric convective organization is manifested as coherent structures within fields of clouds. The fact that coherent structures occur in many fields of science (e.g., fluid dynamics, physics, chemistry, biology, and combustion) attests to the fundamental nature of convective organization.

Convective organization implies an upscale cascade of energy and has dynamical connotations involving wind shear, convection-wave interaction, and the maintenance of the atmospheric circulation against dissipation. The organization of certain shallow (nonprecipitating) cloud systems is rooted in the dynamical instability of the base state, e.g., boundary layer "cloud streets" as the Kelvin-Helmholtz instability of shear flow, and cellular convection as gravitational/diffusive Rayleigh instability, structures which may be maintained through to finite amplitude. On the other hand, moist convection is "multiscale" involving systems up to hundreds or even thousands of times the size of cumulonimbus and "multistructural" evolving into different morphological structures as time progresses. The evolution involves shear and latent heating, evaporative cooled downdraft outflows, and convectively generated waves among other processes. These systemic properties are inadequately represented by parameterizations, which compromises the interactions between moist convection, the global circulation, and the climate system.

The organization of precipitating convection has been observed for over a century [*Ludlam*, 1980]. While vertical shear had been known much earlier to affect the organization of moist tropical convection [e.g., *Hamilton and Archbold*, 1945], a quantification of the effects of shear on convective precipitation awaited weather radar [e.g., *Newton and Newton*, 1959; *Browning and Ludlam*, 1962]. Dynamical models formalized the effects of shear on convective organization and quantified its upscale properties [*Moncrieff and Green*, 1972; *Moncrieff and Miller*, 1976]. Numerical models simulated the three-dimensional (3-D) effects of shear on cumulonimbus and severe storms [e.g., *Miller and Pearce*, 1974; *Klemp and Wilhelmson*, 1978]. *Lilly* [1983] suggested that even a small amount of kinetic energy transferred upscale by convective outflows could affect synoptic-scale motion. Mesoscale circulations have a downscale effect on cumulus convection [*Cotton et al.*, 1976]. The backscatter procedure by which small-scale kinetic energy gets injected back to large-scale models has been used as a way to parameterize the upscale cascade [*Shutts*, 2005].

The assumption of a scale gap between cumulus convection and synoptic-scale motion used in contemporary cumulus parameterization offers useful simplifications such as the neglect of lateral transport of mass, energy, and momentum. Contrary to observations and dynamical theory, in terms of parameterization, the scale-gap assumption relegates convective organization to a secondary consideration. Observations have long confronted this assumption, e.g., the Global Atmospheric Research Program (GARP) Atlantic Tropical Experiment (GATE) clearly showed that mesoscale cloud clusters populate the scale gap (see the review by *Houze and Betts* [1981]). The existence of a mesoconvective continuum rather than a scale gap has been quantified by observations, simulations, and theory over decades. Lateral fluxes are an important consideration for organized systems, especially those that have a strong vertical tilt.

Ignoring the effects of organized convection undoubtedly retarded the formulation of physically based parameterizations. Until recently, parameterization was the only way by which the effects of precipitating convection in global prediction systems could be estimated. This is no longer the case. Cloud system resolving models (CRMs) simulate multiscale convective organization and its scale interactions. High-resolution global weather prediction models explicitly represent convective organization, albeit as underresolved circulations. The multiscale organization of convection can be addressed with completeness at the intersection of weather and climate (timescales up to seasonal) where high resolution is an affordable option.

This paper focuses on the organization of moist convection, its dynamical approximation and simulation by fine-scale numerical models, and its representation in global weather and climate models. The following section involves global-scale convective organization and propagating precipitation systems. The controls on moist convection are addressed in section 3, followed by fundamentals of mesoscale convective organization in section 4, and the multiscale organization of tropical convection in section 5. Multiscale convective organization in a hierarchy of numerical models is the subject of section 6, followed by its parameterization in section 7. The paper concludes with discussion in section 8 and conclusions in section 9.

2. GLOBAL CONVECTIVE ORGANIZATION

The organization of clouds into coherent systems and their association with the global atmospheric circulation is abundantly clear from satellite observations, e.g., the midlatitude baroclinic systems, the subtropical convective complex, and tropical cloud systems. The correlation between convective organization and the large-scale atmospheric circulation implies that convective organization can, in principle, be represented as functions of the resolved-scale variables, i.e., parameterized.

2.1. Midlatitude Baroclinic Systems

The baroclinic systems within the midlatitude storm tracks have long been understood as a baroclinic instability of the zonal flow, which is a convective process. The kinetic energy of motion derives from a slantwise (almost horizontal) buoyant exchange of mass by two global airstreams: a warm conveyor belt originating in the subtropics and the return cold branch from the polar regions. The meridional convergence of the meridional transport of zonal momentum associated with this mass exchange maintains the jet stream and the westerly vertical shear of midlatitudes.

A hierarchy of moist convective organization is embedded in these airstreams. Rainbands of various descriptions occur within the warm conveyor belt. In the cold branch (category A, Figure 1), flow-parallel shallow bands form near the polar ice sheets and transition downstream into open and closed cellular convection (stratocumulus). Near the cold front,

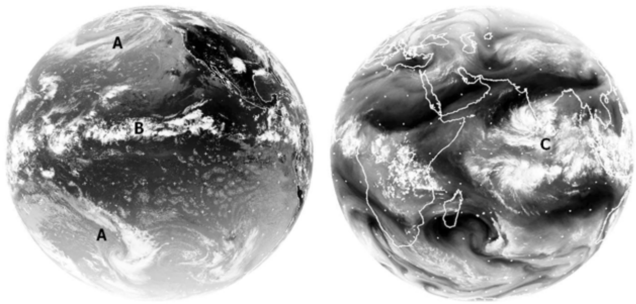

Figure 1. (left) Global image of the large-scale organization of convection, e.g., Intertropical Convergence Zone, subtropical cloud bands, and polar outbreaks. (right) Multiscale organization of deep convection, large mesoscale convective systems (superclusters), and incipient tropical cyclones associated with a Madden-Julian Oscillation (MJO) episode in the Indian Ocean. Image from NERC Satellite Receiving Station, University of Dundee, Scotland, U. K.

convective organization is manifested by clusters of cumulonimbus, rainbands, and squall lines. The largest atmosphere-ocean heat exchange on Earth (~1000 W m^{-2}) near the ice sheets cools the ocean surface and drives deep oceanic convection, forming the thermohaline circulation.

