Assessment of a General Circulation Model with Modified Convection and Clouds

Zachary Alan Mangin
Iowa State University

Follow this and additional works at: http://lib.dr.iastate.edu/etd
Part of the Climate Commons, Environmental Indicators and Impact Assessment Commons, and the Meteorology Commons

Recommended Citation
Mangin, Zachary Alan, "Assessment of a General Circulation Model with Modified Convection and Clouds" (2013). Graduate Theses and Dissertations. 13059.
http://lib.dr.iastate.edu/etd/13059

This Thesis is brought to you for free and open access by the Graduate College at Iowa State University Digital Repository. It has been accepted for inclusion in Graduate Theses and Dissertations by an authorized administrator of Iowa State University Digital Repository. For more information, please contact digirep@iastate.edu.
Assessment of a general circulation model with modified convection and clouds

by

Zachary Alan Mangin

A thesis submitted to the graduate faculty
in partial fulfillment of the requirements for the degree of
MASTER OF SCIENCE

Major: Meteorology
Program of Study Committee:
Xiaoqing Wu, Major Professor
Tsing-Chang (Mike) Chen
William Gutowski
Raymond W. Arritt

Iowa State University
Ames, Iowa
2013
Copyright © Zachary Alan Mangin, 2013. All rights reserved.
# TABLE OF CONTENTS

## GENERAL ABSTRACT

### CHAPTER 1. GENERAL INTRODUCTION

1. Background and motivation 1
2. Thesis organization 4

### CHAPTER 2. GENERAL MODEL BACKGROUND

1. The NCAR General Circulation Model 5
2. The Iowa State University General Circulation Model 5

### CHAPTER 3. IMPACTS OF CONVECTION AND CLOUD PROCESSES ON GLOBAL CLIMATE SIMULATIONS

1. Abstract 7
2. Introduction 8
3. GCM simulations and observational data 12
   a. Control and experimental simulations 12
   b. Observational datasets 12
4. Results 14
   a. Precipitation characteristics 14
   b. Radiation budgets 17
   c. Impacts of clouds 18
   d. Diurnal cycle of precipitation 22
   e. Surface heat fluxes 25
5. Discussion 26
6. Conclusions 28
7. Acknowledgements 30
8. References 30
9. List of Tables and Figures 37
CHAPTER 4. CONVECTION SCHEME IMPACTS ON WATER VAPOR FLUX DURING NORTHERN SUMMER IN GCM SIMULATIONS

1. Abstract 79
2. Introduction 80
3. GCM simulations and observational data 83
   a. Control and experimental simulations 83
   b. Data 84
4. Methodology 84
5. Results 85
   a. Vertically-integrated moisture 85
   b. Stationary modes: $\Psi_Q$ and $\bar{\chi}_Q$ 86
   c. Perturbations: $\Psi'_Q$ and $\chi'_Q$ 87
6. Conclusions 88
7. Acknowledgements 89
8. References 90
9. List of Tables and Figures 93

CHAPTER 5. GENERAL CONCLUSIONS 99
1. Mean climate differences 99
2. Water vapor flux differences 99
3. References 100
4. Acknowledgements 104
GENERAL ABSTRACT

General circulation models (GCMs) allow atmospheric scientists to tinker with atmospheric processes and study the resulting climate trends. Atmospheric trends, such as temperature fluctuations, wind shifts, and precipitation patterns are extensively studied in an attempt to realize their impacts on people, places, and other natural processes. Although useful, GCMs have shortcomings with respect to the representation of subgrid-scale meteorological processes, and thus, parameterization is required. One of the toughest components to simulate in climate models is that of clouds, as they are variable over time and spatial scales. Cumulus parameterizations, used to represent convection, have major implications for the precipitation. Cloud-resolving model (CRM) experiments have aided in the improvement of convection parameterizations. Depending on convection closure and trigger mechanisms, precipitation may be suppressed or occur more often. The cumulus scheme also alters the radiation budget as radiation processes are coupled with hydrological ones. The National Center for Atmospheric Research (NCAR) General Circulation Model (CTL) and the Iowa State University General Circulation Model (EXP) are two such models used to study differences in parameterizations, specifically those to convection. Convection scheme modifications in EXP (based on CRM studies) are found to produce closer to observed mean climate simulations in precipitation, convection, and cloud-related variables. A diurnal cycle of precipitation more resembles observations in EXP than CTL. EXP’s precipitation occurs less frequently but with more vigor than CTL. Through decomposition of the water vapor flux into rotational and divergent wind components, we find EXP to have a more distinguishable Southeast Asian Monsoon trough and generally stronger convergent centers in monsoon regions. This agrees with precipitation in EXP being less frequent but more vigorous than CTL. Eddy components of the water vapor flux for each model simulation appropriately indicate poleward water vapor transport.
CHAPTER 1. GENERAL INTRODUCTION

1. Background and motivation

Researchers apply general circulation models (GCMs) in climatological studies to compare long-term means to observations (Hack et al. 2006; O’Gorman and Schneider 2009; Watanabe et al. 2010; Kay et al. 2012), to understand complex feedbacks (Guilyardi et al. 2009; Sun et al. 2009; Huybers 2010; Gettelman et al. 2012), and to improve climate simulations through various parameterizations (Zhang and Mu, 2005; Ming et al. 2007; Wu et al. 2007; Chao 2012). GCMs can be uncoupled, that is, they can represent one system component, such as the atmosphere, or they can be coupled atmosphere-ocean models. Regardless, parameterization is required in both GCM types, because not all processes can be resolved on the coarse grid spacings. The research that follows focuses on uncoupled atmospheric models. Subgrid processes must be represented and cumulus convection is no exception. Since clouds exist on many temporal and spatial scales, this adds to simulation difficulty.

Several other issues regarding the modeling of convection and cloud fields are at the forefront of climate research. For example, insufficient cloud output over the southern oceans is one present problem that probably results from too few extratropical cyclones (Naud et al. 2012). Morrison et al. (2011) also stated cloud deficiencies exist in the southern oceans, particularly due to biases in shortwave radiation. Such findings are also coincident with Trenberth and Fasullo (2010).

A second shortcoming with current GCMs, their production of a double Inter-Tropical Convergence Zone (ITCZ), has been identified in several recent studies (Lin 2007; Delworth 2012; Noda et al. 2012). The double ITCZ causes the southern branch of the ITCZ to parallel the northern branch instead of slanting. Excessive precipitation results. The spurious double precipitation structure is usually amplified in ocean-atmosphere models. Another way to identify this problem is by describing it as an error in the depiction of the South Pacific Convergence Zone (SPCZ). Too cold sea surface temperatures (SSTs) along the Equator and warmer SSTs to its south characterize the double ITCZ (Mechoso et al. 1995). Flux-adjusted GCMs have
somewhat corrected the double ITCZ as found by Brown et al. (2011), but double ITCZ precipitation is still a known problem in the atmosphere-only simulations. Since uncoupled atmospheric model runs also suffer from a double ITCZ, Lin (2007) referred to the problem as “intrinsic” to the model. Yet another recent area of investigation in regards to GCMs is that of excessive precipitation over steep orography. This may be due to inadequate representation of upward heat transport. Chao (2012) corrected the bias with a subgrid-scale heated-slope-induced vertical circulation parameterization, but it has yet to be determined if this parameterization can be combined in a cumulus scheme.

A third area of concern for GCMs is that of the diurnal cycle of convection, primarily in coastal monsoon areas (May et al. 2012), and near orographic features and small islands (Ploshay and Lau 2010). Land-sea regions, for example, are very difficult to simulate, because a variety of thermal and dynamical processes act to establish diurnal differences. Studies have shown that diurnal convection is stronger over land than over oceans (Dai 2001). Tropical land areas have precipitation peaks in mid-afternoon to late evening due to the heating of the planetary boundary layer. Temperature increases throughout the day over land and instability grows, which results in the stronger precipitation (Yang and Slingo 2001). Mesoscale convective systems (MCSs) contribute to peak convective intensity in the late evening (Nesbitt and Zipser 2003). Other studies have shown convection over oceans peaks in early morning hours (Gray and Jacobson 1977; Hirose et al. 2008; Rauniyar and Walsh 2011). Woolnough et al. 2004 found that precipitation peaked 2-3 hours before cold cloud maxima. The differences in precipitation peak times over land and open ocean are still not completely understood. In fact, several hypotheses have been proposed to explain early morning peaks in convection over ocean. These hypotheses include: nighttime destabilization of the atmospheric profile by infrared cooling and stabilization of the profile during day (Randall et al. 1991), vertical circulations induced by differences in cloudy and cloud-free heating rate (Gray and Jacobson 1977), and the lifetime associated with convective systems triggered by near-surface heating (Chen and Houze 1997). Modification to convection parameterization is an appropriate way to test and improve the aforementioned GCM deficiencies, but one must first know the basics of such schemes.

Cumulus parameterization requires a set of equations to govern the collective effects of clouds. Manabe et al. (1965) laid the groundwork for cumulus parameterization by introducing the simplistic moist-convective adjustment scheme. This scheme took into account latent heat
release as moist air moves up and causes the model atmosphere to become stable. In other words, when large-scale destabilization occurs, nature attempts to stabilize the atmosphere through convection. Net convective transport needs to equal zero between cloud base and cloud top in Manabe et al.’s scheme. Additional requirements for this scheme include an environmental lapse rate equal to the moist adiabatic lapse rate and relative humidity of one hundred percent after convective adjustment. Convective closure in Manabe et al.’s scheme is an assumption on equilibrium states while condensation takes place. Additional closure in the moist-convective adjustment constrains condensation that couples $Q_1$ (apparent heat source) and $Q_2$ (apparent moisture sink) as in Yanai et al. (1973).

Arakawa (1969) developed the first three-layer general circulation model that took into account cumulus mass flux and detrainment from clouds. Kuo (1965, 1974) came up with a scheme based on cloud formation from large-scale moisture convergence. Kuo split up moisture convergence into two parts: storage and precipitation, in which $Q_2$ is related to large-scale advective processes. The original study by Arakawa was later extended to include more than one cloud type. Arakawa and Schubert (1974) followed Yanai et al. (1973)’s study on bulk properties of cloud clusters to define budget equations for $Q_1$ and $Q_2$.

As cumulus parameterizations studies progressed into the 1980s and 1990s, deep convection and its atmospheric effects became more stressed. Zhang and McFarlane (1991) published their work concerning stabilization of the atmosphere in the midlatitudes. This preparatory work brought upon new insights concerning atmospheric stabilization. Specifically, upper-air data from the Preliminary Regional Experiment for Storm-scale Operational and Research Meteorology (PRE-STORM) was used in the study and four categories of atmospheric soundings were named: presystem, insystem, postsystem, and environment. Convective available potential energy (CAPE) values in each sounding type were analyzed and it was found that from presystem to postsystem, instability (CAPE) decreased for both diluted and undiluted parcel ascent. Any entrainment added a further decrease in CAPE and in some cases (entrainment rate, $\lambda = .14 \text{ km}^{-1}$), a complete stabilization occurred. In short, results from this 1991 study shed light on the importance of thermodynamic instability and entrainment effects on atmospheric stabilization and subcloud layer stabilization for convection parameterization schemes.

Later, Zhang and McFarlane (1995) implemented a deep cumulus convective scheme (ZM scheme) based on the previous studies. It was not until 2002 that a revised Zhang and
McFarlane deep convection scheme (RZM) sought to modify the parameterization based on midlatitude instability studies. The RZM was mostly based on the result of net change in instability (CAPE) being tied to that of large-scale atmospheric processes. This modified quasi-equilibrium assumption varied from Arakawa and Schubert’s in that CAPE changes from boundary layer thermodynamics are neglected. Convective inhibition is strongest in areas of weak large-scale forcing. Throughout development of this and other cumulus schemes, improved understanding of model implementation has resulted.

Cumulus parameterization is critically important for many reasons. Because few datasets exist, cloud deficiencies are a leading cause of uncertainty in climate modeling, especially during the warm, rainy season. Errors in clouds imply errors in precipitation and other cloud-related variables. Variables, such as surface temperature and sea level pressure are modeled relatively well compared to precipitation (Randall et al. 2007). Clouds and radiation, which dictate atmospheric instability, impact hydrological variables. If not much instability is present, then less vigorous precipitation may occur, for example. Therefore, handling of precipitation by models is not ideal. Precipitation is not continuous like surface temperature or sea level pressure fields (Dai 2006). Thus, more challenge is added to the climate simulation. Therefore, it is of essence that convection parameterization advances to better all aspects of climate modeling. We present our own work with GCM simulations in the journal papers that follow.

2. Thesis organization

This thesis is organized in journal paper format. It contains three chapters. Chapter 2 briefly discusses model background. Chapters 3 and 4 are stand-alone papers. Both will be submitted to the American Meteorological Society for publication in *Journal of Climate*. The first paper, comprising Chapter 3, determines the mean climate state of a general circulation model with modified convection, radiation, and cloud schemes (from cloud-resolving experiments). The second paper, comprising Chapter 4, breaks down the hydrological circulations in regards to water vapor and compares the two GCM simulations during Northern Hemisphere summer. In Chapter 5, implications are made connecting results from Chapters 3 and 4 and vice versa.
CHAPTER 2. GENERAL MODEL BACKGROUND

1. The National Center for Atmospheric Research General Circulation Model

A version of the National Center for Atmospheric Research (NCAR) General Circulation Model, the Community Climate Model Version 3.0 (CCM3) served as the control (CTL) model in the research that follows. Sufficient computing power made this GCM the most efficient to work with for a scientific study. The uncoupled atmospheric CCM3 is composed of 18 vertical hybrid sigma levels. Resolution in the horizontal is of truncation 42 (T42), or approximately 2.8° x 2.8°. Because of insufficiencies in previous NCAR models with cloud radiative forcing, surface land temperature, and an overly active hydrologic cycle, changes to parameterization were employed in the updated model. Specifically, clouds are now able to form in any model layer except closest to the surface, a relative humidity threshold controls mid and upper-level cloud formation, and a previous minimum cloud fraction of 20% from CCM2 has been deleted. Cloud microphysics in the CCM3 are treated diagnostically. Cloud water concentration treats cloud condensate based on empirical results. This is similar to CCM2. Cloud droplet sizes over continental and maritime areas are also distinguished in the NCAR GCM. In the CCM3, the effective radius of the cloud droplet is 10 µm over water, but it varies based on temperature over land. For deep convection, a Zhang and McFarlane (1995) cumulus scheme is used, which has the closure condition in which convective available potential energy (CAPE) is consumed at an exponential rate by cumulus convection (Kiehl et al. 1998).

