Early Summer Rains and Their Mid-Summer Cessation Over the Caribbean: Dynamics, Predictability, and Applications.

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EARLY SUMMER RAINS AND THEIR MID-SUMMER CESSATION OVER THE CARIBBEAN: DYNAMICS, PREDICTABILITY, AND APPLICATIONS

By
Theodore Allen
A DISSERTATION

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

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EARLY SUMMER RAINS AND THEIR MID-SUMMER CESSION OVER THE CARIBBEAN: DYNAMICS, PREDICTABILITY, AND APPLICATIONS

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The annual rainfall cycle in the Caribbean is characterized by a bimodal pattern with peaks in the late spring (“early rainfall season”) and late summer (“late rainfall season”) with a mid-summer minimum (“mid-summer drought”). The time average rainfall pattern during the early rainfall season reveals a distinct southwest to northeast spatial pattern, known as the Caribbean rain-belt, that is similar to other northern hemisphere subtropical rain-belts. A series of Caribbean farmer interviews guided my decision to focus on the dynamics and evolution of the Caribbean rain-belt. Results from farmer interviews reveal that their livelihoods are more vulnerable to variability in the timing and amount of the early season rains rather than variability in the mid-summer drying. Therefore, there is a strong social and economic relevance to understand rainfall dynamics during the Caribbean early rainfall season.

The atmospheric dynamics that contribute to the Caribbean rain-belt are diagnosed from the quasi-geostrophic omega equation from daily observations. Forcing for ascent at the upper troposphere is supported by positive zonal wind at 200hPa and jet streaks, while positive temperature advection from the tropics at 500hPa provides forcing for ascent in the mid-troposphere. Moisture availability for the Caribbean rain-belt is regulated by local sea surface temperature and by moisture advection from the tropics in
the lower troposphere. The forcing for ascent weakens throughout the Caribbean and strengthens in the North Atlantic during the mid-summer drought period. Therefore, the mid-summer drought may be diagnosed in terms of weakened uplift dynamics.

A liner inverse model (LIM) is built from OLR, u200, and u850 using 15 degree longitude space channels rather than the traditional reduced EOF space. A physical space based channel LIM to predict tropically propagating OLR anomalies shows positive OLR hindcast prediction skill relative to climatology up to 3 weeks lead time. Confidence in the construction and sub-seasonal prediction skill of the physical based channel LIM allows for the construction of a similar LIM to predict the time evolution of the dynamical parameters that shape the Caribbean rain-belt. The Caribbean LIM is trained between 1982-2010 from an anomalous state vector of ingredients identified from the Caribbean rain-belt diagnostic section and includes the daily anomaly time series of SST, OLR, u200, u850, and v200. SST anomalies have the highest prediction skill over a 3 week lead time, while the positive prediction skill of OLR is muted after 5 days. Prediction skill for the wind channels of the Caribbean LIM remains positive up to one week lead time. The simultaneous and time lagged covariance statistics produce a dynamical operator matrix that describes the internal dynamics between each parameter and identifies SST anomalies to exert the most influence on OLR anomalies. In addition, SST anomalies have the highest signal to noise ratio of each Caribbean LIM state vector channel. Caribbean LIM prediction skill is dependent on how well the state vector represents the linear dynamical system diagnosed from Caribbean rain-belt conceptual model. Low LIM prediction skill may be due to underestimating the non-linear system forcing or from omission of relevant parameters in the state vector.
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Chapter 1: Motivation

Economic prosperity in some industries can be linked to a predictable climate. Evidence of this connection can be seen in the Caribbean where a large percentage of the population is employed within the agriculture sector. However, the percent of GDP from agriculture in the Caribbean is much lower compared to the percent of workforce employed in agriculture (Table 1). The high ratio between agricultural employment force and agricultural productivity creates a volatile condition where slight changes in socioeconomic factors such as farming can have significant economic impacts. One of the many physical constraints that may shape an agricultural economy is its vulnerability to changes in expected rainfall. A stable rainfall pattern is highly predictable and lends itself towards sustainable agricultural productivity. Rainfall variability on the other hand, is less stable and can cause shock to expected crop yields. The related shock can then affect the economic prosperity and overall livelihood of the farming population. Understanding the dynamics that lead to Caribbean rainfall can lead to improvements its prediction. Successful predictions of anomalous rainfall can then support related applications (agriculture, disease, economic, etc.) to offset vulnerability to rainfall variability. However, successful modeling of anomalous rainfall in the Caribbean results only after a complete understanding of the region’s dynamical processes have been established.
<table>
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<th>COUNTRY</th>
<th>% of total population employed in agriculture</th>
<th>% of national GDP from agriculture</th>
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<tr>
<td>Jamaica</td>
<td>17</td>
<td>6.4</td>
</tr>
<tr>
<td>Dominica</td>
<td>40</td>
<td>13.6</td>
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<tr>
<td>Haiti</td>
<td>38.1</td>
<td>24.7</td>
</tr>
<tr>
<td>Cuba</td>
<td>19.7</td>
<td>3.8</td>
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Table 1. The percentage of people employed in agriculture and the total economic output from agriculture for a specific country.