2.2. Subtropical Convective Complex

The subtropical convective complex (category B, Figure 1) is identified with the Intertropical Convergence Zone (ITCZ), fields of trade wind cumulus, and stratocumulus decks. Occurring in conditions of anticyclonic cool advection, the subtropical convective complex has evolutionary properties in common with the polar branch of midlatitude baroclinic systems, e.g., the downstream transition of shallow cumulus into deep convection. Marine stratocumulus has received much attention because of its cooling effect on the climate system. The ITCZ in the Atlantic and Pacific is multi-structural, populated by synoptic-scale easterly waves, tropical cyclones, and mesoscale cloud systems. The ITCZ in the Indian Ocean is modulated by the Asian-Australian monsoon. During boreal summer, the northward migration of the ITCZ into the Bay of Bengal affects the onset of the summer monsoon, the variability of precipitation, agriculture, and livelihood on a continental scale.

2.3. Propagating Convective Systems

Propagating precipitation systems populate sheared environments such as the midlatitude jet streams during the warm season, and subtropical jet streams and tropical wave disturbances throughout the year. Examples are mesoscale convective systems (MCS), the Madden-Julian Oscillation (MJO) [*Madden and Julian*, 1972], and convectively coupled equatorial waves. These systems were reviewed by *Houze* [2004], *Zhang* [2005], and *Kiladis et al.* [2009], respectively. Category C in Figure 1 shows the multiscale and multi-structural nature of the MJO in the Indian Ocean associated with severe weather, heavy precipitation, and floods, e.g., tropical cyclones and superclusters.

Organized propagating precipitation systems are truly a "missing process" in climate models because the pertinent dynamics are not approximated by parameterizations, and the model resolution is insufficient to represent them explicitly. Distinctions between extratropical and tropical convection feed through to parameterization. In weather prediction models, parameterization has a high fidelity in midlatitudes because, being well-resolved, the baroclinic systems provide realistic moisture and vertical shear controls for moist convection (see sections 4.1 and 4.2). Rather than being subject to downscale control, tropical convection is responsive to if not generated by an upscale cascade of energy.

3. CONTROLS ON MOIST CONVECTION

Latent heat released by moist convection is the principal source of energy for the large-scale tropical circulations, whose effects may be transmitted globally by Rossby wave propagation. These circulations are the product of nonlinear interactions among moist processes rather than a dynamical instability of the base state. Latent heating is dispersed by inertial-gravity waves up to the Rossby radius of deformation (~1000 km). The absorption of heat by evaporating liquid precipitation and melting ice drives downdrafts that cool and dry the lower troposphere: Earth's natural air

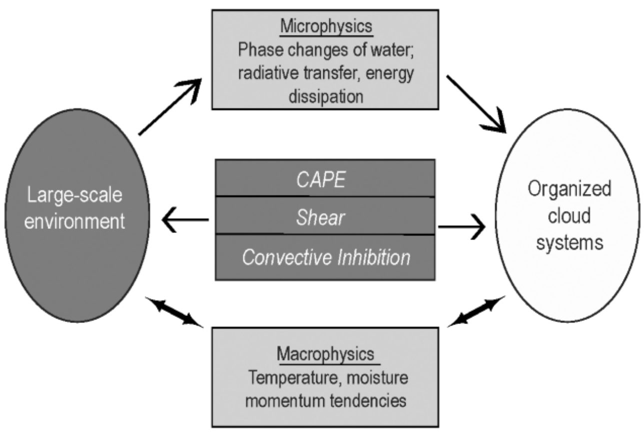

Figure 2. Association of moist convection involving convective available potential energy (CAPE), vertical shear, and convective inhibition.

conditioning system. Propagating for hundreds of kilometers, downdraft outflows (density currents) modulate atmosphere-ocean exchange. Dynamical lifting of planetary boundary layer at density-current fronts triggers new convection. In the tropics, convectively generated gravity waves foster the clustering of cumulonimbus. Vertical shear organizes deep cumulonimbus into long-lasting mesoscale systems. The "top-heavy" profile of heating (tropospheric latent heating and lower-tropospheric evaporative cooling) associated with mesoscale systems affects the tropical circulation through potential vorticity dynamics.

The following section summarizes convective available potential energy, convective inhibition, and vertical shear controls on precipitating convection (Figure 2).

3.1. Convective Available Potential Energy

The integrated buoyancy of vertically displaced moist air parcels defines the convective available potential energy (CAPE) [*Moncrieff and Miller*, 1976] for the up and down branches of convective overturning. The concept of CAPE is demonstrated by exchanging two fluid parcels of density ρ_1, ρ_2 initially at the heights z_1, z_2, respectively, where $z_2 > z_1$ and $\rho_2 > \rho_1$, i.e., the fluid is unstably stratified. The initial and final total potential energies per unit volume are $\rho_1 g z_1 + \rho_2 g z_2$ and $\rho_1 g z_2 + \rho_2 g z_1$, respectively. The total change of potential energy is $g(z_2 - z_1)(\rho_2 - \rho_1)$, and the total kinetic energy of convective overturning is $\frac{1}{2}\rho_1 W^2 + \frac{1}{2}\rho_2 W^2 = \bar{\rho} W^2$, where $\bar{\rho} = \frac{1}{2}(\rho_1 + \rho_2)$ is the average density of the exchanged parcels. The symmetry of this simple model requires that the potential energy release be shared equally by the up and down branches. Equating the potential energy to the kinetic energy for the up branch results in $\frac{1}{2}W^2 = g(z_2 - z_1)\,(\rho_2 - \rho_1)/\bar{\rho} =$ CAPE, the parcel theory of convection.