2. The Iowa State University General Circulation Model

Based on the NCAR GCM, the Iowa State University General Circulation Model (ISUGCM) is also an uncoupled atmospheric model. This experimental (EXP) model is of the same horizontal and vertical resolutions, but modifications have been made to the convection scheme in ISUGCM. A revised Zhang and McFarlane deep cumulus scheme has been implemented that links destabilization in the troposphere to large-scale moisture and thermal advection (revised convection closure). A trigger condition allows deep convection to occur once an increase in CAPE of at least 65 J kg⁻¹ h⁻¹ occurs. Convective momentum transport (CMT) is included in the model because it was found to redistribute horizontal momentum in the vertical. Cloud-resolving model (CRM) studies validated the CMT scheme (Zhang and Cho 1991a, b; Wu
et al. 2003; Zhang and Wu 2003). Redistribution of liquid ice and water is considered as cloud water path uses a scaling factor from CRM simulations (Wu and Liang 2005; Wu et al. 2007). Radiation is treated with a mosaic cloud scheme in which different clouds and their optical properties are accounted. Liang and Wang (1997) developed the mosaic cloud scheme, in which GCM grid columns are divided into cloudy and clear subcells. The cloudy subcells then are treated as one of the following: convective (Cc), anvil cirrus (Ci), or stratiform (Cs). This improves subgrid variability and allows the model to simulate more closely to observed high cloud amounts. It is to be determined whether modifications made to CTL impact mean climate as well as cloud and convection fields.
CHAPTER 3. IMPACTS OF CONVECTION AND CLOUD PROCESSES ON GLOBAL CLIMATE SIMULATIONS

A paper to be submitted to *Journal of Climate*

Zachary A. Mangin and Xiaoqing Wu

1. Abstract

In this study, modifications made to a global climate model’s convection parameterization scheme are under study to determine their impacts on mean climate variables related to convection/cloud processes. The convection parameterizations in the experimental model and a control model vary in closure conditions, although CAPE is key to the performance of each. The Iowa State University General Circulation Model (ISUGCM) is a modified version of the National Center for Atmospheric Research General Circulation Model (NCAR GCM). In ISUGCM (EXP), revised closure relates destabilization of the tropospheric layer to large-scale thermal and moisture advections. With NCAR GCM (CTL), a trigger condition initiates convection when the threshold value of CAPE is reached. On the other hand, when the rate of CAPE increase exceeds a threshold rate, EXP initiates deep convection. Differences in trigger conditions between CTL and EXP have implications for the large-scale advection processes. Subgrid variability in EXP is represented by the cloud mosaic radiation scheme, which helps bring radiation values closer to those observed. Cloud amount, liquid water, and ice tend to be better accounted for through the cloud mosaic method. Generally, EXP creates more vigorous but less frequent convection than CTL. A mean climate state more reflective of observations, especially in the precipitation field, helps produce a climate closer to other fields, too. Previous studies have shown correct seasonal migration of the major tropical rain belts by including convective momentum transport. The diurnal cycle of precipitation reveals that EXP achieves a closer to observed frequency of precipitation, intensity, and timing of precipitation maxima for most locations. A better represented diurnal cycle of precipitation aids in more realistic sensible and latent heating, which affect radiation, and go hand-in-hand with the mean climate simulation.
2. Introduction

Convection, cloud, radiation and precipitation processes are key components of the global water and energy cycle and operate on a wide range of spatial and temporal scales (e.g., Chahine 1992; Houghton et al. 2001). Water exists in three phases (vapor, liquid and solid) in the atmosphere. The formation and growth of clouds and precipitation are associated with water phase changes. Convective clouds affect local thermodynamics, large-scale circulations and wave disturbances through the release of latent heat and redistribution of heat, moisture and momentum; especially precipitation. The vertical and horizontal distributions of clouds influence the atmospheric radiation budgets and the radiative heating/cooling rates through the reflection, absorption and emission of solar and terrestrial radiation. Since solar radiation is the key factor driving the water cycle, and the interaction of water vapor and clouds with radiation modulates the transformation of energy, the energy cycle is closely coupled with the water cycle.

General circulation model (GCM) simulations of the water and energy cycle should be able to produce the mean, diurnal, seasonal, annual, and interannual characteristics of global precipitation, cloud and radiation simultaneously consistent with surface and satellite observations. However, much uncertainty in parameterization of convection and clouds limits the ability of GCMs to reproduce past and current climate mean state and climate variability (e.g., Browning 1994; Slingo et al. 1994; Arakawa 2004; Kiehl and Gent 2004; Lau 2002, 2005; Zhang 2005; Dai 2006; Deser et al. 2006; Rasch et al. 2006). The Fourth Assessment Report of the Intergovernmental Panel on Climate Change (Solomon et al. 2007) highlights clouds as the largest uncertainties in climate model predictions. Long-standing biases present in the global distribution of precipitation in GCM simulations include the South Pacific convergence zone (SPCZ) extending too far east and too zonal, the peak of Inter-Tropical Convergence Zone (ITCZ) precipitation remaining north of the equator throughout the year, and the overestimation of light precipitation and the underestimation of heavy precipitation (e.g., Hack et al. 1998, 2006; Dai 2006). Numerous issues associated with clouds are present in GCMs, and satellite data have reinforced this fact. The International Satellite Cloud Climatology Project (ISCCP) data reveal the most recent GCMs are unable to accurately produce cloud simulations. Comparing model output to that of ISCCP, three main sources of error exist, which include a lack of total clouds, too many thick clouds, and too few midtopped clouds (Kay et al. 2012). These problems are
largely due to the tuning of cloud amount and/or cloud liquid and ice water contents in order to maintain the global radiation budget closer to satellite observations (e.g., Liang and Wang 1997; Wu and Liang 2005; Song and Zhang 2012).

The cloud-resolving models (CRMs) explicitly resolve convection, and there have been attempts to implement them into GCMs in place of convection and cloud schemes (Grabowski 2004; Khairoutdinov et al. 2005; Zhu et al. 2009). A CRM approach seeks to simulate cloud features given large-scale conditions. The drawback to including CRMs into the GCM framework (superparameterization) is the amount of required computer time. Since finer cloud features are resolved with CRMs, the computational cost to run such a model is several hundred times greater than the traditional single column model (Randall et al. 2003). Use of CRMs though does indicate the ability to depict more realistic cloud systems, including cloud geometric and radiative properties under the large-scale conditions collected in the field experiments, such as the Global Atmospheric Research Program Atlantic Tropical Experiment (GATE) and the Tropical Ocean Global Atmosphere-Coupled Ocean Atmosphere Response Experiment (TOGA-COARE) (e.g., Grabowski et al. 1996, 1998, 1999; Wu et al. 1998, 1999; Wu and Moncrieff 2001).

Knowledge gained from the analysis of CRM simulations has led to improvements in parameterizations of convection, clouds, and radiation in GCMs. They include a revised closure assumption, a trigger condition, convective momentum transport (CMT), and mosaic treatment of subgrid cloud distribution. The closure assumption in the original Zhang and McFarlane (1995) scheme, currently used in most versions of NCAR GCM, is based on convective available potential energy (CAPE). The scheme assumes convective updrafts exist when the atmosphere is conditionally unstable, and CAPE is used up at an exponential rate during parcel ascent (Kiehl et al. 1998a). Zhang (2002) developed the revised closure assumption using the observations from the Atmospheric Radiation Measurement Program (ARM) and TOGA-COARE. New closure relates convection to the destabilization by the large-scale advection processes, which is consistent with the concept of CRMs. Wu et al. (2007) applied the cloud-resolving simulations in deriving the trigger condition for deep convection. The convection is activated when the CAPE increase due to the large-scale advection of temperature and moisture exceeds a certain threshold. Zhang and Wu (2003) used the cloud-resolving simulations to evaluate the CMT parameterization scheme developed by Zhang and Cho (1991) and Wu and Yanai (1994). The
CMT scheme considers the vertical redistribution of the horizontal momentum by convection, and accounts for the role of perturbation pressure field generated by the interaction of convection with large-scale circulation in vertical momentum transport. The interaction between CMT and the thermodynamic effects of convection plays an important role in shaping up tropical convection and the Hadley circulation (e.g., Wu et al. 2003; Song et al. 2008).

The mosaic approach of incorporating subgrid cloud distribution into the radiation calculation was developed by Liang and Wang (1997). Liang and Wu (2005) and Wu and Liang (2005) evaluated the mosaic treatment of cloud distributions against the CRM simulations (Wu and Moncrieff 2001). The approach divides the GCM grid column into subcells so that an individual layer within a subcell is either completely overcast or cloud-free. Each overcast subcell layer contains a specific cloud genus with distinct optical properties. The most important consideration is to distinguish, within individual cloudy subcells, two cloud fractions: one with and another without inherent geometric association. In particular, convective (Cc), anvil cirrus (Ci) and stratiform (Cs) clouds in each layer are defined to be geographically distinct and minimally overlapped; Cc clouds are assigned to a single subcell column, while Ci fill consecutively the subcells that are equally divided over the remaining grid area; Cs is distributed into random-ordered subcells with an identical sequence for adjacent layers (maximal overlap) and otherwise independent sets for random overlap. Use of mixed overlap has shown to produce better agreement in total cloud cover with observations (Tian and Curry 1989). At a given layer, one subcell may contain the residual partial cloud fraction to conserve the grid total cloud amount. Separate ICA (Independent Column Approximation) radiation calculations are then performed for each subcell with clouds, whereas clear sky radiative fluxes are computed only once and used for all subcells when needed. Grid mean radiative heating rates and fluxes are then the area averages over all subcells. Consequently, the mosaic approach can adequately address the cloud macrogrouping (geometric association) and inhomogeneity (within-cloud optical property variance) effects on radiation.

Over the last decade, NASA satellite observations have obtained a suite of invaluable global data including water vapor (NVAP, SSM/I, AIRS), precipitation (TRMM, GPCP, CMAP), cloud liquid water path (SSM/I, AMSR-E, MODIS, ISCCP), cloud ice water path (MODIS, MLS), cloud amount (ISCCP, MODIS, VIRS), ocean evaporation (SSM/I, AMSR), and the top of the atmosphere (TOA) radiative fluxes (ERBE, CERES, ISCCP). Recently,
progress has been made to retrieve ice water path (IWP) in ice-over-water cloud systems by a multilayered cloud retrieval system (MCRS) using combined microwave (MW), visible (VIS) and infrared (IR) satellite measurements (e.g., Lin et al. 1998a, b; Ho et al. 2003; Huang et al. 2005, 2006; Minnis et al. 2007). The development of oceanic climatology (from 1998 to the present) of IWP from satellites Aqua and TRMM data is now quite feasible. The uncertainties of retrievals at night and for precipitating clouds can be further reduced with the aid of the MW precipitation retrievals (Lin and Rossow 1997; Kummerow et al. 2001), the cloud radar on CloudSat (Stephens et al. 2002) and the CALIPSO lidar (Winker et al. 2003). The improved GCM physical parameterization scheme and retrievals of NASA satellite observations provide a unique opportunity to systematically and consistently evaluate all aspects of precipitation, cloud liquid and ice properties and radiative fluxes in the global water and energy cycle, which impose a strong constraint on the formulation of convection, cloud and radiation processes in GCMs.

Our goal is to evaluate the combined effects of previously mentioned modifications in GCM output, specifically the changes made to convection and cloud. Effects of each individual change to the convection scheme have been studied in previous GCM research. Thus, we will assess combined effects of all convection parameterization changes. It is hypothesized that the synthesis of such modifications can provide a constraint on the simulation and produce a mean climate that more resembles observations. We want to understand how such modifications affect the climate simulation. Since the convection and cloud schemes are the only changes from early versions of NCAR GCM, our experimental model should be sufficient in providing evidence of modification impacts. Also, our model is simpler than newer adaptations of NCAR GCM (e.g. clouds are diagnostic), so simulations are less time-consuming and make it easier to distinguish impacts of revisions made to the convection parameterization. This paper is organized in the following way. Section 2 describes the model simulations under study and outlines modifications made to the experimental model. Observations used for comparison and model simulations are also described in Section 2. In Section 3, results are displayed, which include the effects of the combined convection scheme changes. A discussion and concluding remarks follow in Sections 4 and 5.
3. GCM simulations and observational data

a. Control and experimental simulations

In this research work, two atmosphere-only GCMs are under study, the National Center for Atmospheric Research General Circulation Model (NCAR GCM) and the Iowa State University General Circulation Model (ISUGCM). Both simulations are run for the ten-year period 1980-89 and are forced by observed monthly sea surface temperatures. The control model (CTL) is a version of the NCAR GCM, and it is of T42 horizontal resolution—approximately 2.8° x 2.8°. CTL has 18 hybrid sigma levels in the vertical with a model top at 2.9 mb (Kiehl et al. 1998a). A land surface model from Bonan (1998) is incorporated as well. Radiation at the top of the atmosphere in this model is tuned to agree with ERBE fluxes (Kiehl et al. 1998b). Implemented in CTL is a Zhang and McFarlane (1995) deep convection scheme, which is based on local CAPE change. The experimental model (EXP) is the ISUGCM. It is based on NCAR GCM, but modifications have been made to the deep convection scheme. Specifically, convection closure has been revised, so that now destabilization rate of the troposphere by large-scale thermal and moisture advection is now considered (Zhang 2002) and used in place of the previous closure. A trigger threshold based on change in convective available potential energy (CAPE) controls the initiation of deep convection. As in NCAR GCM, the Zhang and McFarlane deep convection scheme is coupled with the Hack (1994) shallow convection scheme. The Hack scheme moistens the atmosphere and preconditions it for deep convection. In EXP, the radiation parameterization known as the mosaic cloud approach (Liang and Wang 1997) has been implemented. Grid points in the GCM are divided into subcells with one cloud genus describing each subcell: convective (Cc), anvil cirrus (Ci), and stratiform (Cs). Inclusion of mosaic clouds accounts for different radiative and optical properties of cloud genera. An average over all subcells theoretically improves the radiation calculations, since various cloud properties are taken into account. Previous climate simulations have shown the Hadley circulation has been modified through addition of CMT and results in a closer to observed eastward propagation of the Madden-Julian Oscillation (MJO) of 5 m s⁻¹ speed (Deng and Wu 2010; Deng and Wu 2011).

b. Observational datasets

Many observation datasets were used in this study, most of which are related to precipitation, clouds, and radiation. If any data were of a different horizontal resolution
compared to the models, regridding through bilinear interpolation was performed before
comparison. An important point to consider is that many observation datasets may be quality-
controlled, but discrepancies still exist. There is much uncertainty with global observations,
because of spatial and temporal sampling issues. Missing data, especially over oceans, add to the
lack in confidence of some observations. Regardless, it is crucial to assess several available
datasets to determine which ones are best for the task at hand.