Late spring rainfall in the Caribbean is important because it supplies moisture to nourish the early summer “secondary” crops. Income generated from secondary crops can finance the materials needed to capitalize on the more productive late summer “primary” growing season. The focus of this dissertation is on the atmospheric dynamics connected to late spring rainfall.
Chapter 2. THE ONSET OF EARLY SEASON RAINFALL AND ITS MIDSUMMER CESSIONATION IN THE CARIBBEAN: A NEW PERSPECTIVE OF THE MID-SUMMER DROUGHT

2.1. Caribbean rainfall background

The annual rainfall cycle for the Caribbean basin reveals a distinct bimodal pattern with peaks during the late spring and late summer months. A relative minimum during the mid-summer, known as the mid-summer drought (MSD) (Magaña et al. 1999) separates the Primera or the early rainfall season (ERS) from the late rainfall season (LRS). Definitions of the ERS period range throughout related literature and vary from April to July (Angeles et al. 2010), May to July (Spence et al. 2004) and May to June (Chen and Taylor 2002). Despite the range in timing and duration of the ERS, rainfall during the ERS is critical for Caribbean agriculture during the early growing season (Gamble et al. 2010). Reduced rainfall after the ERS results in a decline in vegetation health during the MSD (Allen et al. 2010). Following the MSD, the second and wetter rainfall peak during the LRS coincides with passing tropical storms. Extreme rainfall connected to tropical storms and warmer sea surface temperatures during the late summer may explain the higher rainfall totals found in the second peak of the bimodal rainfall pattern, however, this topic requires further investigation. Descriptions of spring rain onset dynamics in the Caribbean are absent in the research literature. Can seasonal patterns in the general circulation be identified and further help diagnose the cause of early season rainfall onset?

The sea surface temperature (SST) threshold for deep convection is between 26–27°C in the tropics, depending upon region and season (Waliser and Gautier 1993). The
Atlantic Warm Pool (AWP), a warm body of water (SST > 28.5C) west of 50W and north of the equator, satisfies the SST threshold criteria for deep convection in the Caribbean (Wang and Enfield 2001). The AWP promotes surface evaporation leading to elevated column integrated water vapor content or precipitable water (PW). The variability of the AWP onset date exhibits similar variability with ERS onset dates (Misra et al. 2014). Previous studies that describe the relationship between SST and rainfall suggest that Caribbean rainfall may be driven by fluctuations in local SST (Curtis 2013). The seasonal cycle of Caribbean SST may provide clues to understanding the onset of early season rainfall.

Inter-annual variability is observed in annual rainfall totals, length of the rainy season, and length and timing of the MSD (Angeles et al. 2010; Curtis and Gamble 2007; Gamble et al. 2008; Jury et al. 2007; Magaña et al. 1999). Various mechanisms have been proposed to explain the MSD that include evolving regional SST anomalies (Magaña et al. 1999), a westward extension of the North Atlantic Subtropical High (NASH) (Giannini et al. 2000; Kelly and Mapes 2011) variations in the summertime Caribbean Low Level Jet (Muñoz et al. 2008), contributions from Saharan dust (Angeles et al. 2010), and the biannual crossing of the solar declination (Karnauskas et al. 2013). However, the dynamics responsible for the onset and decay of rainfall in the Caribbean during the ERS along with its associated variability remain understudied and poorly understood. Could additional knowledge gleaned from diagnosing ERS related rainfall cessation be used to diagnose the onset of the Caribbean MSD? It is possible that a modified explanation of the MSD that takes the cessation of late spring rains into account may help to explain mid-summer drying.
The El Niño-Southern Oscillation (ENSO) phenomenon is known to affect the inter-annual variability of rainfall in the Caribbean region (Giannini et al. 2001; Chen and Taylor 2002). Early season rainfall is reduced prior to a mature El Nino and is enhanced following a mature El Nino and during La Nina years. Drier than normal summertime conditions are exacerbated when a positive winter NAO phase is observed immediately prior to a developing El Nino (Giannini et al. 2001). Intra-seasonal variability in Caribbean rainfall is also observed and may be linked to phases of the Madden Julian Oscillation (MJO). Intra-seasonal variability in Caribbean rainfall is largest during the LRS, but some modulation of rainfall by the MJO appears in all seasons (Martin and Schumacher 2011). In addition, variability in Atlantic and Caribbean tropical storm frequency has been attributed to the MJO (E. D. Maloney and Hartman 2000; Klotzbach 2010). The dependence of tropical rainfall on the MJO may affect the rainfall totals during the LRS and perhaps the ERS as well.