In the above simple example for an unsheared environment, CAPE is the sole source of energy. In a sheared environment, and for precipitating convection in particular, the kinetic energy of shear and propagation and the work done by the horizontal pressure gradient organize convective overturning (see section 4.1). For a moist atmosphere, CAPE is based on similar principles except that moisture affects density, and compressibility introduces potential temperature. For a moist atmosphere, CAPE $= \int_{z_1}^{z_2} g\left(\frac{\delta\theta_v}{\bar{\theta}_v} - l\right) dz$ where θ_v the virtual potential temperature represents the effects of water vapor on buoyancy, and l is the water loading. In the tropics, the water loading can deplete CAPE by 30%.

CAPE is generated by the transport of heat and moisture from the surface into the planetary boundary layer, and the large-scale advection of temperature and moisture. Dry adiabatic ascent in cyclonic regions of the midlatitude storm tracks and tropical disturbances cools and destabilizes the troposphere and generates CAPE.

3.2. Convective Inhibition

The planetary boundary layer is usually stably stratified. Therefore, a vertically displaced air parcel will be negatively buoyant unless some finite-amplitude mechanism lifts boundary layer parcels above the level of free convection: the planetary boundary layer is "metastable." The convective inhibition or negative CAPE is the vertical integral of the negative buoyancy below the level of free convection. Two mechanisms (local and nonlocal) can break the metastability barrier. The local mechanism is associated with weakly sheared environments. During daytime, the planetary boundary layer is deepened by the turbulent heat flux from the solar-heated surface. In mountainous terrain, the horizontal gradient of temperature generates upslope flow and initiates deep convection (see section 4.5). The nonlocal mechanism involves boundary layer convergence involving density currents, frontal boundaries, solitary gravity waves on the boundary layer inversion, and nocturnal downslope flow. Density currents have long been used to trigger deep convection in numerical models [*Thorpe et al.*, 1980; *Thorpe and Miller*, 1978].

3.3. Vertical Shear

The controlling effect of deep shear and its association with CAPE was demonstrated by early dynamical models and numerical simulations of squall lines [e.g., *Moncrieff and Green*, 1972; *Moncrieff and Miller*, 1976; *Thorpe et al.*, 1982] and severe convective storms [e.g., *Weisman and Klemp*, 1982]. The interaction between low-level shear and density currents initiates families of cumulonimbus multiscale squall lines and mesoscale convective systems in both midlatitudes [*Rotunno et al.*, 1988] and the tropics [*Lafore and Moncrieff*, 1989]. This dynamical triggering is most effective when the wind and wind-shear vectors point in the opposite direction [*Moncrieff and Liu*, 1999]. Baroclinic systems generate vertical shear.

The following section sets the organization of moist convection onto a rigorous basis with emphasis on propagating mesoscale systems.

4. FUNDAMENTALS OF MESOSCALE CONVECTIVE ORGANIZATION

MCSs have been extensively observed, numerically simulated, and dynamically modeled. Quoting *Houze* [2004,

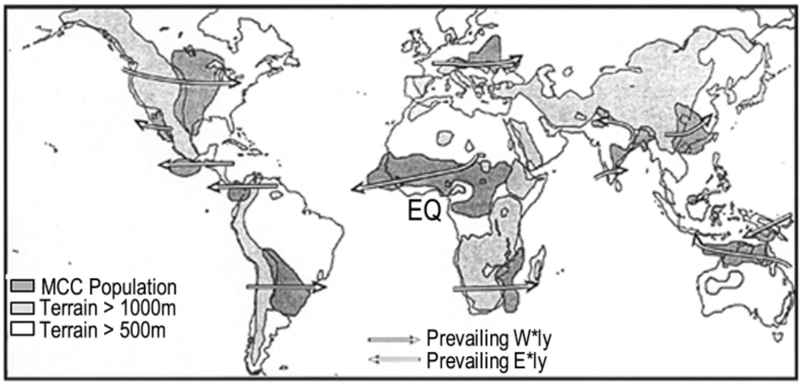

Figure 3. Global distribution of mesoscale convective complexes associated with mountainous terrain and the midlatitude/subtropical jet streams. From *Laing and Fritsch* [1997]. Copyright Royal Meteorological Society, reprinted with permission.

pp. 38–39], "Much of what we know about MCSs . . . has come from field projects and modeling studies carried out in the 1970s and 1980s." Early observations revealed extensive MCSs over the tropical oceans [e.g., *Zipser*, 1969; *Houze*, 1977; *Houze and Betts*, 1981; *LeMone et al.*, 1984]. MCS are embedded within tropical waves [e.g., *Nesbitt et al.*, 2000; *Jakob and Tselioudis*, 2003], synoptic-scale superclusters and the MJO [*Nakazawa*, 1988], and convectively coupled Kelvin waves [*Straub and Kiladis*, 2002; *Haertel and Kiladis*, 2004]. Their propagation and longevity means that MCSs affect the atmosphere and atmosphere-ocean coupling across a range of scales.

Figure 3 shows MCS and mesoscale complexes (MCC) over continents initiated in the neighborhood of mountain ranges, e.g., Rocky Mountains in the United States, the Ethiopian Highlands in Africa, the Andes in South America, the Tibetan Plateau in China, and the eastern Ghats in India. These systems propagate great distances downstream [*Laing and Fritsch*, 1997; *Carbone et al.*, 2002]. The MCC is a special subset of the global MCS population. *Maddox* [1980] defined MCCs in terms of size and longevity: cloud top area with temperature ≤−32°C over a horizontal area of 100,000 km^2 or greater and a cloud top temperature ≤−52°C over an area of 50,000 km^2 or greater, size definitions that must be maintained for at least 6 h.

4.1. Slantwise Layer Overturning in the Vertical Plane

The propagation, dynamical morphology, and longevity of MCS and the accompanying transports of mass, heat, moisture, and momentum is succinctly posed in terms of vorticity. As a class of convective motion, MCS have dynamical properties in common with density currents [*Benjamin*, 1968; *Moncrieff and So*, 1989]. The fact that evaporation-cooled descent occurs rearward of an MCS has basic consequences (see section 4.2), including hydraulic properties that make the MCS a highly efficient, if not the optimally efficient, regime of convective overturning. These aspects were unified in a nonlinear theory of steady convective overturning in shear by Moncrieff and colleagues. Originally applied as a model of squall lines and MCS (this section), this theory has been generalized to model the large-scale organization of tropical convection such as superclusters (section 5).