Precipitation rate (PRECT) was the first and most important variable analyzed in this
research, so a few different datasets were used. For global PRECT data, Global Precipitation
Climatology Project (GPCP) was first used for making comparison between models and
observations. The Xie-Aarkin precipitation data, also known as Climate Prediction Center (CPC)
Merged Analysis of Precipitation (CMAP) data, were also used. These two datasets have several
main differences, such as CMAP having higher PRECT values over tropical ocean but lower
ones over polar oceans. Most differences occur over oceans versus over land, since fewer
observations exist over the oceans. Atoll data incorporated into CMAP may not be ideal either
(Yin et al. 2004). Either way, both datasets were used for model to observation comparison. A
Tropical Rainfall Measuring Mission (TRMM) dataset, similar to that of Lin et al. 2000, was
used for the diurnal analysis of PRECT. TRMM data span 50°N-50°S and were originally of 2.5°
\times 2.5° resolution before regridding. Evaporation rates were also important to analyze, so the 15-
yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-15)
was used for this variable. The National Center for Environmental Prediction (NCEP) NCAR
Reanalysis 1 data were critical in assessing model surface temperature, surface sensible heat
flux, and surface latent heat flux (Kalnay et al. 1996). Assimilation of land surface, buoy,
rawinsonde, satellite, and other data are integrated into these NCEP products (Kalnay et al.
1996). Monthly NCEP data were used in this research. Cloud radiation fields were taken from a
two different datasets: the International Satellite Cloud Climatology Project (ISCCP) and the
Earth Radiation Budget Experiment (ERBE) data. First produced in 1982, ISCCP sought to
produce a climatology of cloud variations (Rossow et al. 1985; Rossow and Schiffer 1991). Low
cloud, medium cloud, and high cloud fractions were all taken from ISCCP for comparison to the
models. ERBE data were collected by NASA satellites during the mid-1980s in an effort to
obtain monthly averages of several top of atmosphere (TOA) fluxes (Barkstrom 1984). Due to its
global domain, ERBE avoids intercalibration uncertainties. Also, ERBE have broadband
measurements and yield TOA information for obtaining reflectance (Breon et al. 1994). In regards to ice and liquid water, a product was relied upon called CloudSat. CloudSat observations were taken over oceans only, because brightness temperatures over land are erroneous (Trenberth et al. 2005). Waliser et al. 2009 emphasized the lack of ice data in both observations and IPCC AR4 models, and they discussed an ice water path dataset called CloudSat. Backscatter is measured from the planet’s surface with the Cloud Profiling Radar (CPR) on the CloudSat satellite. A radar reflectivity factor helps convert backscatter to an appropriate variable for finding IWP. Since the satellite observes a vertical profile, precipitation and cloud information are provided in the atmospheric profile.

4. Results

a. Precipitation characteristics

Since a more distinct signal in temperature and other atmospheric variables exists for June-August (JJA) and December-February (DJF), examination of these two extreme seasons, is usually the focus of past climate studies. Transition seasons, March-May (MAM) and September-October (SON), are often neglected. The mean climate state is not usually as well-portrayed for transition seasons, which may be due to inaccurate tropical heating. As a result, precipitation errors in the Tropics occur (Jia et al. 2012). Shukla et al. 2003 found that GCMs simulate clearer signals in boreal winter, for example, when tropical Pacific sea surface temperature (SST) anomalies related to El Niño-Southern Oscillation (ENSO) are strong. Variability is greater in transition seasons, and thus, there is more noise. Regardless, we want to assess GCM performance during all seasons as errors during intermediate seasons can lead to biases for the winter and summer in many atmospheric fields.

Annual mean plots are first created to identify differences between CTL and EXP. In Fig. 1, we start by displaying precipitation rate (PRECT) between the models in a global mean ANN plot. The upper subplot of the panel shows a mean PRECT that is .40 mm day\(^{-1}\) too high compared to GPCP observations (bottom subplot). EXP is not nearly as high with global ANN PRECT (~.20 mm day\(^{-1}\) too high). Both models struggle with simulating monsoon zone precipitation. In fact, EXP displays a weaker (~2-3 mm day\(^{-1}\) less) global mean value of ANN precipitation over central Africa. A weaker precipitation zone is also shown over North America,
Central America, and South America in EXP. On the other hand, the Southeast Asian Monsoon looks stronger in EXP than CTL. Since the GCMs are of coarse horizontal resolution, CTL and EXP cannot fully resolve precipitation over small islands like the Maritime Continent (Indonesian islands and Philippines). Neither model captures the correct ANN PRECT over the Asian Monsoon, but the structure of precipitation bands does look similar between models and GPCP. In model output and observations, the Inter-Tropical Convergence Zone (ITCZ) has a southern branch that extends into the South Pacific known as the South Pacific Convergence Zone (SPCZ). This prominent feature is important to tropical dynamics and circulation as it produces strong upwelling and advection of cooler water to the west of South America (Kiladis et al. 1989). Therefore, it should be portrayed in the mean precipitation simulations.

Breaking down PRECT into seasonal averages may be the better way to observe structure and progression of rain belts. Figure 2 includes the 1980-89 seasonally-averaged precipitation for CTL. During December-February (DJF), a bulk of high PRECTs are located south of the Equator, although the model does erroneously have the greatest values of 14+ mm day\(^{-1}\) extending a couple degrees north of the Equator. March-May (MAM) is a transition season, in which mean PRECTs pick up, but maxima are not as large as DJF. In June-August (JJA), model convection is most vigorous and some of the higher PRECTs protrude into the Southern Hemisphere, which is not reflective of observations (Figs. 4 and 5). September-November (SON), another transition season, shows diminished mean PRECTs but an ITCZ that has a double-branched shape. Observations do not show a double ITCZ structure.

The EXP simulation is shown in Fig. 3. For each season in EXP’s output, there is lower global mean PRECT than CTL. Peak PRECT stays to the south of the Equator in DJF and to the north of the Equator in JJA. The DJF subplot for EXP clearly shows a SPCZ that is diagonal in structure. Parallel rain belts straddling north and south of the Equator characterize the undesired double ITCZ structure as in the CTL simulations. MAM shows weaker global mean PRECT than CTL but it is ~.04 mm day\(^{-1}\) higher than DJF. Overall EXP PRECT structure, especially in the Tropics, more resembles the observations. A double ITCZ is less prominent for EXP than CTL during SON. Observed PRECT data are plotted in Figs. 4 and 5 (GPCP and Xie-Arkin, respectively). The main difference between observations is that the Xie-Arkin dataset, sometimes referred to as CMAP, has a more pronounced ITCZ, but mean PRECT amounts are similar between the two observation datasets for all but JJA. On average, PRECT in the Tropics is
higher in CMAP, probably because atoll rain gauges are integrated in the dataset and the gauges are uncorrected for wetting evaporation and aerodynamic effects (Gruber et al. 2000). Although this may contribute to most differences between observed datasets, the input of satellite data may also cause some differences.

A Student t-test (assuming independence of monthly means) has been performed to indicate areas which are significantly different between models. In Fig. 6, seasonal PRECT differences are plotted between EXP and CTL. Dashed contours indicate significant differences between models at the 95% significance level. From the panel’s subplots, it is seen that most seasonal differences between model runs occur in the Tropics. Max DJF and JJA differences follow the location of the ITCZ, but the models do have a few variations in the storm tracks (40º-70º). Up to 2 mm day\(^{-1}\) less PRECTs in the southern storm tracks occur in EXP during DJF. Northern storm track differences are less statistically significant in DJF than the southern ones, probably coincident with position of the ITCZ. JJA shows fewer statistically significant PRECT differences over the storm tracks than DJF. MAM shows very few places of significant statistical difference besides along the Equator and southern storm tracks. SON has one main statistically different swath of rain that extends from the eastern Maritime Continent (120ºE) to about 150ºW in the Pacific Ocean. This seems to be an area in which EXP overestimates rainfall along the SPCZ.

To better explain differences in the ANN PRECT between models, zonal average plots help provide latitudinal distributions. Figure 7 shows zonally-averaged PRECT for the two models and two observation datasets over the four seasons. A bimodal peak occurs in PRECT in DJF and JJA. EXP is closer to the CMAP peak in both of these seasons than CTL (blue). An outlier is seen in DJF with CTL as the peak in PRECT is too high compared to both observed datasets. CTL’s peak PRECT during JJA should be south of the Equator, not at its location slightly north of the Equator. Zonally-averaged JJA PRECT for the two models is more similar, but EXP has a stronger peak of ~1 mm day\(^{-1}\) at 8ºN. Outside of the deep Tropics, EXP is slightly closer to both observation datasets during most of JJA. It is interesting to note that during March-May (MAM), a bimodal peak feature shows up again, quite similar to DJF. EXP and CTL are similar in representation of this transition season. During September-November (SON), this transition period more reflects JJA representation of PRECT. In general, transition seasons are
more similar in zonal structure to their preceding season (DJF or JJA) due to seasonal lag in SSTs compared to the solar cycle.

b. Radiation budgets

To prove consistency between atmospheric variables, low values of outgoing longwave radiation (OLR) should match up in model simulations with areas of stronger convection, as OLR has been shown to be negatively correlated to areas of deep convection (Xie and Arkin 1998). Earth Radiation Budget Experiment (ERBE) OLR data from polar-orbiting satellites are used for comparison. Figures 8-11 display OLR for: a) ERBE, b) CTL, and c) EXP. Figure 8 shows the average OLR for DJF and observations in the top subplot indicate that OLR is greatest in areas with few clouds. The poles are the exception though as the high surface albedo (due to ice) reflects back shortwave radiation and thus, there is less longwave radiation absorbed. This results in less longwave radiation that escapes to space at the poles. Wherever the ITCZ is not present, there is more OLR, fewer clouds, and hence, less precipitation. Global DJF average OLR for EXP is 2 W m$^{-2}$ smaller than the observations, whereas CTL is about 4 W m$^{-2}$ greater. Areas of difference between EXP and CTL exist throughout the Tropics, as what may be thicker clouds in EXP decrease OLR. Fig. 9 is of average 1980-89 MAM OLR, and it shows the global average OLR observations stay about the same from DJF to MAM. The JJA OLR (Fig. 10) shows a higher global average than MAM and DJF probably because clouds during JJA are more penetrable by longwave radiation. Strong PRECTs over the Bay of Bengal in the models create a bias in OLR as it is underestimated. EXP does have slightly lower OLR than EXP. OLR off the west coast of South America more resembles ERBE during JJA, but over Australia, EXP underestimates OLR. There are probably more clouds over Australia in EXP. This will be determined in output of related fields. EXP has a better handle on Eastern Pacific lower SON OLR amounts (Fig. 11).

Global averages of TOA shortwave radiation indicate EXP has less shortwave energy at TOA than CTL (which may be in agreement with more ice and hence more reflective clouds in EXP). The TOA shortwave radiation in model output for 1980-89 averaged DJF (Fig. 12) has TOA shortwave radiation that is smaller south of the Equator along the SPCZ. Smaller TOA shortwave radiation values follow the ITCZ, and the pattern is consistent in other seasons.
Surface shortwave radiation is also important in the radiation scheme since solar radiation may produce temperature perturbations at the lowest levels that promote buoyancy. Fig. 13 shows average DJF shortwave radiation at the surface for CTL and EXP along with differences. In most of the deep Tropics, EXP is around -50 W m$^{-2}$ less in terms of surface shortwave radiation than CTL. This agrees with the OLR as there is less OLR in deep convective zones in EXP output. Location of the decreased shortwave radiation also occurs in the location of the ITCZ and hence coincides with the seasonal cycle. In Fig. 14, JJA surface shortwave radiation is displayed, and again, the transitioning of the main ITCZ precipitation branch northward agrees with smaller values in surface shortwave radiation. EXP has less surface shortwave radiation in most convective areas. MAM and SON (not shown) have similar trends in that the smaller values of SFC shortwave radiation follow the ITCZ. Surface (SFC) longwave radiation is another important part of the radiation budget as this component is determined by clouds and water vapor in the first few km (Morcrette 2002). The 1980-89 averaged DJF SFC longwave radiation is shown in Fig. 15 and the effects of EXP modifications are seen—decreased longwave radiation in most of the Tropics and more longwave radiation closer to the poles. With the CMT included in EXP, the main precipitation branch of the ITCZ moves to the appropriate hemisphere and the double ITCZ is alleviated through revised convection closure. The 1980-89 averaged MAM, JJA, and SON plots (not shown) for this variable indicate the same main difference between EXP and CTL.

c. Impacts of clouds

Radiative flux effects at the TOA are influenced by clouds and are best represented by longwave (LWCF) and shortwave cloud forcing (SWCF) (Marat et al. 2005). Cloud radiative forcing (CRF) is defined as the effects of longwave plus shortwave radiation at the top of the atmosphere due to clouds. Longwave radiative cloud forcing (LWCF) may be defined as:

$$LWCF = E_{clr} - E$$  \(1\),

where $E$ is total longwave radiative flux and $E_{clr}$ corresponds to clear sky radiative flux. Shortwave cloud forcing (SWCF) is defined by a second equation:

$$SWCF = S - S_{clr}$$  \(2\),

where $S$ is shortwave total scene reflected flux and $S_{clr}$ is reflected flux for completely clear skies.
The average 1980-89 DJF LWCF is plotted in Fig. 16 for CTL, EXP, and ERBE data. Areas in which clouds have more of a “trapping” effect on radiation have positive LWCF. Regions of deep convection tend to have higher LWCF as the possibly thicker clouds block radiation that would otherwise escape and radiate to space at cooler temperatures. EXP has on average ~5 W m$^{-2}$ greater LWCF than CTL. A maximum LWCF of 100 W m$^{-2}$ is present in the warm pool region in both models and observations, which coincides with the idea that EXP has more optically thick clouds. ERBE data do not amplify LWCF in the warm pool region as much as the models though. For other seasons (not shown), EXP always has greater global average LWCF than CTL. CTL has a closer global mean to that of ERBE data though.