Other than global teleconnections like ENSO, the MJO, and the NAO, there are no physical mechanisms proposed that offer dynamical explanations for spring rain onset in the Caribbean. Therefore, identifying Caribbean rainfall dynamics may be difficult if examined in isolation. Instead, can placing the ERS rainfall in context with other seasonal rain patterns in the subtropics help to explain the trigger mechanisms that are responsible for the first peak of the Caribbean bimodal rainfall pattern?

Global average precipitation estimates for June through August reveal a zonally elongated rain-belt feature on the western edge of subtropical anticyclones (figure 1.1). These features are reviewed in related literature with various names: subtropical precipitation zones, subtropical convergence zones, quasi-stationary rain-belts, quasi-
stationary frontal zones, and subtropical frontal zones (Kodama 1992; Ninomiya 1984; Sampe and Xie 2010; Enomoto et al. 2003). Here we will use the term quasi-stationary rain-belt. The dynamics of these features are distinct from mid-latitude polar fronts and the ITCZ, but common characteristics are observed.

![15 May - 31 Aug TRMM climatological accumulated rainfall](image)

Figure 1.1. May 15 – August 31 climatological accumulated rainfall averaged from 1998-2012 from TRMM.

Active convection with steady and substantial precipitation in the quasi-stationary rain-belts is observed, similar to the ITCZ, while stronger winds aloft and a low level baroclinic wind structure resembles a mid-latitude front (Y. Kodama 1992). Mid to upper tropospheric westerly flow is a large scale common feature found in all STCZs where subtropical jets flow in the sub-tropical latitudes (Kodama 1993). Moist adiabatic ascent throughout these rain-belts is supported by the presence of frontogenesis along a westerly subtropical jet, warm mid-level temperature advection, and low level poleward moisture
Low level poleward flow around the western edge of subtropical anticyclones is a fundamental process that supplies moisture for STCZs in both hemispheres (Kodama 1993). Heavy precipitation occurs within these regions during the more active summer months when all of the above key processes are met. The three strongest rain-belt features that produce in excess of 400 mm/month of rainfall during their active phase are the South Atlantic Convergence Zone, the South Pacific Convergence Zone, and the Meiyu-Baiu Frontal Zone (East Asia).

The weaker North Atlantic Convergence Zone (NACZ) reaches into the Atlantic during July when upper tropospheric westerly flow remains north of the subtropics. Prior to July, a distinct time averaged rain-belt feature is observed in the Caribbean Sea during the ERS (figure 1.2). The Caribbean rain-belt pattern extends northeastward from the Caribbean coast of Central America towards the Bahamas. A distinct dry zone is seen during the ERS throughout the Gulf of Mexico, which suggests that passing late season mid-latitude fronts do not fully explain the presence of rainfall in the Caribbean during the ERS. Higher rainfall amounts in the Gulf of Mexico would be expected if trailing fronts alone contributed to the regional rainfall during the ERS. Since this is not the case, we conclude that there are dynamical factors other than frontal dynamics alone that produce rainfall along the Caribbean rain-belt during the ERS.
An annual bimodal rainfall pattern exists throughout the Caribbean Sea and Gulf of Mexico (Magana et al. 1999). However, bimodality, or the differences between the relative minimum and the average of the first and second rainfall peaks, is not uniform across the region (figure 1.3).
Weak rainfall bimodality is seen in the Gulf of Mexico and the eastern Caribbean with the strongest bimodal signals observed in the Caribbean rain-belt region. Angeles et al. (2010) calculate a bimodal rainfall index that confirms the highest bimodal signal in the Caribbean is found along the Caribbean rain-belt region (Angeles et al. 2010 figure 1.3). High rainfall bimodality along the Caribbean rain-belt suggests that processes leading to the Caribbean rain-belt are distinct from other areas in the southeast Caribbean and Gulf of Mexico where both the ERS rainfall is less and the bimodal index is smaller.

Subtropical frontal zones are observed over areas with weak surface temperature gradients. For example, the South Pacific Convergence Zone and the Caribbean rain-belt form over uniformly warm waters. If the Caribbean rain-belt is a front, then perhaps the horizontal temperature gradient that offsets thermal wind balance exists aloft. Molinari and Vollaro (2012) show how upper level quasi-geostrophic dynamics can affect the
lower troposphere by relating clouds in the North Pacific gyre to “the dynamical forcing of vertical motion” in the 400-200 hPa layer (Molinari and Vollaro 2012). In the atmosphere, the thermal wind equation dynamically links the vertical shear of the geostrophic wind to horizontal density gradients. Therefore, fronts and jet streams are inseparably linked by the thermal wind equation. But, if forcing can be found in the upper layers of the atmosphere, then how does upper level quasi-geostrophic forcing relate to the lower troposphere to govern convection?