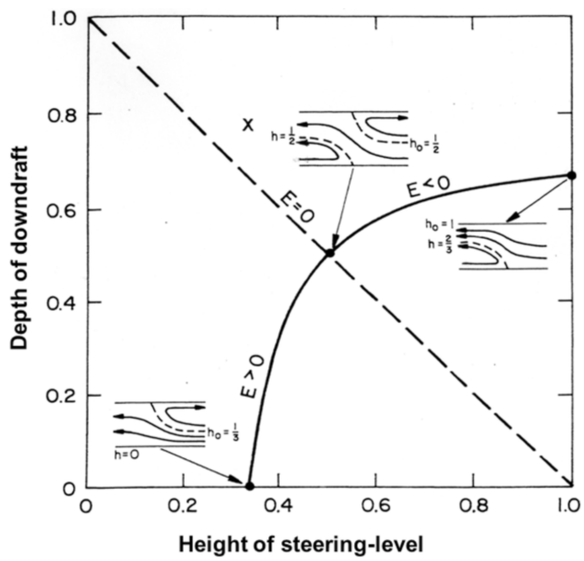

Figure 4. Regimes of archetypal organization each featuring the backward tilt of slantwise layer overturning. Rightmost inset diagram for $E = -8/9$ is purely propagating, i.e., the up branch approaches from the right everywhere. Uppermost inset diagram for $E = 0$ has symmetric up branch and down branches. Leftmost inset diagram for $E = 1$ has a hydraulic jump-like up branch but no down branch, a density current in low-level shear. From *Moncrieff* [1992].

On Figure 4, the uppermost inset diagram displays quasi-laminar branches or "slantwise layer overturning" in the vertical plane that distinguish the Moncrieff models: (1) an upward jump-like branch flows through the system without change of direction resembling a hydraulic jump, (2) an overturning upward branch, (3) an overturning downward branch. Plate 1 casts slantwise layer overturning in terms of the mesoscale circulation associated with the standard observational description of an MCS [*Houze et al.*, 1980]. The organized systems travel eastward/westward in westerly/easterly shear.

As well as the thermodynamic energy (CAPE) normally associated with deep convection, two dynamic forms of energy are fundamental to slantwise layer overturning: the kinetic energy of shear and propagation, AKE $=\frac{1}{2}(U_0-c)^2$ and the work done by the horizontal pressure gradient, WPG $= \Delta p/\rho$. The quantities WPG and AKE are functionally related through the Bernoulli work-energy principle, i.e., the change in the kinetic energy per unit mass along the bottom boundary $\left(\frac{1}{2}U_0^2-\frac{1}{2}U_1^2\right)$ equals the work done by the horizontal pressure gradient ($\Delta p/\rho$).

Quotients of CAPE, AKE, and WPG define two dimensionless quantities, the convective Richardson number R = CAPE/AKE and E = WPG/AKE. These quantities control the organization of precipitating deep convection [*Moncrieff*, 1981], rather than CAPE, shear, or pressure-work on their own. The work done by the horizontal pressure gradient expressed by E represents the hydraulic (Bernoulli) character of slantwise layer overturning. The effects of the work done by the horizontal pressure gradient on the generation and maintenance of mesoscale downdrafts were quantified in numerical simulations of tropical squall lines [*Lafore and Moncrieff*, 1989].

The effects of the convective Richardson number were illustrated by a numerical simulation of convective organization in conjunction with the variation of CAPE and shear during the passage of an easterly wave in the eastern Atlantic during GATE [*Grabowski et al.*, 1998]. Plate 2 shows transitions between nonsquall cloud clusters, a squall cluster with a trailing stratiform region, and scattered cumulus over the period of a week. The squall cluster occurred for strong vertical shear and weak CAPE, i.e., small R.

The Moncrieff 2-D models of steady convective overturning in shear are solutions of an elegant general nonlinear integral-differential equation, "the structure equation for the vertical slantwise layer overturning":

$$\nabla^2\psi - G(\psi) - \int_{z_0}^{z}\left(\frac{\partial F}{\partial \psi}\right)_{z'} \mathrm{d}z' = 0 \qquad (1)$$

where $z_0(\psi)$ is the inflow height of the stream function (ψ) defined as ($u = \partial\psi/\partial z$, $w = -\partial\psi/\partial x$). $F(\psi,z)$ is the buoyancy along streamlines or trajectories in steady flow. The first term in equation (1) is the vorticity along trajectories, the second the inflow shear, and the third the vorticity generated by the horizontal gradient of buoyancy. Equation (1) is derived from the vorticity and thermodynamic equations for 2-D flow derived from conserved Lagrangian quantities [*Moncrieff*, 1981].

Equation (1) represents each of the three airflow branches in Plate 1. Far-field solutions give the propagation speed and the lateral boundary conditions for the 2-D near-field problem. The three branches must fit together, defining a "free-boundary problem" where the shape and orientation of interfaces between the branches must be calculated as part of the solution. (Continuity of pressure is the dynamic boundary condition at free boundaries.) As shown in section 4.2, backward tilted free boundaries are vitally important for slantwise layer overturning. Special cases of equation (1) are the Helmholtz equation (neutral overturning) and Laplace's equation (unsheared inflow and neutral overturning). More mathematically tractable than equation (1), which has no known analytic solution, these simplified equations model 2-D convective overturning.

The archetypal model is the canonical regime of overturning [*Moncrieff*, 1992]. A solution of equation (1) for the hydrodynamic limit for CAPE = 0 (R = 0), the archetypal model is defined by constant inflow for the jump branch and constant inflow shear for the up and down overturning branches. Solutions exist only in the range $-8/9 \le E \le 1$. For illustration, three regimes are sketched on Figure 4: (1) the purely propagating density-current-like regime (E = 1) generalizes the *Benjamin* [1968] model to include circulation in the density current, (2) regime for E = 0 is symmetric slantwise layer overturning, and (3) the jump-like regime ($E = -8/9$) identifies the hydraulic nature of the slantwise overturning.