As for the average SWCF, this field shows a better match to ERBE observations in EXP output than that of CTL. SWCF is negative wherever there is abundant cloud cover, because SWCF represents the cooling effects of clouds. The 1980-89 average DJF SWCF (Fig. 17) has more negative values where the ITCZ is located. This is because optical properties, solar irradiance, and amount of cloudiness all impact SWCF. The addition of the cloud mosaic scheme allows output of SWCF to better resemble ERBE. Heterogeneity of cloud features becomes clearer with cloud mosaic, since cloud properties are accounted for in subgrid fashion. The EXP 1980-89 DJF global means and for each other season (MAM, JJA, and SON not shown) are closer to ERBE.

Heating rate profiles, both shortwave and longwave, show differences in model thermal structures and may help explain differences in the overall simulation of convection in each model. More heating at lower levels should allow for increased instability, whereas lower heating rates build up instability more slowly. In this sense, heating rates are connected to surface fluxes. Displayed in Figs. 18 and 19, are the zonally-averaged longwave heating rate (LW HTG) differences between EXP and CTL for DJF and JJA, respectively. A main difference between models happens in the deep southern Tropics near 1,000 hPa in which decreased heating of up to -1 K day$^{-1}$ occurs in EXP compared to CTL during DJF. Transitioning of the ITCZ to the hemisphere experiencing its warm season in EXP seems to be consistent, since this model simulation general shows less precipitation than CTL. Warmer heating rates at lower pressure levels in EXP are located in the opposite hemisphere as the main ITCZ branch. MAM and SON (not shown) show slight shifts toward their extreme season counterpart (JJA and DJF, respectively). The shortwave heating rate (SW HTG) differences are shown in Figs. 20 and 21.
Again, trends in EXP heating are in agreement with ITCZ location. During DJF, the smaller SW HTGs are smaller in EXP than CTL south of the Equator because the main ITCZ cloud bands are located here in EXP. Consistent with previous variables, thicker clouds may be located in the Northern Hemisphere of EXP during boreal summer that block solar radiation (Fig. 21).

Table 1 summarizes the net radiative fluxes at top of atmosphere (TOA) and surface (SFC) for longwave (LW) and shortwave (SW) radiation. Observation values for net radiation in Table 1 are ERBE data. LWP and IWP observations originate from SSM/I and CloudSat datasets, respectively. Inclusion of mosaic cloud in EXP accounts for the subgrid variability in the model by treating the subcells with individual cloud properties. Effects of these properties on radiation are varied, but radiation values are altered once averaged over all subcells. Resulting net radiation values in LW and SW at TOA and SFC are up to ~10 W m\(^{-2}\) greater in EXP than CTL. Still, realistic radiation budgets are maintained with cloud mosaic incorporation. As in Wu and Liang (2005), radiative-effective cloud is reduced, and we find a crucial change in the liquid and ice content of clouds. Arguably the most important impacts of the cloud mosaic scheme are made on LWP and IWP because of the impacts made on radiation. Higher LWPs and IWPs result from the mosaic cloud approach and closer to observed values are obtained, especially for LWP as ice observations contain much uncertainty. Consequently, cloud mosaic affects cloud fractions. Annual mean medium cloud (CLDMED) and high cloud (CLDHGH) fractions decrease in EXP through the inclusion of the mosaic scheme by about 1\% and 3.5\%, respectively. ISCCP mean data for CLDMED and CLDHGH are slightly closer to mean values in EXP, but as stressed numerous times, cloud observations have high uncertainty. CloudSat cloud fractions are higher in their global annual means than ISCCP. CLDHGH is much closer to EXP’s simulated value, which is in agreement with the increased IWP. The annual low cloud fraction (CLDLOW) increases by ~0.5\% in EXP, which is probably due to increased low clouds near the poles. In CloudSat, CLDLOW more resembles EXP, and CLDMED does not have much difference between simulations. Global mean total cloud fraction (CLDTOT) between EXP and CTL has decreased by 1-3\% (not shown). Higher total cloudiness is indicative of more precipitation, so in EXP smaller global mean PRECT is in agreement with lower CLDTOTs for each season. Although seasonal CLDTOT has decreased by only a few percent with the convection scheme modifications, the key differences are broken down into cloud types: low, medium, and high. CLDLOW, CLDMED, and CLDHGH are summarized up by seasonal
differences (EXP-CTL) in Table 2. CLDLOW increases for all seasons by up to 8% in EXP.
Lower clouds are more associated with thunderstorms, so this may be in agreement with the idea of less often but more vigorous convection in EXP than CTL. Middle level cloud increases by ~2% in EXP for all seasons, so less sunlight is absorbed at the surface in EXP. This is probably another reason why instability must takes longer to build up before deep convection starts.

CLDHGH is quite different between models, and each season shows a decrease of as much as 14% in EXP to CTL comparisons. Since higher clouds contain more ice, this has quite the impact on radiation. Fewer high clouds means more incoming radiation to penetrate lower levels and less overall cloudiness as seen by the decrease in all seasons for EXP’s CLDTOT of 1-3%.

Lowered CLDTOTs in EXP imply lower PRECTs, which is shown in the last line of Table 1. Higher mean global CLDTOT in CTL is primarily due to the tuning of cloud water to balance radiation in CTL. Another possibility though is that cloud fractions are altered by detrainment of water by the revised convection parameterization. In EXP, the mosaic radiation scheme allows for clouds to have various optical properties and more resemble observations.

As stated earlier, assessment of liquid and ice water is critical in understanding the new mean climate produced by EXP. The total liquid water path (LWP) in EXP is greater than CTL over almost the entire globe for each season. The 1980-89 JJA average LWP is displayed in Fig. 22, and SSM/I observations are used as comparison for LWP output. Maximum LWPs occur in the midlatitudes and low latitudes over areas of frequent convection. Both models show this, but EXP has on average over twice as high LWPs as CTL. CTL may have a closer global mean LWP to SSM/I, but the models simulate LWP over land, whereas observations do not. EXP shows the much more consistent pattern of LWP over convective areas. Clouds are much more abundant over the ITCZ and storm tracks in EXP output. This is also true for 1980-89 averaged DJF, MAM, and SON (not shown).

Since clouds are more optically thick in EXP simulations, total ice water path (IWP) is also on average 23 g m\(^{-2}\) greater in EXP. In Fig. 23, average boreal summer (JJA) IWP is plotted. Both models have maxima in IWP over the SE Asian Monsoon region, an area known for deep convection. Other seasons show a shift in the maxima with the transition of the ITCZ. On average, CloudSat observations in Fig. 23c look better represented by EXP than CTL. A caveat to the CloudSat observations though is that they are averaged over the months spanning July 2006 – June 2010. IWP data are extremely limited, so this set of observations is better than none
at all. CTL is about three times too low in terms of average IWP, whereas EXP is ~10 g m\(^{-2}\) higher than the observations. With EXP having increased LWP, IWP, and hence thicker clouds, this may improve cloud forcing No observations are available for comparison, but both models have maxima in IWP over the SE Asian Monsoon region, an area known for deep convection. Other seasons show a shift in the IWP maxima with the transition of the ITCZ. As a result, there has been an overestimate in the SWCF in CTL simulations (Norris and Weaver 2001).

d. Diurnal cycle of precipitation

Day-to-day fluctuations of PRECT are important to assess, because they are components of the mean climate state. Since most GCMs precipitate too early and too frequently, it is of interest to inspect the two models for differences. The observation dataset used for comparison in the diurnal cycle inspection is a version of the Tropical Rainfall Measuring Mission (TRMM) similar to that used by Lin et al. (2000). For direct comparison, observations were again gridded to T42 resolution. PRECT frequency was found by taking all hours and counting the number of times PRECT was above a nonzero threshold of .01 mm day\(^{-1}\). For the models, this was done over the years 1980-89, but TRMM data were of years 1998-2005. Although the years are different, similar trends between data and models should exist.

Figure 24 shows PRECT frequency for three different sources: TRMM (green), CTL (blue), and EXP (red). Solid lines denote frequencies over land, whereas dashed lines indicate values over oceans. TRMM peak PRECT frequency over land occurs at 16 local solar time (LST) with max frequencies of close to 30%. Over oceans, DJF TRMM PRECT frequencies max out around 25% between 0-4 LST. Precipitation for CTL happens too often at frequencies between 25-45% over land and 50-60% over oceans. The peak PRECT frequency occurs a few hours too early in CTL over land but is about right over oceans. As for DJF EXP PRECT frequencies, they are about 20% less over oceans in EXP and between 15-20% less over land than CTL. EXP’s land curve is less jagged than the TRMM curve, but EXP’s general trend more resembles TRMM. The dashed red EXP curve (for over oceans) shows the same trend as CTL except that the EXP’s curve is shifted down about 20%. This shift is probably due to the revised convection trigger in the deep convection scheme. In EXP, CAPE builds up and when a change in CAPE of at least 65 J kg\(^{-1}\) h\(^{-1}\) occurs, deep convection initiates. This trigger holds back the stronger convection and since oceans have higher heat capacity than land, instability from
heating takes longer at the surface. Since EXP and CTL are uncoupled atmosphere-only models, this is yet another reason for little change in shape of the PRECT frequency curves over oceans. During the other extreme season, JJA, PRECT frequencies (Fig. 25) look similar to DJF. Overall JJA frequencies for EXP are again less than CTL. Timing of the peak PRECT occurs in the mid-afternoon for the models, but an abrupt drop-off in frequency is seen in each simulation, whereas TRMM PRECT frequencies do not decrease as fast. Although neither model quite resembles TRMM’s diurnal cycle, EXP is much closer to the TRMM curve.

Timing of max PRECT frequency is shown on a global contour plot in Fig. 26 for DJF. The three-panel plot has TRMM data, CTL, and EXP. TRMM data show early morning PRECT frequency peaks over open oceans in the early morning hours during DJF (Fig. 26a). Over land, timing of max PRECT frequency occurs closer to the mid-afternoon as indicated by the lighter yellow and orange colors. Islands tend to have PRECT frequencies maximize around late evening to early morning (e.g. over the Indonesian Islands). Land-ocean interaction may also alter the diurnal cycle in special cases when stronger low-level winds provide moisture farther inland. This can somewhat be seen in the area just east of the U.S. Rocky Mountains but is much more prevalent in boreal summer. In Figs. 26b and c, both models struggle to depict Northern Hemisphere winter max timing of PRECT frequency. Since the convection trigger in each model is based on CAPE, it is much more difficult to simulate the cold season when less instability is present. Austral summer shows more frequent convection around 1 LST in EXP than CTL, especially over South Africa and central South America. This may be due to jet-like features around these locations that influence the moisture content needed for thunderstorms. It is difficult to determine this in the TRMM data for West Africa though because of missing data. The JJA output for timing of max PRECT frequencies (Fig. 27) shows a more sound Northern Hemisphere simulation in EXP than CTL over most land areas. For example, east of the Rockies, EXP seems to pick up on the early morning convection, which is mostly attributed to nocturnal low-level jet rain events during summer. West Africa also looks more accurately resembled by EXP. The exception though is India, in which most convection takes place in early morning in EXP but during mid-afternoon in CTL. Tropical Southern Hemisphere PRECT frequencies tend to occur a couple hours later in EXP than CTL. Similar to DJF, the models mostly precipitate in early mornings over the oceans. Until interactive ocean is included, this should not change much between simulations.
Station plots created for ten different locations during the warm season are plotted in the Fig. 28 panel. Similar to stations used by Dai et al. (2007), model PRECT frequencies are shown. TRMM observations are not available at some locations, so we turn to Dai et al. (2007) when comparison data are needed. The 10 stations for comparison during the warm season include: a) S.E. U.S., b) Western Equatorial Pacific, c) Europe, d) the Equatorial Indian Ocean, e) Eastern China, f) Sahel, g) the North Pacific, h) South America, i) the North Atlantic, and j) S.E. Australia. PRECT frequency for CTL (blue) and EXP (red) illustrate several different points. First, in warm season locations, PRECT peak frequencies in EXP tend to be lower than that of CTL (Figs. 28e, f, h, and j). This is consistent with the idea that the convection trigger condition prevents convection from occurring as frequently. The peak frequency of PRECT for E. China is the only diurnal land plot in which EXP has a slightly higher frequency than CTL. The diurnal cycle peak PRECTs over oceans show slightly higher peak frequencies of PRECT than CTL (Figs. 28b, d, g, and i). Compared to TRMM (not shown), models still tend to precipitate too frequently. Peak timing of PRECT frequency looks to have shifted 1-2 hours later in EXP than CTL.