Generalizations of the archetypal model include 2-D buoyant overturning for $R \neq 0$ [*Thorpe et al.*, 1982] and density-current-like phenomena such as cold-frontal rainbands [*Carbone*, 1982; *Moncrieff*, 1989; *Moncrieff and So*, 1989; *Moncrieff and Liu*, 1999]. In the *Moncrieff and Miller* [1976] tropical squall-line model, 3-D overturning occurs in the plane transverse to the direction of propagation modeling the "crossover zone" observed in tropical squall lines [*Zipser*, 1969].

Slantwise layer overturning was originally developed to explain MCS-type convective organization on the ~100-km scale. *Moncrieff and Klinker* [1997] showed that this concept also explains the ~1000-km scale superclusters observed during the Tropical Ocean Global Atmosphere

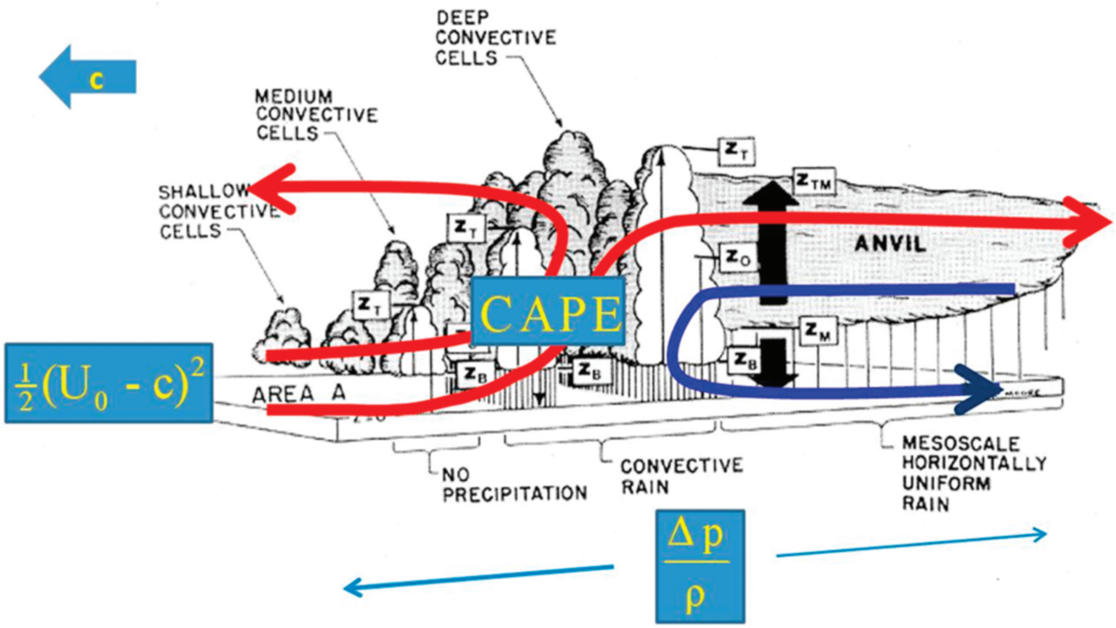

Plate 1. Underlying diagram is the standard observational description of a mesoscale convective system (MCS) propagating leftward [*Houze et al.*, 1980] consisting of shallow cumulus, medium convective cells and deep convection ahead, and a stratiform anvil region and downdraft to the rear. Overlying this diagram is the slantwise layer overturning circulation consisting of a jump up branch, an overturning up branch, and an overturning down branch and the associated three forms of energy, per unit mass: (1) CAPE, (2) the kinetic energy of relative inflow, $\frac{1}{2}(U_0 - c)^2$, and (3) the work done by the horizontal pressure gradient, $\Delta p/\rho$. Adapted from *Tao and Moncrieff* [2009].

Coupled Ocean-Atmosphere Response Experiment (TOGA COARE) simulated by the European Centre for Medium-Range Weather Forecasts (ECMWF) model. This scale invariance between MCS and synoptic-scale superclusters remains to be fully exploited.

4.2. An Existence Principle for Slantwise Layer Overturning

The Lagrangian basis of the Moncrieff models means that the far-field solutions are obtainable, along with the corresponding transports of mass, energy, momentum, and vorticity, without requiring near-field solutions. However, the far- and near-field solutions must be thermodynamically and dynamically consistent. Thermodynamic consistency of 2-D steady overturning requires that the up branches tilt backward (overlie) the down branch enabling precipitation to fall into, evaporate, and sustain the cool down branch. Dynamical consistency requires that the vertical tilt and hence the near-field momentum transport be consistent with the far-field inflow/outflow.

The upward jump is vital. Without it, the system tilts forward, contradicting the thermodynamic consistency [*Moncrieff*, 1978]. The archetypal model demonstrates this point. The upward jump produces the required backward tilt by its effects on the pressure distribution. That the upward jump is crucial for slantwise layer overturning is consistent with the trailing stratiform region being basic to MCS-type organization. The next section affirmatively answers the question: Is the existence principle upheld by numerical models and observations?

4.3. Representativeness of the Slantwise Layer Overturning

The Moncrieff dynamical models were developed side-by-side with numerical simulations [e.g., *Moncrieff and Miller*, 1976; *Thorpe et al.*, 1980, 1982; *Dudhia et al.*, 1987; *Lafore and Moncrieff*, 1989; *Liu and Moncrieff*, 2001], so these dynamical models are, by design, representative of numerical simulations. The intriguing possibility that the slantwise layer overturning model has a general application is based on the following statement made from the observational perspective. *Houze* [2004] states

> An MCS does not always take the form of a crisply defined leading convective line with a trailing-stratiform region; however, it tends to always have a stratiform region with a middle level inflow guided into the system by the environmental relative wind. The rear inflow behind squall lines appears to be a particularly clear example of the more general phenomenon of middle level inflow into and mesoscale descent within the lower reaches of a stratiform region of an MCS.

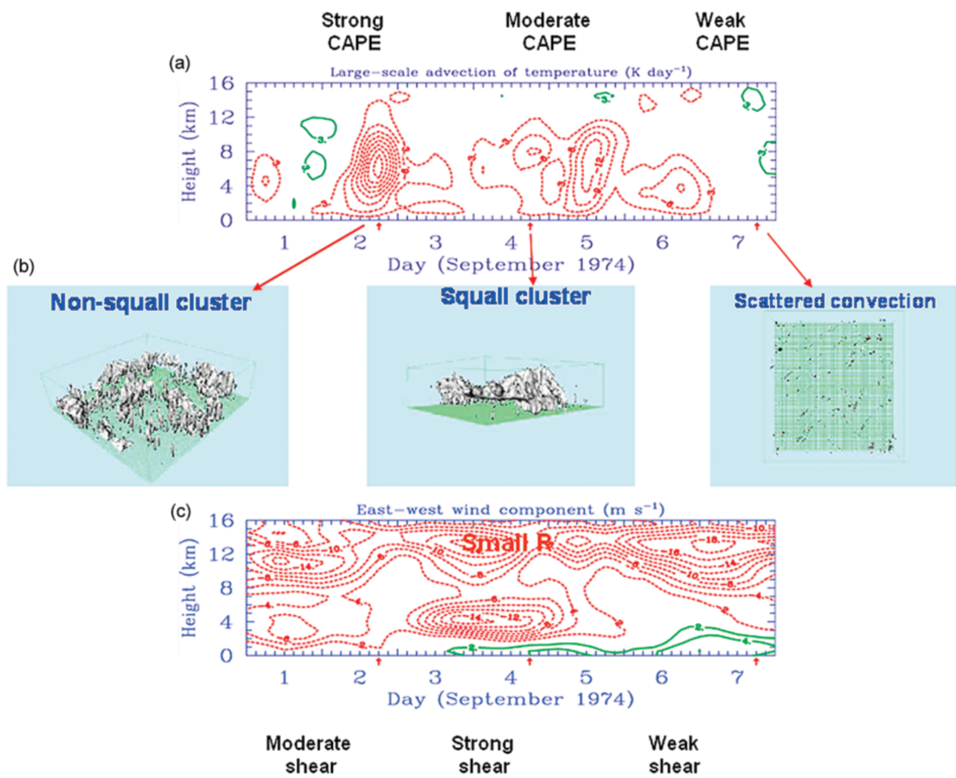

Plate 2. Effects of shear and CAPE (convective Richardson number, R) on the organization of tropical convection in a cloud system resolving model (CRM) simulation showing three regimes of convection: (a) nonsquall cluster for large CAPE and moderate shear, (b) squall cluster for weak CAPE and large shear, and (c) scattered convection for weak CAPE and weak shear. The squall cluster has the backward tilt of MCS-type convective organization. *Tao and Moncrieff* [2009].

The existence principle (section 4.2) is consistent with this quotation.

While observations do not give a precise estimate of the global representativeness of the slantwise layer overturning model, evidence on regional scales and for different climate states does support its validity. *Fritsch et al.* [1986] estimated the contribution of precipitation from mesoscale convective weather systems (74 MCCs and 32 MCSs) over the continental United States during the warm season (April–September). Examining two climatic scenarios, a “normal” year (1982) and a drought year (1983), Fritsch et al. found that mesoscale convective weather systems account for 30–70% of the warm season precipitation in the region from the Rocky Mountains to the Mississippi. The contribution is even larger in midsummer. The implication is that propagating convective weather events are “very likely the most prolific precipitation producers in the United States” and “may be a crucial precipitation-producing deterrent to drought.”

In a study of stratiform rain in the tropics estimated from the precipitation radar on Tropical Rainfall Measuring Mission (TRMM) over a 3-year period, *Schumacher and Houze* [2003] estimated that stratiform precipitation associated with slantwise overturning accounts for 73% of the rainy area and 40% of the total rain.

Kingsmill and Houze [1999] examined the momentum fields in all the MCSs observed by airborne Doppler radar in TOGA COARE. These systems contained the fundamentals of the Moncrieff 2-D model (see Plate 1). They also showed 3-D aspects of the MCSs and how the overturning and jump components of the 2-D model fit into the more complex 3-D context of natural MCSs. The Kingsmill and Houze study shows that even though MCSs in nature are 3-D, the fundamental properties of the Moncrieff model remain.

4.4. Downgradient and Upgradient Convective Momentum Transport

The convective momentum transport (CMT) per unit volume and unit length in the transverse (y) direction is $\langle \rho u'w' \rangle = \frac{1}{L}\int_0^L \rho u'w' \mathrm{d}x$, where L is the dynamical scale. The

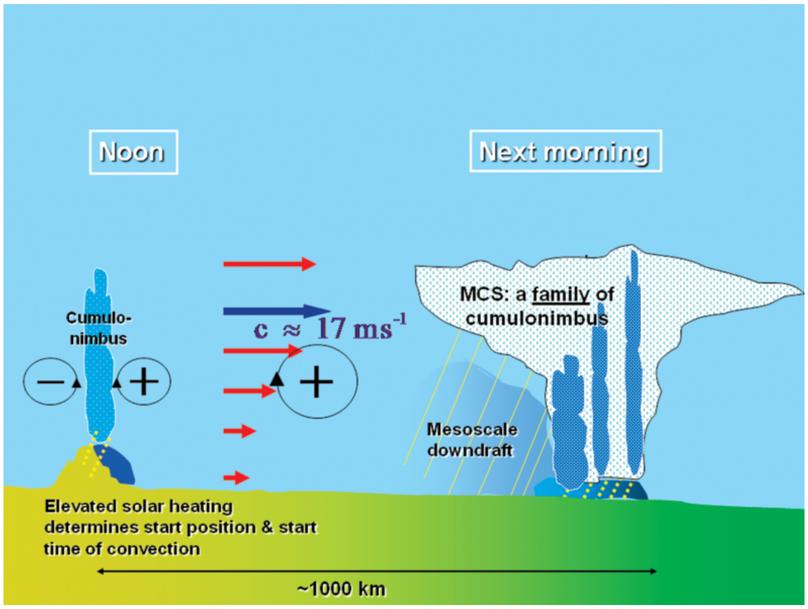

Plate 3. Conceptual model of an MCS originating over the Continental Divide. Cumulonimbus initiated by the elevated heating (baroclinic generation of horizontal vorticity) evolve in a sheared environment into multiscale systems over the Great Plains sustained by the large-scale advection of moisture in the low-level jet originating over the Gulf of Mexico.