Another diurnal aspect of precipitation is that of intensity. This variable is found by counting all time steps in which precipitation occurred and dividing PRECT by the number of events during that specific time step (e.g. each hour). To check the phase PRECT intensity, or timing of max PRECT intensity, averages over each LST hour are found. The max intensities are plotted for DJF and JJA (Figs. 29 and 30). In the TRMM subplot for DJF, max PRECT intensities occur mostly in mid to late afternoon over land and in the early morning over oceans (similar to that of max PRECT frequency timing). Oceans show little variability in the models compared to TRMM. The TRMM data were originally of higher resolution, so after regridding occurred some of the islands’ timings of max PRECT are a combination of land/ocean effects. Neither model depicts max timing of PRECT intensities that well for DJF, but EXP does produce most intense convection earlier in the mornings (~6 LST), whereas CTL has its peak intensities around 8 LST. TRMM data indicate 11 LST as the average time of PRECT max intensity though, so this season is not well-portrayed by either model. During JJA (Fig. 30), timing of max PRECT intensity is clearly different between models. TRMM data display a mean max timing of intensity at 12 LST, whereas CTL and EXP are several hours earlier on average. CTL again has more variable timing of maximums in the Northern Hemisphere oceans than EXP. EXP though
does attempt to resolve inland locations influenced by ocean moisture though, such as east of the U.S. Rockies. A dipole exists there in which richer Gulf moisture provides later morning intense thunderstorms. The average timing of max PRECT intensity in EXP is three hours closer to that of TRMM’s average during JJA.

e. Surface heat fluxes

Sensible heat flux (SHFLX) and latent heat flux (LHFLX) may have implications for the simulations of PRECT. Chao (2012) found that excessive SHFLX and/or lack of ventilation of boundary layer heat along steep slopes may cause inaccurate PRECT over rough orography. This may be the case for erroneous regions of PRECT over mountainous areas in GCM simulations. Surface SHFLX values are displayed in Fig. 31. Similar to Fig. 1, model and observations are shown in the panel plot. A major problem area with the models is that of the Tibetan Plateau. SHFLX in the area has a weaker dipole in model simulations over the Tibetan Plateau. A stronger dipole exists in the NCEP NCAR Reanalysis-1 data for SHFLX. As for the hypothesis that there may be stronger SHFLX over higher elevations in the models, this does not seem to be the case. On average, CTL has twice as large SHFLX as NCEP data, whereas EXP values are about 1.5 times as large. In Fig. 32, seasonal SHFLX differences between EXP and CTL are plotted similar to PRECT in Fig. 6. The general trend between models shows that EXP decreases SHFLX over oceans by about -10 W m$^{-2}$. Highly convective areas, termed “convective hot spots,” have increases in SHFLX of up to 60 W m$^{-2}$ in EXP.

Since it is difficult to see some of the finer differences between models, zonal average plots were created of SHFLX in Fig. 33. For DJF, EXP is closer to the NCEP data at nearly every point. According to observations, SHFLX is negative at higher latitudes and positive in midlatitudes. A main two-hump feature is present with one hump in each hemisphere for each season. SHFLX is close to zero at the Equator. The main regions of bias between models and observations are the polar areas. Near 80ºN, DJF zonally-averaged SHFLX is ~30 W m$^{-2}$ too high in EXP but CTL is almost twice as high as NCEP. The MAM and JJA model simulations both lack in representation of polar regions. Near 90ºS, the largest bias between models and observations occurs. Neither of the two models simulates values of SHFLX less than -30 W m$^{-2}$, which is quite biased from NCEP values of near -60 W m$^{-2}$ at this extreme latitude. At 90ºN, MAM and JJA are in better agreement with NCEP, but a bias up to 25 W m$^{-2}$ still exists.
Problematic regions during SON are again limited to the higher latitudes. Since both models are uncoupled atmosphere-only, they probably struggle with SHFLX values in extreme latitudes without interactive ocean and ice. Besides, the main effects of EXP modifications tend to alter the tropical regions.

Surface latent heat flux (LHFLX), another important parameter, is the difference between the two lowest moisture layers of the atmosphere. LHFLX is released when convection occurs and it is a means of atmospheric stabilization. Values of LHFLX vary around the world, but higher values of this variable may indicate more active storm regions. Like the previous zonally-averaged plot, a similar one is displayed in Fig. 34 for LHFLX. It seems that for each season, CTL tends to have higher LHFLX values than EXP in the Tropics. This coincides with the fact that each season has shown to have a higher global mean PRECT for CTL than EXP. During DJF, three peaks in LHFLX show up for both models and NCEP. Peaks of LHFLX on either side of the Equator represent areas in which most latent heat has been released to stabilize the atmosphere and likely are areas of max PRECT. It was thought that areas of max PRECT would match up with those of max LHFLX, but looking at DJF, this does not seem to always be the case, especially in the Tropics. Looking at Fig. 7 (Xie-Arkin observations) again, max PRECT is located at 8°N during JJA, but for Fig. 34, max LHFLX is near 15°S, but there is a secondary peak near 8°N. A difference in peak PRECT and LHFLX may occur because the tropical atmosphere is already more moist in the deep Tropics. Therefore, areas in which there is a drier atmosphere and intense convection occurs, higher LHFLX values exist. More variability exists in the JJA subplot for LHFLX between the model runs, especially in EXP. This may be due to inclusion of convective momentum transport in EXP, which modifies the tropical circulation through enhancement of the Hadley circulation.

5. Discussion

Combined effects of convection parameterization changes in EXP may actually be summed up through use of a Taylor diagram. A Taylor diagram helps illustrate differences between EXP and CTL and whether amplitude variation and pattern variation match better with observations. Fig. 35 displays 1980-89 DJF global means for CTL (blue) and EXP (red). Some new variables are introduced in this Taylor diagram such as surface temperature (T) and
evaporation minus precipitation (E-P). Although the new variables in the Taylor diagram are not highlighted in earlier analysis, they reinforce what earlier research has already shown.

First, PRECT is closer to representing amplitude variation (x and y-axes) and centered (global mean removed) pattern correlation (curved axis). This reinforces what was shown earlier in Figs. 2-4. EXP, with its modifications, shows a global DJF mean PRECT closer to observations than CTL. Surface temperature (T) has not changed much in the Taylor diagram between model simulations. It is simply plotted to instill the point that some mean atmospheric variables do not change much even with the modifications made to the convection scheme. E-P (not shown as a horizontal plot) has a global mean close to balance in both models for each season. On the Taylor diagram for DJF, it is interesting that EXP actually shows a lower root mean square (RMS) difference (distance from reference point) for E-P than CTL, and variation amplitude is slightly closer to ECMWF in EXP. SHFLX and LHFLX are closer to the reference point for EXP. The CLDTOT fractions in EXP have better pattern correlations to ISCCP by about +0.1 compared to CTL. CLDHGH in CTL looks to be the only cloud fraction variable that is closer to the reference point. EXP tends to have fewer high clouds than CTL. The other extreme season, JJA, is shown in Fig. 36 in Taylor diagram form. This time, the pattern correlation and RMS difference for model PRECT and GPCP is nearly the same in CTL and EXP. T is slightly off in terms of amplitude variation in EXP. E-P, LHFLX, and SHFLX in EXP output better match observations and reanalysis though. Cloud fractions are similar between models and observations, but CLDHGH again has a lower RMS difference with ISCCP than EXP.

Transition seasons are not displayed in Taylor diagrams, but similar results are obtained in which radiative surface fluxes better match observations in amplitude variation and pattern correlation for EXP outputs. CLDHGH better matches the variation in amplitude and pattern correlation for CTL. Table 3 highlights the tropical annual means and tropical means for eight variables: precipitation rate, evaporation rate, surface temperature, surface sensible heat flux, surface latent heat flux, and cloud fractions for low, middle, and high clouds. Precipitation rate is still too high in EXP but it is closer to GPCP observations than CTL. Evaporation in the Tropics is also closer in EXP than CTL to the ECMWF data. Surface temperature is slightly higher in EXP than CTL. Surface sensible heat and latent heat fluxes are closer to NCEP observations than CTL. A primary difference between models is in cloud output, in which EXP simulates a higher
low cloud and middle cloud fraction but lower high cloud fraction than CTL. All cloud fractions are closer in EXP to observations than CTL for the Tropics.

From the horizontal plots shown earlier of PRECT, the deep convection changes made to EXP seem to increase the low and medium cloud and thus the LWCF. The CTL LWCF is more closely matched to observations than EXP, but not by much. SWCF though is different between models. EXP increases the SWCF for each season by $\sim33\%$. This increases cooling, especially in the Tropics. Less solar forcing reaches the surface. The deep convection trigger suppresses convection until the required CAPE change threshold is reached.

Diurnal precipitation fields analyzed earlier revealed EXP to have closer mean frequency of PRECT over oceans and land when compared to TRMM most of the time. On average, EXP’s timing of the max frequency events occurred a couple hours later than CTL. Timing of max PRECT intensities for EXP were earlier in the morning than CTL, but not necessarily closer to the observations. The change in convection closure and trigger in EXP lead to later convection, primarily during the warm season. This in turn, modifies mean climate fields related to the convection, especially PRECT and radiative fluxes. These fields are then simulated closer to observations with the aforementioned modifications.

6. Conclusions

To realize the impacts of convection scheme changes, this study compared two different model simulations. The control model, NCAR GCM (CTL), was found to have typical problems that occur with atmosphere-only GCMs, such as a double ITCZ structure, poor resemblance of transition season climate fields, and exaggerated precipitation in monsoon regions and areas of steep mountains. An altered convection scheme in the experimental model, ISUGCM (EXP), allowed for investigation of the combined parameterization changes to be studied. Comparison between CTL and EXP aided in the understanding of how the modified convection scheme affected certain mean climate fields.

Student t-tests showed most of the significant differences between the two models in mean climate fields occur in the Tropics. Generation of more low and medium but fewer high cloud fractions in EXP and the cloud mosaic scheme influence the radiative and optical properties in the model atmosphere. A caveat to this though is that EXP produces more optically thick clouds (higher ice and liquid content). The main influence of clouds on radiative forcing is
broken down into LWCF and SWCF. When compared to observations, CTL showed a better agreement with LWCF by ~5 W m$^{-2}$. EXP actually had higher global mean LWCF for each season, and since LWCF describes the trapping of heat by clouds, the experimental model lessens the OLR. LW HTG is increased in the lowest atmospheric levels in areas of thick cloud cover. On the other hand, the EXP SWCF more closely matched with observational data by ~10 W m$^{-2}$. Since SWCF describes the overall cooling effect of clouds, it becomes clear that since SWCF is larger than the LWCF, SWCF dominates in EXP. This means less insolation reaches the surface in EXP during the warm season. SW HTG rates agreed with this statement. Hence, less instability should be able to build up as rapidly from solar forcing. Thermal and moisture advection are tied in with destabilization of the troposphere and when there is ample instability, the deep convection trigger activates. More vigorous, but less frequent precipitation may occur.

Diurnal analysis of PRECT frequencies and intensities showed this to be mostly true.

As for shortcomings in AGCMs mentioned earlier, EXP shows a reduction in the double ITCZ problem especially during DJF and SON. CTL has the two-branch parallel structure in precipitation that characterizes the double ITCZ. Global mean precipitation fields in EXP output are closer to observations. The transition of the ITCZ is more clearly represented in EXP as the maximum in precipitation during DJF and JJA occurs at ~8°S, and ~8°N, respectively. Convective momentum transport may be responsible for the resultant transition of the ITCZ in EXP. Although these facets of the model look improved, monsoon regions are still exaggerated. This may be more of a resolution problem though, since the GCM grids are so coarse. Monsoon features are still one of the most difficult features to correct simulate in GCMs and without the addition of an ocean model, EXP is inept at obtaining correct PRECT amounts.

Ongoing research will further analyze the frequency and vigor of the precipitation generated in EXP. The diurnal cycle may need to be more fully inspected to determine if the change in convection parameterization is correctly working on small temporal scales. To determine the circulation differences between models, a global water vapor budget should be assessed for each season. A water vapor budget also provides a view of various modes of transport, whether they be the total, standing, or transient components and their impacts on circulations. An extension of the research in this paper will eventually involve coupling the atmosphere and ocean. These components will add to the working knowledge of the modified convection scheme.
7. Acknowledgements

We would like to thank Daryl Herzmann for much of the computing aid required to complete this research. Also, Dave Flory was helpful in providing computer storage for mass amounts of data. This research was partly supported by the Biological and Environmental Research Program (BER), U.S. Department of Energy under Grant DE-FG02-08ER64559, and by the National Science Foundation under Grant ATM-0935263.

8. References


Chao, W. C., 2012: Correction of excessive precipitation over steep and high mountains in a GCM. J. Atmos. Sci., 69, 1547-1561.


9. List of Tables and Figures

Table 1. Annual 10-year 1980-89 mean global averages of net longwave (LW) and shortwave (SW) radiative fluxes (W m\(^{-2}\)) at top of atmosphere (TOA) and surface (SFC) from control (CTL), experimental (EXP) model, and observations (OBS). Radiation observations are ERBE data. Liquid water path (LWP) observations (g m\(^{-2}\)) are from SSM/I and ice water path (IWP) (g m\(^{-2}\)) are from CloudSat. Cloud fractions (%) are from ISCCP and CloudSat. Precipitation rates (mm day\(^{-1}\)) are from CMAP and GPCP.

Table 2. Global 1980-89 mean cloud fraction differences (in percent), EXP – CTL for four seasons. CLDxxx represents cloud fractions, where xxx is replaced by LOW, MED, HGH, and TOT stand for low, medium, high, and total cloud, respectively.

Table 3. Tropical (30°N-30°S) annual 1980-89 means for some model atmospheric variables and their respective observed values with units. P denotes precipitation rate (mm day\(^{-1}\)), E is evaporation rate (mm day\(^{-1}\)), TS is surface temperature (K), SH is surface sensible heat flux (W m\(^{-2}\)), LH is surface latent heat flux (W m\(^{-2}\)), and CLDxxx are cloud fractions (%) for low, middle, and high clouds.

Fig. 1. 1980-89 annual (ANN) mean precipitation rate (mm day\(^{-1}\)). Top and middle panels are CTL and EXP, respectively. GPCP data are in bottom panel regridded to T42 grid.

Fig. 2. 1980-89 mean CTL precipitation rate (mm day\(^{-1}\)) for the four seasons and global means.

Fig. 3. Same as Fig. 2 except for EXP.

Fig. 4. Same as Fig. 3 except for GPCP.

Fig. 5. Same as Fig. 4 except for Xie-Aarkin data.

Fig. 6. Average precipitation differences, 1980-89 EXP – CTL. Dashed contours represent differences that are significant at the 95% significance level (SL).

Fig. 7. Zonally-averaged precipitation rate over four seasons for CTL (blue), EXP (red), CMAP (long, black dash) and GPCP (short, black dash).

Fig. 8. Average outgoing longwave radiation during 1980-89 DJF for: a) ERBE, b) CTL, and c) EXP. All color-filled contours are of units W m\(^{-2}\).