momentum transport by cumulus convection, called cumulus friction by *Schneider and Lindzen* [1976], is parameterized by $\langle \rho u'w' \rangle = M_c(u_c - \bar{U})$, where $M_c = \sigma_c \rho \bar{w}_c$ is the updraft mass flux, σ_c the fractional area of cloud in the grid box, $\bar{w}_c(z)$ the horizontally averaged updraft speed, $u_c(z)$ the in-cloud momentum, and $\bar{U}$ the mean-flow momentum per

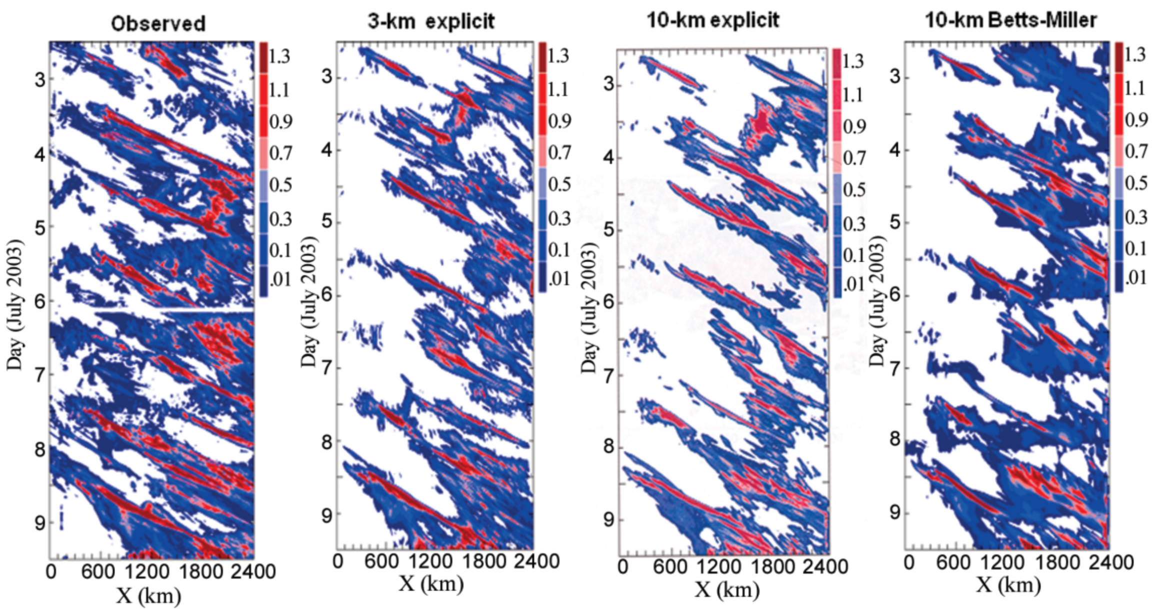

Plate 4. Precipitation rate in mm h^{-1} (left to right): Next Generation Weather Radar analysis [*Carbone et al.*, 2002], 3-km grid simulation, 10-km grid simulation, and 10-km grid simulation including the *Betts* [1986] convective parameterization. From *Moncrieff and Liu* [2006]. Copyright American Meteorological Society.

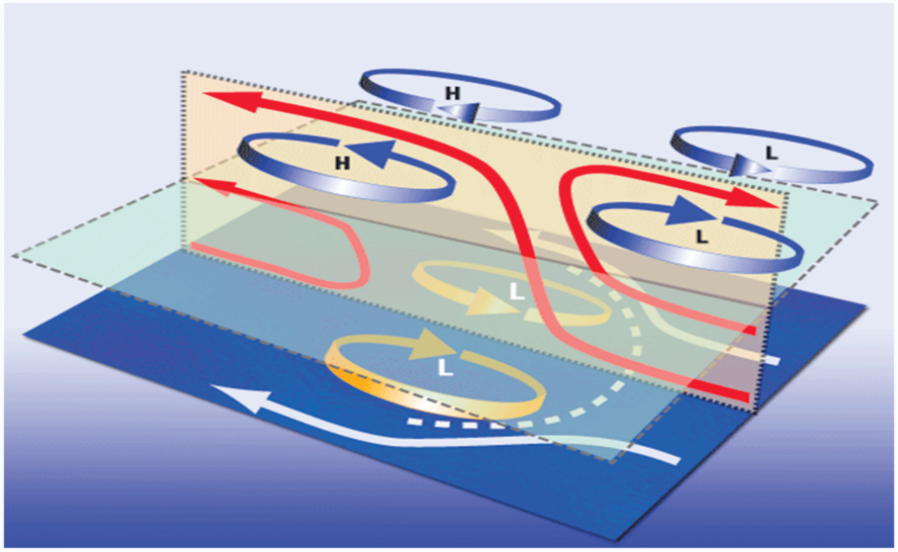

Plate 5. Dynamical models of the 3-D MJO-like system in the *Grabowski* [2001] superparameterized simulation showing vertically tilted MCS-like superclusters interlocked with a Rossby-gyre circulation approximated by a two-level model of slantwise layer overturning in the horizontal plane. These two circulations satisfy simplified forms of equations (1) and (2), respectively. From *Moncrieff* [2004a]. Copyright American Meteorological Society.

unit mass. The mean-flow acceleration is the negative of the vertical gradient of the momentum transport. Schneider and Lindzen assumed that in-cloud updraft momentum is conserved and equal to the cloud-base value. However, momentum is not normally conserved in convective updrafts due to the horizontal pressure gradient. Based on CRM simulations, *Kershaw and Gregory* [1997] approximated the pressure gradient effects on the in-cloud momentum. By

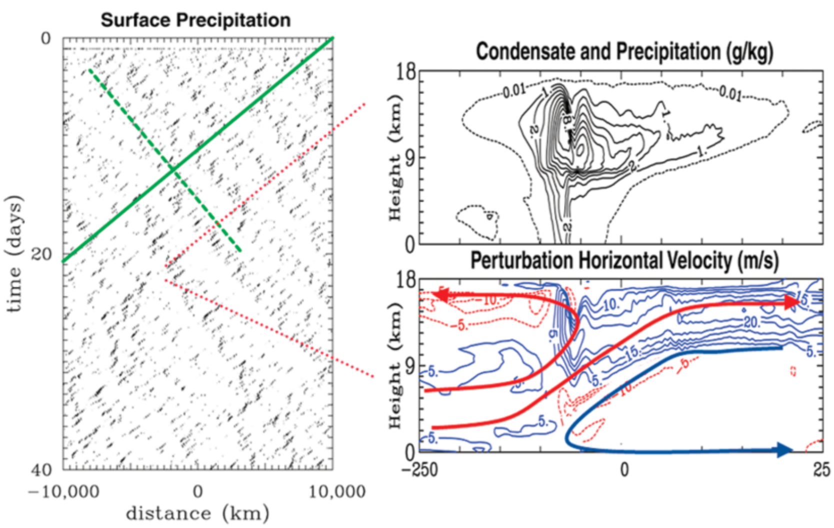

Plate 6. Multiscale convective organization simulated in a 2-D global CRM. (left) Hovmöller diagram of westward propagating precipitation systems embedded in eastward propagating cloud envelopes. (top right) Vertical section of the condensate and precipitation. (bottom right) Westward propagating MCS-like systems approximated by slantwise layer overturning. Adapted from *Grabowski and Moncrieff* [2001]. Copyright Royal Meteorological Society, reprinted with permission.

reducing the difference between in-cloud and mean-flow momentum, the pressure gradient brought the parameterized momentum transport into closer agreement with the CRM simulations. The convective momentum transport represented in the above way is not necessarily downgradient. When $u_c > \bar{U}$, upgradient transport occurs because the mean flow is accelerated.

The organization of moist convection is associated with distinctive mesoscale momentum transport (MMT). The vertical integral of momentum flux divergence is zero for steady flow bounded above and below by horizontal boundaries. In other words, although horizontal momentum can be redistributed, should shear increase in a particular layer, it must decrease in another, i.e., both upgradient and downgradient transport of momentum will occur. The sign of the MMT is opposite to that of the propagation vector, i.e., an eastward propagating system is associated with westward momentum transport. Its magnitude peaks near the middle of the convective layer, consistent with field-experiment analysis [*LeMone et al.*, 1984; *Wu and Yanai*, 1994]. The archetypal MMT agrees with numerical simulations [*Wu and Moncrieff*, 1996] and observations [*LeMone and Moncrieff*, 1993]. The kinetic energy generation is comparable to the rate of change of CAPE [*Wu and Moncrieff*, 1996]. More information can be found in the work of *Moncrieff* [1997].

Houze et al. [2000] gave empirical evidence for how the mesoscale circulations associated with MCSs can feedback either positively or negatively to the large-scale circulation of the MJO. In the strong westerly wind zone of the MJO, the MMT reinforces the larger-scale structure. *Mechem et al.* [2006] present model results that support the empirical evidence that MMT feeds back to the larger-scale wave. These downdraft-related transports of momentum present complications that may need to be considered in a complete representation of momentum transport by MCS. The Kelvin-Rossby wave structure of the MJO also organizes convection as seen in the analysis of TOGA COARE observations [*Houze et al.*, 2000].

Tung and Yanai [2002a, 2002b] studied convective momentum transport associated with the MJO, tropical waves, squall, and nonsquall MCSs. They examined the momentum budget deduced from the objectively analyzed observations during TOGA COARE in the intensive flux array (IFA) at $2.5° \times 2.5°$ areal resolution. The IFA-mean kinetic energy transfer is downscale for about 60–65% of time in the lower troposphere, but in the upper troposphere, upscale and downscale kinetic energy transfers occur with similar frequency. In other words, different kinetic energy transfers are associated with different regimes of convective organization (recall the role of R and E). Upscale kinetic energy transfer occurs in the line-normal direction of squall lines. During the westerly wind phase (burst) of the MJO, the convective momentum transport is upgradient, and the upscale kinetic energy transfer assists the westerly wind burst. In the subsequent strong low-to-midlevel westerlies, the momentum transport is mostly downgradient reducing the shear in midtroposphere.

4.5. Orogenic Mesoscale Convective Systems

Using brightness temperature obtained from satellite-based observations as a proxy for deep convection, *Laing and Fritsch* [1997] showed a relationship between MCCs, orography, and the midlatitude/subtropical jet streams (Figure 3). Using data from the surface-based network over the continental United States, *Carbone et al.* [2002] showed that during the warm season (May–October), episodes of MCS originate over the Continental Divide, propagate eastward for ~1000 km over the continental United States in the westerly shear flow characteristic of that region. The episodic nature of these MCSs is indicative of upper tropospheric eastward traveling short waves, which episodically generate CAPE and shear. The MCSs may evolve nocturnally into MCCs over the Great Plains (Plate 3) when the low-level jet of moisture from the Gulf of Mexico penetrates deep into the Midwest. The nocturnal maximum of precipitation is partly due to CAPE generated by the advection of moisture by the low-level jet originating over the Gulf of Mexico. The diurnal cycle of energy is affected on a continental scale [*Knievel et al.*, 2004]. The large nocturnal systems tend to be more 3-D than MCS and during the later stages of evolution may develop synoptic-scale vortices that further prolong their life.

Tripoli and Cotton [1989] simulated diurnal convection in the lee of the Rocky mountains and proposed a conceptual model of the life cycle of orogenic propagating convection. The *Moncrieff and Liu* [2006] 3-D simulations were initialized and forced by global analysis provided by the National Centers for Environmental Prediction (NCEP). The simulated squall lines resemble those of *Davis et al.* [2003] for other observed episodes. The precipitation patterns produced by explicit convection at 3-km grid spacing, explicit convection at 10-km grid spacing, and hybrid (explicit plus parameterized) convection at 10-km grid spacing were compared with radar measurements (Plate 4). The MCS propagation and the distribution of precipitation are similar. The precipitation is mostly from the explicit (grid-scale) circulation, not the parameterized convection. The grid-scale circulations do not approximate MCS unless the grid spacing is at least 10 km.

The simulated MCS over the U.S. continent displays the backward tilt characteristic of slantwise layer overturning.