Fig. 9. Same as Fig. 9 except for 1980-89 MAM.
Fig. 10. Same as Fig. 9 except for 1980-89 JJA.
Fig. 11. Same as Fig. 10 except for 1980-89 SON.
Fig. 12. Average 1980-89 DJF TOA shortwave radiation for a) CTL, b) EXP, and c) EXP – CTL. Dashed line in c) indicates the 95% level of significance.
Fig. 13. Average 1980-89 DJF SFC shortwave radiation for a) CTL, b) EXP, and c) EXP – CTL. Dashed line in c) indicates the 95% level of significance.
Fig. 14. Same as Fig. 13 except for 1980-89 JJA.
Fig. 15. Same as Fig. 13 except for surface longwave radiation.
Fig. 16. 1980-89 average DJF LWCF for: a) CTL, b) EXP, and c) ERBE data. Units are W m\(^{-2}\).
Fig. 17. Same as Fig. 16 except for SWCF.

Fig. 18. Mean 1980-89 DJF longwave heating rate for EXP – CTL. Units are K day\(^{-1}\).
Fig. 19. Same as Fig. 18 except for 1980-89 JJA.
Fig. 20. Same as Fig. 19 except for shortwave heating rate.
Fig. 21. Same as Fig. 20 except for 1980-89 JJA.

Fig. 22. Total liquid water path (g m\(^{-2}\)) for averaged 1980-89 JJA for: a) CTL, b) EXP, and c) SSM/I.

Fig. 23 Average 1980-89 JJA total ice water path for: a) CTL, b) EXP, and c) CloudSat. Units are g m\(^{-2}\). CloudSat observations are difficult to obtain and the ones used in c) are actually for July 2006 – June 2010 from Waliser et al. 2009.

Fig. 24. Mean DJF precipitation frequency versus local solar time. Solid lines indicate frequencies over land and dashed lines are for frequencies over oceans. Different colors are for various data: green (TRMM), blue (CTL), and red (EXP). TRMM data are for years 1998-2005 and model simulations are for 1980-89.
Fig. 25. Same as Fig. 24 except for JJA.
Fig. 27. Same as Fig. 26 except for JJA.
Fig. 28. Mean precipitation frequencies for 10 stations during their warm seasons: a) SE US, b) Western Equatorial Pacific, c) Europe, d) Equatorial Indian Ocean, e) E. China, f) Sahel, g) Northern Pacific, h) S. America, i) N. Atlantic, and j) SE Australia during the warm season. The x-axis displays LST hour and y-axis is precipitation frequency (%).

Fig. 29. Same as Fig. 26 except for timing of max precipitation intensity.

Fig. 30. Same as Fig. 29 except for JJA.

Fig. 31. Same as Fig. 1 except for sensible heat flux (SHFLX). NCEP data are in bottom panel regridded to T42 grid. Units are W m$^{-2}$.

Fig. 32. Same as Fig. 6 except for surface sensible heat flux differences. Units are W m$^{-2}$.

Fig. 33. Same as Fig. 7 except for SHFLX and black line represents NCEP observations. Units are W m$^{-2}$.

Fig. 34. Same as Fig. 33 except for LHFLX. NCEP observations are in black.

Fig. 35. Taylor diagram of average 1980-89 DJF of various atmospheric variables for CTL (blue) and EXP (red).

Fig. 36. Same as Fig. 35 except for 1980-89 average JJA.
Table 1. Annual 10-year 1980-89 mean global averages of net longwave (LW) and shortwave (SW) radiative fluxes (W m\(^{-2}\)) at top of atmosphere (TOA) and surface (SFC) from control (CTL), experimental (EXP) model, and observations (OBS). Radiation observations are ERBE data. Liquid water path (LWP) observations (g m\(^{-2}\)) are from SSM/I and ice water path (IWP) (g m\(^{-2}\)) are from CloudSat. Cloud fractions (%) are from ISCCP and CloudSat. Precipitation rates (mm day\(^{-1}\)) are from CMAP and GPCP.

<table>
<thead>
<tr>
<th>Global</th>
<th>CTL ANN</th>
<th>EXP ANN</th>
<th>OBS ANN</th>
</tr>
</thead>
<tbody>
<tr>
<td>(F_{\text{LW}}) (TOA)</td>
<td>238.7</td>
<td>233.5</td>
<td>233.9</td>
</tr>
<tr>
<td>(F_{\text{SW}}) (TOA)</td>
<td>240.6</td>
<td>236.4</td>
<td>234.0</td>
</tr>
<tr>
<td>(F_{\text{LW}}) (SFC)</td>
<td>60.8</td>
<td>64.6</td>
<td>49.4</td>
</tr>
<tr>
<td>(F_{\text{SW}}) (SFC)</td>
<td>171.5</td>
<td>165.9</td>
<td>165.9</td>
</tr>
<tr>
<td>(LWP_{\text{SSM/I}})</td>
<td>53.5</td>
<td>119.0</td>
<td>97.4</td>
</tr>
<tr>
<td>(IWP_{\text{CloudSat}})</td>
<td>10.6</td>
<td>33.4</td>
<td>21.9</td>
</tr>
<tr>
<td>(\text{CLDLOW}_{\text{CloudSat}/\text{ISCCP}})</td>
<td>39.4</td>
<td>40.0</td>
<td>48.8 (24.9)</td>
</tr>
<tr>
<td>(\text{CLDMED}_{\text{CloudSat}/\text{ISCCP}})</td>
<td>29.8</td>
<td>28.2</td>
<td>36.4 (23.1)</td>
</tr>
<tr>
<td>(\text{CLDHGH}_{\text{CloudSat}/\text{ISCCP}})</td>
<td>33.8</td>
<td>30.3</td>
<td>37.3 (13.0)</td>
</tr>
<tr>
<td>(\text{PRECT}_{\text{CMAP}/\text{GPCP}})</td>
<td>3.13</td>
<td>2.87</td>
<td>2.69 (2.61)</td>
</tr>
</tbody>
</table>
Table 2. Global 1980-89 mean cloud fraction differences (in percent), EXP – CTL for four seasons. CLDxxx represents cloud fractions, where xxx is replaced by LOW, MED, HGH, and TOT stand for low, medium, high, and total cloud, respectively.

<table>
<thead>
<tr>
<th>EXP–CTL</th>
<th>DJF</th>
<th>MAM</th>
<th>JJA</th>
<th>SON</th>
</tr>
</thead>
<tbody>
<tr>
<td>CLDLOW</td>
<td>+8%</td>
<td>+7%</td>
<td>+6%</td>
<td>+7%</td>
</tr>
<tr>
<td>CLDMED</td>
<td>+3%</td>
<td>+2%</td>
<td>+2%</td>
<td>+2%</td>
</tr>
<tr>
<td>CLDHGH</td>
<td>−12%</td>
<td>−14%</td>
<td>−10%</td>
<td>−11%</td>
</tr>
<tr>
<td>CLDTOT</td>
<td>−2%</td>
<td>−3%</td>
<td>−1%</td>
<td>−1%</td>
</tr>
</tbody>
</table>
Table 3. Tropical (30°N-30°S) annual 1980-89 means for some atmospheric variables and their respective observed values with units. P denotes precipitation rate (mm day$^{-1}$), E is evaporation rate (mm day$^{-1}$), TS is surface temperature (K), SH is surface sensible heat flux (W m$^{-2}$), LH is surface latent heat flux (W m$^{-2}$), and CLDxxx are cloud fractions (%) for low, middle, and high clouds.

<table>
<thead>
<tr>
<th>Tropics</th>
<th>CTL ANN</th>
<th>EXP ANN</th>
<th>OBS ANN</th>
</tr>
</thead>
<tbody>
<tr>
<td>P_GPCP</td>
<td>3.87</td>
<td>3.51</td>
<td>2.91</td>
</tr>
<tr>
<td>E_ECMWF</td>
<td>4.30</td>
<td>4.00</td>
<td>4.06</td>
</tr>
<tr>
<td>TS_NCEP</td>
<td>298.19</td>
<td>298.49</td>
<td>298.28</td>
</tr>
<tr>
<td>SH_NCEP</td>
<td>24.28</td>
<td>22.45</td>
<td>22.59</td>
</tr>
<tr>
<td>LH_NCEP</td>
<td>124.95</td>
<td>116.29</td>
<td>110.13</td>
</tr>
<tr>
<td>CLDLOW_ISCCP</td>
<td>25.0</td>
<td>32.0</td>
<td>28.0</td>
</tr>
<tr>
<td>CLDMED_ISCCP</td>
<td>7.0</td>
<td>9.0</td>
<td>14.0</td>
</tr>
<tr>
<td>CLDHGH_ISCCP</td>
<td>40.0</td>
<td>28.0</td>
<td>13.0</td>
</tr>
</tbody>
</table>
Fig. 1. 1980-89 annual (ANN) mean precipitation rate (mm day$^{-1}$). Top and middle panels are CTL and EXP, respectively. GPCP data are in bottom panel regridded to T42 grid.
Fig. 2. 1980-89 CTL precipitation rate (mm day$^{-1}$) for the four seasons and global means.
Fig. 3. Same as Fig. 2 except for EXP.
Fig. 4. Same as Fig. 3 except for GPCP.
Fig. 5. Same as Fig. 4 except for Xie-Arkin data.
Fig. 6. Average precipitation differences, 1980-89 EXP – CTL. Dashed contours represent differences that are significant at the 95% significance level (SL).
Fig. 7. Zonally-averaged precipitation rate over four seasons for CTL (blue), EXP (red), CMAP (long, black dash) and GPCP (short, black dash).
Fig. 8. Average outgoing longwave radiation during 1980-89 DJF for: a) ERBE, b) CTL, and c) EXP. All color-filled contours are of units W m$^{-2}$.
Fig. 9. Same as Fig. 8 except for 1980-89 MAM.
Fig. 10. Same as Fig. 9 except for 1980-89 JJA.
Fig. 11. Same as Fig. 10 except for 1980-89 SON.
Fig. 12. Average 1980-89 DJF TOA shortwave radiation for a) CTL, b) EXP, and c) EXP – CTL. Dashed line in c) indicates the 95% level of significance.
Fig. 13. Average 1980-89 DJF SFC shortwave radiation for a) CTL, b) EXP, and c) EXP – CTL. Dashed line in c) indicates the 95% level of significance.
Fig. 14. Same as Fig. 13 except for 1980-89 JJA.
Fig. 15. Same as Fig. 13 except for surface longwave radiation.
Fig. 16. 1980-89 average DJF LWCF for: a) CTL, b) EXP, and c) ERBE data. Units are W m$^{-2}$. 
Fig. 17. Same as Fig. 16 except for SWCF.
Fig. 18. Mean 1980-89 DJF longwave heating rate for EXP – CTL. Units are K day$^{-1}$. 
Fig. 19. Same as Fig. 18 except for 1980-89 JJA.
Fig. 20. Same as Fig. 18 except for shortwave heating rate.
Fig. 21. Same as Fig. 20 except for shortwave heating rate.
Fig. 22. Total liquid water path (g m$^{-2}$) for averaged 1980-89 JJA for: a) CTL, b) EXP, and c) SSM/I.
Fig. 23. Average 1980-89 JJA total ice water path for: a) CTL, b) EXP, and c) CloudSat. Units are g m\(^{-2}\). CloudSat observations are difficult to obtain and the ones used in c) are actually for July 2006 – June 2010 from Waliser et al. 2009.
Fig. 24. Mean DJF precipitation frequency versus local solar time. Solid lines indicate frequencies over land and dashed lines are for frequencies over oceans. Different colors are for various data: green (TRMM), blue (CTL), and red (EXP). TRMM data are for years 1998-2005 and model simulations are for 1980-89.
Fig. 25. Same as Fig. 24 except for JJA.
Fig. 26. Timing of max precipitation frequency for DJF. Panel subplots are for: a) TRMM (1998-2005), b) CTL (1980-89), and c) EXP (1980-89). Time is local solar time.
Fig. 27. Same as Fig. 26 except for JJA.
Fig. 28. Mean precipitation frequencies for 10 stations during their warm seasons: a) SE US, b) Western Equatorial Pacific, c) Europe, d) Equatorial Indian Ocean, e) E. China, f) Sahel, g) Northern Pacific, h) S. America, i) N. Atlantic, and j) SE Australia during the warm season. The x-axis displays LST hour and y-axis is precipitation frequency (%).
Fig. 29. Same as Fig. 26 except for timing of max precipitation intensity.
Fig. 30. Same as Fig. 29 except for JJA.
Fig. 31. Same as Fig. 1 except for sensible heat flux (SHFLX). NCEP data are in bottom panel regridded to T42 grid. Units are W m$^{-2}$. 
Surface Sensible Heat Flux: EXP – CTL, 1980–89

Fig. 32. Same as Fig. 6 except for surface sensible heat flux differences. Units are W m$^{-2}$.
Fig. 33. Same as Fig. 7 except for SHFLX and black line represents NCEP observations. Units are W m$^{-2}$. 
Fig. 34. Same as Fig. 33 except for LHFLX. NCEP observations are in black.
Fig. 35. Taylor diagram of average 1980-89 DJF of various atmospheric variables for CTL (blue) and EXP (red).
Fig. 36. Same as Fig. 35 except for 1980-89 average JJA.
CHAPTER 4. CONVECTION SCHEME IMPACTS ON WATER VAPOR FLUX DURING NORTHERN SUMMER IN GCM SIMULATIONS

A paper to be submitted to *Journal of Climate*

Zachary A. Mangin, Xiaoqing Wu, and Tsing-Chang Chen

1. Abstract

Two general circulation model simulations were used to test impacts of convection parameterization on the June-August mean hydrological climate state. A decomposition of the water vapor budget into stationary and transient components was performed as in Chen (1985). Similar results were found with the models under study compared to observations from Chen (1985). The stationary nondivergent mode of $\Psi_Q$ explains most of the global water vapor flux in both the National Center for Atmospheric Research General Circulation Model (CTL) and the Iowa State University General Circulation Model (EXP). Overall magnitude and structure of the field was similar to NCEP-2 Reanalysis data. The stationary $\chi_Q$ field, which explains the divergent mode was also simulated in both models, but EXP again showed better strength and overall areal extent. This field explains the maintenance of water vapor by the Walker Circulation. Transient modes ($\Psi_Q'$ and $\chi_Q'$) are indicative of cyclones in the climate system. Both models simulated the transient zonal gradient along the storm track regions of each hemisphere. This is important because storm systems traverse this zone and bring synoptic weather patterns eastward. Water vapor transport moves poleward from the tropics for the transient mode. The convective scheme does not change overall structure of stationary and transient components of the water budget for $\Psi_Q$ and $\chi_Q$, but there are subtle differences. EXP displayed stronger gradients in the stationary components. Stronger circulations of up to $\sim$15 kg s$^{-1}$ for $\Psi_Q$ and $\sim$5 kg s$^{-1}$ for $\chi_Q$. These differences coincide with the more vigorous convection that occurs in EXP. As for transient fields, EXP again showed slightly tighter gradients than CTL, and thus, cyclone systems are less prevalent but more vigorous in the EXP due to convection being tied to large-scale advections. These general conclusions between model simulations help reinforce previous findings with the modified convection scheme.
2. Introduction

Clouds are linked with radiation and hydrological processes through solar heating, which drives surface evaporation and adds water vapor to the atmosphere. Global circulations must produce sufficient water vapor for clouds to produce precipitation though. Past hydrological research has focused on water vapor transport, which was initiated by Starr and Saltzmann (1966) during the planetary circulation project. Rasmusson (1972), Salstein et al. (1980), Salstein et al. (1983), and others continued water vapor flux studies through use of streamfunction and velocity potential. Streamfunction and potential functions allowed for the breakdown of circulations into rotational and divergent components. Some studies were simply completed to obtain a visual distribution of water vapor content (Chen 1985), whereas others sought to explain interannual differences in water vapor (Rosen et al. 1979; Berbery and Collini 2000; Long et al. 2000). Chen (1985) used similar tools to describe the maintenance of water vapor flux during two extreme seasons. In the study, water vapor flux was broken down into stationary (long-term seasonal mean) and transient (daily) components. It was found that the nondivergent component of water vapor flux explains most of the total water vapor transport. The divergent mode maintains high water vapor content over equatorial regions. Poleward transport is generated by transient components (cyclone systems). It is important that these circulation features show up because the Hadley and Walker circulations facilitate meridional and zonal extent of water vapor. Chen (1985) utilized streamfunction of water vapor flux ($\Psi_Q$) and velocity potential of water vapor flux ($X_Q$) to illustrate observational water vapor transport and maintenance.

No known general circulation model (GCM) studies have analyzed the water vapor flux in the same manner as Chen (1985). By breaking down water vapor into stationary and transient terms as well as nondivergent and divergent variables, one can determine the effectiveness of the GCM to simulate water vapor transport and associated circulations. Another aspect of GCM simulations that should be examined is that of the hydrological diurnal cycle. It is the diurnal cycle, or day-to-day variations in atmospheric variables, that helps determine microphysical and parameterization effects. Parameterization impacts are most easily seen and evaluated on shorter temporal scales (Randall et al. 1991). With regards to the diurnal hydrological cycle, analysis of precipitation has been performed in several previous studies. Yang and Slingo (2004) used Cloud Archive User Service (CLAUS) brightness temperature data from the International Satellite
Cloud Climatology Project (ISCCP) to examine diurnal variation of tropical convection. Coincident with previous research, diurnal amplitude of brightness temperature was most pronounced over clear-sky land. As a response to solar heating, a peak in convection occurred in the late afternoon over land. Over the oceans, convection was most frequent in the early morning—corresponding to infrared cooling that led to destabilization (Randall et. al 1991). Liu and Moncrieff (1998) reemphasized this idea through use of a cloud-resolving model (CRM) simulation that showed a similar result. Land-ocean areas, such as the Maritime Continent (15°N-25°S, 90°E-170°E), are varied in regards to diurnal amplitude and timing of precipitation. Greatest precipitation usually occurs over the islands, but there is a secondary peak over the nearby oceans, likely due to mesoscale sea breeze interaction.

Similar to uncoupled GCM studies, Dai and Trenberth (2004) found summer rainfall to occur prematurely by a few hours in the coupled GCM. Convective versus nonconvective ratios of precipitation were too high as well. Later, Dai (2006) performed an intercomparison of eighteen coupled models. Models that were not flux-corrected displayed spurious precipitation as a feature known as the “double Intertropical Convergence Zone (ITCZ).” The double ITCZ may be due to warm bias in sea surface temperatures or simply the convection parameterization. Song and Zhang (2009) demonstrated the ability of a GCM with revised deep convection scheme to simulate a more accurate ITCZ. Lastly, most models still produced too little stratiform precipitation, too early timing in the max peak of precipitation, and too frequent rainfall at lower than observed intensities.

Computer-demanding models that explicitly simulate clouds are known as cloud-resolving models (CRMs). Experiments involving these models have aided in convection, radiation, and cloud scheme improvements. Given large-scale conditions, CRM techniques may be applied to GCMs, because direct implementation of CRMs into GCM subcells is draining on computer memory. Improved cloud geometry and radiative properties result through use of experimental large-scale conditions from the Global Atmospheric Research Program Atlantic Tropical Experiment (GATE) and the Tropical Ocean Global Atmosphere-Coupled Ocean Atmosphere Response Experiment (TOGA-COARE) (e.g., Grabowski et al. 1996, 1998, 1999; Wu et al. 1998, 1999; Wu and Moncrieff 2001). From such GCM experiments, a revised closure assumption, a trigger condition, convective momentum transport (CMT), and mosaic treatment of subgrid cloud distribution (Liang and Wu 2005) have resulted. The closure assumption in the
The original Zhang and McFarlane (1995) scheme, currently used in most versions of NCAR GCM, is based on convective available potential energy (CAPE). A revised closure assumption using the observations from the Atmospheric Radiation Measurement Program (ARM) and TOGA-COARE was developed by Zhang (2002). New closure relates convection to the destabilization by the large-scale advection processes. Wu et al. (2007) applied the cloud-resolving simulations in the development of a trigger condition that initiates deep convection when the CAPE increase due to the large-scale advections exceeds a certain threshold. The CMT scheme developed by Zhang and Cho (1991a, b) and Wu and Yanai (1994) was evaluated in CRM simulations by Zhang and Wu (2003). The CMT scheme considers the vertical redistribution of the horizontal momentum by convection, and accounts for the role of perturbation pressure field generated by the interaction of convection with large-scale circulation in vertical momentum transport. The interaction between CMT and the thermodynamic effects of convection plays an important role in shaping up tropical convection and the Hadley circulation (e.g., Wu et al. 2003; Song et al. 2008).

Through use of a GCM that has modified convection based on CRM studies, we want to analyze the global water vapor budget in terms of streamfunction and velocity potential during June-August (JJA), the peak of the Southeast Asian Monsoon (SAM). It is believed that combined convection scheme modifications should lead to stronger velocity potentials and possibly stronger streamfunction values in mean climate that more resemble mean observations. The SAM is extremely important to mean climate since it plays a major role in distribution of differential heating and potential energy and the eventual transformation into kinetic energy (Krishnamurti et al. 1998). Much of the planet’s latent heat release originates from the tropical atmosphere and the SAM, characterized by strong atmospheric heating and high precipitation rates (PRECTs) (e.g. Riehl and Malkus 1958; Riehl and Simpson, 1979; Zuluaga et al. 2010). With CMT included in the modified GCM, better seasonal propagation of water vapor circulations should be present (Wu et al. 2007). Also decreased eddy activity should likely result in the experimental model since it is thought to suppress convection for longer periods. Since no major modifications have been made from early versions of the National Center for Atmospheric Research General Circulation Model (NCAR GCM), our simulations should reveal key circulation differences that result from parameterization changes. This paper is organized in the following way. Section 2 contains information regarding model simulations and outlines changes
made to the experimental model. Section 3 describes detailed methods behind the water vapor flux decomposition. In Section 4, results of the water vapor flux decompositions with model simulations are discussed. Sections 5 and 6 include summary and concluding remarks.

3. GCM simulations and observational data

a. Control and experimental simulations

The two GCMs used in this research include a version of the National Center for Atmospheric Research General Circulation Model (CTL) and the Iowa State University General Circulation Model (EXP). EXP is a modified version of the NCAR GCM. Both models are run for ten years (forced by monthly sea surface temperatures) and simulate years 1980-89. Alterations made to EXP include: a revised convection closure assumption based on observations from the Atmospheric Radiation Measurement Program (ARM) and Tropical Ocean and Global Atmosphere Coupled Ocean-Atmosphere Response Experiment (TOGA-COARE), a trigger condition based on cloud-resolving (CRM) model simulations, convective momentum transport (CMT) that accounts for interaction between convection and the large-scale, a mosaic radiation approach that incorporates cloud properties, and an improved liquid water path calculation.

Revised closure in EXP relates deep convection to the destabilization of the tropospheric layer by large-scale advection of temperature and moisture (Zhang 2002). A trigger condition in EXP initiates deep convection when CAPE change of at least 65 J kg\(^{-1}\) h\(^{-1}\) takes place. CMT is included in the climate simulation and it aids in the simulation of the equatorial regions through reduction of heating and drying to limit convection (Wu et al. 2007). Mosaic radiation approach is integrated in the model to allow for subgrid scale variability in cloud optical property. GCM grid points are split into subcells, and each subcell is defined by one of the following cloud genera: convective (Cc), anvil cirrus (Ci), and stratiform (Cs). The genera have their own cloud optical properties and when averaging occurs over all subcells, cloud macrogrouping and inhomogeneity are represented. Impact of this mosaic approach on radiation has shown warmer tropospheric temperatures (>1°C) in low and mid-latitudes. Cloud amount and precipitation are reduced in the tropics due to stabilization from a dipole structure of cooling near the surface and warming at upper levels (Liang and Wang 1997). Modified liquid water path calculations in EXP use a vertical scaling factor, \(f(z)\), to constrain model radiation budgets closer to observations.
With the new scaling factor, water decreases with height more slowly and more cloud ice water is located at upper levels. These findings came from CRM simulations with ARM cloud systems.

b. Data

The main datasets used in this research include the National Center for Environmental Prediction Department of Energy (NCEP-DOE) II Reanalysis (R-2) observations (Kanamitsu et al. 2002). A caveat exists with these data though. The R-2 observations do not output specific humidity, but relative humidity and temperature at various pressure levels made it possible to find a value for specific humidity through use of the Clausius-Clapeyron equation. To find specific humidity \( q \), we needed to input pressure, relative humidity, and temperature into a version of Clausius-Clapeyron. R-2 data have been corrected for several errors related to sea level pressure, snow cover, ocean albedo, relative humidity, and snowmelt. Data are of the same spatial (T62) and temporal (6-hourly) resolution as Reanalysis-1 data (Kanamitsu et al. 2002). NCEP R-2 were regridded to T42 resolution thought to match model data.

4. Methodology

To test CTL and EXP’s handles on maintenance of water vapor during years 1980-89, several steps were taken. Investigation into differences in the water budget between models became of interest because there were no known studies found that performed similar research. Also, mean climate states of hydrological variables in both models are different and the cumulus parameterization has impact. Daily means were first found for \( \Psi_Q \) and \( \chi_Q \). The streamfunction describes the rotational (nondivergent) component circulation, whereas the potential field represents the divergent circulation. Subscript “Q” refers to water vapor. So, to obtain either value, water vapor fluxes had to first be calculated. Since transition seasons March-May (MAM) and September-November (SON) show less distinct circulations we wished to exclude these from current water budget analysis. Since a preliminary study has been done on June-August (JJA), we perform our analysis on this season.

Following Chen (1985), water vapor transport is represented by the following equation, where \( v \) is total wind (\( \nu = u \hat{t} + v \hat{j} \)),

\[
\bar{Q} = \frac{1}{g} \int_{P_u}^{P_h} \bar{q}\bar{v}dp = \frac{1}{g} \int_{P_u}^{P_h} \bar{q}\bar{v}dp + \frac{1}{g} \int_{P_u}^{P_h} \bar{q}'\bar{v}'dp = Q + Q'.
\]
An overbar signifies a seasonal mean (stationary) component and a transient is represented by an apostrophe ('). The $g$ stands for gravity and $q$ is specific humidity. Limits for the integrations are 300 hPa for $P_u$ and $P_L$ is equal to surface pressure. The upper limit is set to 300 hPa because essentially all water vapor exists below that pressure level. Equation (1) was then used to calculate zonal and meridional components of $Q$ which are $Q_\lambda$ and $Q_\phi$, respectively. Using Helmholtz theorem, Chen (1985) partitioned water vapor transport into rotational and irrotational pieces. Equation (2) shows this decomposition below:

$$
\tilde{Q} = \hat{k} \times \nabla \Psi + \nabla X = \tilde{Q}_\Psi + \tilde{Q}_X = (\hat{k} \times \nabla \Psi + \nabla X) + (\hat{k} \times \nabla \Psi' + \nabla X') = Q + Q'.
$$

The approach to plot water vapor transport was obtained from Rosen et al. (1979) and Salstein et al. (1980). Fields for these circulations in terms of streamfunction and velocity potential become:

$$\nabla^2 \Psi = \hat{k} \cdot \nabla \times Q, \text{and } \nabla^2 X = \nabla \cdot Q.\quad (3)$$

Water vapor transport from (1) was calculated by finding daily means over all 92 days during June-August (JJA). Eddies were found for $u$, $v$, and $q$ by subtracting the seasonal mean from each day. Zonal and meridional water vapor fluxes were found for stationary and mean modes. The mass-weighted integrations were performed. Stationary and transient water vapor transports were obtained and inputted into the streamfunction and velocity potential calculations on a Gaussian grid. This same procedure was then repeated for 6-hourly data. Computations were performed for CTL, EXP, and observations.

5. Results

a. Vertically-integrated moisture

Previous work with CTL and EXP has helped us better understand mean climate field differences between the model simulations. Specifically, we have found a more accurate structure of tropical rainbands across the Tropics. The South Pacific Convergence Zone (SPCZ) more correctly resembles Xie-Arkin and the Climate Prediction Center’s Global Precipitation Climatology Project (GPCP) observations by slanting southeast rather than being parallel to the Equator (not shown). Seasonal migration of the Inter-Tropical Convergence Zone (ITCZ) has shown to take place with EXP through inclusion of CMT. Figure 1 displays the differences of precipitable water (PREH2O) between EXP and CTL for the four seasons. In EXP, higher PREH2O values in excess of 12 kg m$^{-2}$ compared to CTL cross the Equator during the corresponding warm season. Assuming independence of monthly means, a Student t-test has
been performed to indicate the areas of PREH2O that are significantly different between the models. Dashed contours in Fig. 1 show regions with significant differences between models at the 95% significance level (SL). As already stated, the most significant differences between EXP and CTL occur in the Tropics in the location of the warm season ITCZ rain. Past research with EXP has shown exaggerated precipitation still occurs in the regions of monsoon activity, especially during JJA. The SAM area is no exception, especially during JJA. CTL also overestimates PRECTs though. To break down circulation differences between model simulations, we apply a water vapor decomposition analysis.

**b. Stationary modes: $\Psi_Q$ and $\chi_Q$**

As introduced earlier, the streamfunction and velocity potential are used to describe two different atmospheric circulations, the rotational component and the divergent one. Turbulent moisture fluxes are found and then integrated from the surface to 300 hPa to obtain water vapor transport and the streamfunction (velocity potential) of this water vapor transport is called $\Psi_Q$ ($\chi_Q$). The total fields of $\Psi_Q$ and $\chi_Q$ are broken down into long-term seasonal mean (stationary) and eddy (transient) components. Stationary fields are represented by a tilde (¨) and transient by an apostrophe (‘). The breakdown of these variables is as follows:

$$\Psi_Q = \bar{\Psi}_Q + \Psi'_Q \text{ and } \chi_Q = \bar{\chi}_Q + \chi'_Q$$ (4)

The stationary component of the rotational wind, $\bar{\Psi}_Q$, is shown in Fig. 2 for both models and NCEP R-2 data. Observations (Fig. 2c) display a Northern Hemisphere three-cell structure over the Pacific, Atlantic, and Asian monsoon zones. These anticyclonic centers, responsible for eastward and westward propagation in the tropics, are also identified in Chen (1985). These centers show up in both models (Figs. 2a and b) Compared to the First Global GARP Experiment (FGGE) data used in Chen (1985), the $\bar{\Psi}_Q$ in NCEP R-2 is up to 10 kg s$^{-1}$ less in magnitude where circulation centers exist. It has been nearly 30 years since FGGE data were synthesized, so NCEP R-2 is likely the more error-free dataset. Over Asia, a monsoon trough extends through the Indian subcontinent. This climatological feature is essential to agronomy in the area because much rain falls along and just south of this trough (Annamalai 2010). Both CTL and EXP show the monsoon trough in Fig. 2, but the trough is more defined in ISUGCM. This trough is in fact stronger too by up to ~14 kg s$^{-1}$ in EXP. In CTL, the monsoon trough expands a few degrees too far west into northeast Africa. The southernmost three-cell structure located at 20ºS is more
properly resembled by EXP than CTL. A break in zonal extent occurs at 70°E in this three-cell structure in the EXP output, whereas CTL shows a more continuous feature. Near the 60°S, the closed circulation is ~16 kg s\(^{-1}\) too high compared to NCEP R-2. Most contours are more tightly packed in EXP than CTL, thus indicating stronger rotational winds in EXP than CTL. As found by Chen and Wiin-Nielsen, (1976), the nondivergent circulation describes most of the large-scale water vapor transport and intensity. Hence, the cells mentioned in Fig. 2 provide the needed advection of water vapor to various locations.

The streamfunction explains water vapor transport in the Tropics by the nondivergent component but water vapor maintenance is determined by the divergent circulation (Chen 1985). Velocity potential of water vapor, or \(\chi_Q\), is split up into stationary and transient components. Figure 3 shows \(\bar{\chi}_Q\) for observations and the two models. During JJA, three convergent areas of water vapor occur in NCEP R-2 data (Fig. 3c) over three monsoon zones: East Asia, South America, and Africa. These divergent circulations match up with the Hadley (Walker) circulation in the meridional (zonal) direction. \(\bar{\chi}_Q\) is of smaller magnitude than \(\bar{\Psi}_Q\), but during northern summer \(\bar{\chi}_Q\) is enhanced in magnitude compared to winter (not shown). In EXP, stronger convergence occurs over each monsoon region than in CTL. Revised convection closure in EXP ties destabilization of the troposphere to the large-scale advections and with the new trigger condition, suppression of deep convection and stronger moisture convergence occurs. The convergent high center over SE Asia is closer in intensity to that of NCEP R-2. Also, the location of this closed center is more appropriately located in EXP compared to CTL. This is such important, because the SE Asian monsoon is the most powerful source of latent heating in the planetary general circulation. Areal extent of each tropical circulation more resembles the NCEP R-2 data in EXP, too. Eddy modes should also be explored though.

c. Perturbations: \(\Psi'_Q\) and \(\chi'_Q\)

Transient modes describe atmospheric perturbations, and therefore in a hydrological sense the weather systems that redistribute water vapor around the world. \(\Psi'_Q\) and \(X'_Q\) are shown in Figs. 4 and 5, respectively. From the NCEP R-2 subplots in each figure, it is clear that transient terms are much smaller in magnitude than the stationary ones. This makes sense since the transient is equal to the total field minus the stationary. A zonal structure of \(\Psi'_Q\) and \(X'_Q\) is present in the storm track regions (between 40°-70°), whereas the Tropics show a more cellular
structure. Cyclones traverse the storm track regions and transport water vapor zonally. CTL and EXP both simulate the tight gradient in $\Psi^\prime_Q$ but there are visible differences between simulations outside of the Tropics. Over North America, a very distinct trough shows up in CTL (Fig. 4a), whereas the trough is farther north in EXP. EXP’s depiction of this trough is more comparable in strength to NCEP R-2 (Fig. 4c) though than CTL. This trough setup may be a major reason CTL produces more weather (more precipitation events) than EXP in parts of the midlatitudes. The gradient in the transient components in the higher and lower latitudes indicates water vapor upkeep is directed from the Tropics to the more extreme latitudes. Opposite to $\Psi^\prime_Q$, the transient water vapor flux of velocity potential ($\chi^\prime_Q$) is greater in magnitude than that of the streamfunction. The divergent circulation has greater influence on moisture supply for day-to-day weather systems. Maximum $\chi^\prime_Q$ presides in the extreme latitudes and is almost twice the magnitude than the corresponding $\Psi^\prime_Q$ field. Since the models are forced with monthly sea surface temperatures, transient cells in the Tropics are only somewhat varied. EXP displays more accurate placement of the closed divergent center over Southeast Asia (Fig. 5c). Storm track regions display a stronger gradient in EXP than CTL, so poleward water vapor maintenance is stronger in EXP. This is consistent with higher PREH2O (Fig. 1) in EXP than CTL in most locations. Eddy components provide more moisture to the atmosphere in EXP than CTL.

6. Conclusions

A seldom-used diagnostic method, decomposition of the mean June-August hydrological general circulation into nondivergent and divergent circulations, was used in this study to display differences between two models, one with several convection scheme modifications (EXP), and the other without (CTL). Our previous research had found that the revised convection closure from Zhang (2002) which is tied to large-scale processes helps achieve a more realistic mean climate state in cloud, radiation, and precipitation fields for EXP. The addition of CMT enhances the Hadley and Walker circulation. Inclusion of mosaic cloud in the radiation scheme produces much more realistic liquid and ice content while maintaining acceptable radiation budget between the top of the atmosphere and the surface. We wanted to determine more about hydrological transport and maintenance between these GCM simulations though, and $\Psi_Q$ and $\chi_Q$ provided that. By breaking down these variables into their stationary and transient components,
we were able to learn more about the water vapor budget between models and NCEP R-2 observations.

The stationary field of the nondivergent circulation, \( \Psi_Q \), shows the three main anticyclonic tropical centers in NCEP R-2 and model simulations, but a more distinct difference between simulations is that of the Asian monsoon trough. EXP has a stronger and more distinguishable Asian monsoon trough that better resembles observations. As for the stationary divergent mode (\( \chi_Q \)), we found more accurate depiction in EXP than CTL when compared to NCEP R-2 over most monsoon regions. Since the SE Asian monsoon provides most of the atmospheric energetics, it is extremely important to simulate accurately and EXP is found to be closer in strength and structure in terms of \( \chi_Q \) than CTL. Transient modes were also investigated in this study, and both \( \Psi'_Q \) and \( \chi'_Q \) are much smaller than their respective stationary mode counterparts. Both models correctly portray a steep gradient in eddy fields at more extreme latitudes, which is indicative of water vapor transport poleward. Analysis of these aforementioned fields have given us some more insight into EXP’s mean climate state and especially its more vigorous hydrological cycle. The extreme importance in the large-scale advections is reinforced through this water flux budget. Since EXP’s convection closure connects destabilization of the tropospheric layer to thermal and moisture advections, closer to observed water vapor flux budget is achieved.

Ongoing research will further analyze the water vapor flux budget in all seasons in the GCM simulations. A diurnal cycle approach to \( \Psi_Q \) and \( \chi_Q \) may aid in further diagnosing precipitation differences. An extension of the research in this paper will eventually involve coupling the atmosphere and ocean. The ocean component will probably have a profound effect on the global water vapor flux. These components will add even more to the working knowledge of the modified convection scheme.

7. Acknowledgements:

We would like to thank Daryl Herzmann for much of the computing aid required to complete this research. Also, Dave Flory was helpful in providing computer storage for mass amounts of data. This research was partly supported by the Biological and Environmental
Research Program (BER), U.S. Department of Energy under Grant DE-FG02-08ER64559, and by the National Science Foundation under Grant ATM-0935263.

8. References


### 9. List of Tables and Figures

Fig. 1. 1980-89 mean differences of precipitable water for the four seasons: EXP - CTL. Dashed lines indicate significant differences to the 95% significance level (SL). Units of precipitable water are kg m$^{-2}$.

Fig. 2. Mean 1980-89 June-August $\tilde{\Psi}_Q$ for: a) CTL, b) EXP, and c) NCEP R-2. Units are x $10^7$ kg s$^{-1}$.

Fig. 3. Same as Fig. 2 except for $\tilde{\chi}_Q$.

Fig. 4. Same as Fig. 2 except for $\Psi'_Q$.

Fig. 5. Same as Fig. 4 except for $\chi'_Q$. 
Fig. 1. 1980-89 mean differences of precipitable water for the four seasons: EXP - CTL. Dashed lines indicate significant differences to the 95% significance level (SL). Units of precipitable water are kg m$^{-2}$. 
Fig. 2. Mean 1980-89 June-August $\bar{\Psi}_Q$ for: a) CTL, b) EXP, and c) NCEP R-2. Units are $10^7$ kg s$^{-1}$. 
Fig. 3. Same as Fig. 2 except for $\tilde{\chi}_Q$. 
Fig. 4. Same as Fig. 2 except for $\Psi'_Q$. 
Fig. 5. Same as Fig. 4 except for $\chi^\lambda$. 
CHAPTER 5. GENERAL CONCLUSIONS

1. Mean climate differences

Through use of the Iowa State University General Circulation Model (EXP) and its modified convection scheme, several impacts were noticed on mean climate. First and foremost, global mean precipitation fields were reduced significantly, especially in the Tropics. In EXP, tropical rain belts migrated across the Equator during the respective warm seasons, whereas the control model (CTL) kept main precipitation bands north of the Equator. Revised closure in EXP relates the instability build-up to the large scale temperature and moisture advections. The new convection trigger based on cloud-resolving model (CRM) experiments seemed to suppress convection longer. When convection did take place though, it tended to be more vigorous in EXP. Implementation of mosaic cloud accounted for cloud optical properties in GCM subcells (subgrid variability) and brought liquid and ice water content to more respectable levels, although high uncertainty in these fields is present. Total cloud fractions were slightly reduced in EXP, but the more optically thick clouds in EXP provided higher magnitude cloud forcing. As for diurnal analysis of precipitation, EXP’s modifications lowered precipitation rates and intensities. The timing of the precipitation peaks more closely resembled observations (TRMM) over most land and oceans. Plenty of uncertainties exist in convection and clouds processes, but our method of using CRM techniques seems to have improved the mean climate state in several atmospheric fields.

2. Water vapor flux differences

The diagnostic method involving $\Psi_Q$ and $\chi_Q$, first implemented by Chen (1985), was used to further analyze 1980-89 June-August mean simulations of CTL and EXP. Decomposition of the wind into rotational (nondivergent) and divergent components was required, and this breakdown aided in visualization of circulations relevant to hydrological transport and maintenance. Through streamfunction and velocity potential, these fields revealed some key differences between model simulations in terms of the water vapor flux. The stationary modes ($\bar{\Psi}_Q$ and $\bar{\chi}_Q$) showed most of the general features compared to NCEP Reanalysis-2 data. However, EXP was shown to have a more distinguishable Asian monsoon trough, a very
important feature in regards to latent heat and energetics. Stronger convergence also occurred in the SE Asian monsoon in EXP and may be a result of revised convection closure. In regards to model atmospheric vertical structure, EXP was found to be a much moister model than CTL through comparison of precipitable water. This is in agreement with more optically thick cloud and as a result, convection that occurs tends to be stronger. Transient fields of nondivergent and divergent winds ($\Psi_Q'$ and $\chi_Q'$) show the importance of cyclone systems. Although much smaller in magnitude than their stationary mode counterparts, $\Psi_Q'$ and $\chi_Q'$ are simulated reasonably well in both GCM simulations. Tight gradients in these eddy fields indicate the transport of water vapor poleward. Overall water vapor circulation fields in EXP look somewhat improved through our implementation of CRM-based modifications, but the other seasons should be analyzed as well.

3. References


Chao, W. C., 2012: Correction of excessive precipitation over steep and high mountains in a GCM. *J. Atmos. Sci.*, 69, 1547–1561.


4. Acknowledgements

First and foremost, I would like to thank Prof. Xiaoqing Wu for his guidance, patience, and support through the research and writing of this thesis. I also greatly appreciate Prof. Tsing-Chang (Mike) Chen for teaching me new diagnostic methods. Finally, I would like to thank my other thesis committee members for their valuable comments and suggestions to this work: Prof. Ray Arritt and Prof. William Gutowski. Research completed for this thesis was supported by the National Science Foundation under Grant ATM-0935263.