This chapter focuses on the dynamics related to the onset and cessation of the observed Caribbean rain-belt phenomenon during the ERS. The observed climatological Caribbean rain-belt rain pattern during the ERS provokes the primary research question: What makes it rain in the Caribbean rain-belt time-space pattern? This unknown is addressed by answering three related questions.

1) Can we establish an improved definition of ERS onset and cessation from gridded satellite based precipitation estimates?
2) What are the dynamical factors that cause the climatological Caribbean rain-belt phenomenon?
3) By placing the Caribbean rain-belt in context with another northern hemisphere analog like the Pacific meiyu-baiu, can we offer suggestions that explain mid-summer Caribbean drying that coincides with increasing North Atlantic rainfall?

2.2 Data

Gridded precipitation estimates are favored over station rainfall observations throughout this study due to the larger fractional area of water compared to land.
Therefore, precipitation leading to the ERS Caribbean rain-belt feature is detected from the evenly spaced 0.25 degree Tropical Rainfall Measuring Mission (TRMM) (Huffman et al. 2007) at daily intervals from 1998 to 2012. ERS precipitation climatology is defined as the 15 year average rainfall between May 15 and June 15 from 1998 to 2012. The Modern Era Retrospective Analysis for Research and Applications dataset (MERRA, Rienecker et al. 2011) is used over the same period to analyze vertically integrated total precipitable water (PW) which is used to describe integrated atmospheric moisture content. MERRA PW is compared with daily NOAA Optimum-Interpolation SST fields at 0.25 degree resolution to determine the relationship between SST and PW.

The NOAA NCEP-NCAR CDAS-1: Climate Data Assimilation System I; NCEP-NCAR Reanalysis Project (Kalnay et al. 1996) is the primary dataset used in this study to analyze the meteorological fields. Using the same 15 year range mentioned above, the NCEP/NCAR Reanalysis data is projected onto a horizontal grid with regular grid spacing of 2.5 degrees longitude and latitude and is accessed from 10 unevenly spaced vertical levels from 1000hPa to 200hPa.

MERRA 3 hourly analysis of the meteorological fields are used to diagnose case studies of extreme rainfall and flooding during two major flood events in the Caribbean.

2.3 Northern hemisphere rain-belts

The meiyu-baiu and the Caribbean rain-belt shape the late spring to early summer rainfall pattern in the North Pacific and Caribbean respectively. However, there has been scant research focused on diagnosing and predicting Caribbean rainfall during the ERS despite the strong dependence of agricultural success on early season rains. In contrast, the dynamics leading to the meiyu-baiu and its variability have received considerable
research attention throughout the past two decades. Placing the Caribbean rain-belt in context with the meiyu-baiu allows us to gain insight from pre-existing research centered on the Pacific subtropical convergence zone.

### 2.3.1 Meiyu-Baiu

Rainfall within the meiyu-baiu rain-belt may be continuous or intermittent for days to weeks and may include rain showers or thunderstorms (Chen and Chang 1980). The rain-belt first appears near the South China Sea around mid-May and migrates northward towards Japan where it abruptly ends in mid-July. Remnants of the meiyu-baiu propagate towards northern China and Korea in late July and form what is known as the Changma (rain season). The meiyu-baiu frontal zone is characterized by a distinct quasi-stationary rainband that defines the North Pacific Convergence Zone (Sampe and Xie 2010, Kodama 1992). However, dynamical discontinuities exist within the rainband (Chen and Chang 1980). The western section is dominated more by shallow convection while the eastern section resembles a mid-latitude baroclinic front. Sampe and Xie (2010) attribute mid-tropospheric temperature advection (terms D and F) and local diabatic heating (term B) in the thermodynamic energy equation (equation 1.1) as the two primary factors that facilitate adiabatic uplift into the meiyu-baiu.

\[
\frac{\partial T}{\partial t} = \frac{\overline{Q}}{C_p} - \left(\frac{p}{p_0}\right) \frac{R C_p}{\alpha} \frac{\partial \theta}{\partial p} - \nabla \cdot V_p \overline{T} = \left(\frac{p}{p_0}\right) \frac{R C_p}{\alpha} \frac{\partial \theta}{\partial p} - \nabla \cdot V_p \overline{T}
\]

(equation 1.1)

They conclude that ascent along the meiyu-baiu rainband from 30N to 40N closely follows the region of positive temperature advection at 500hPa by a westerly jet. In this
case, horizontal advection (terms D and F) and diabatic forcing (term B) balance the vertical advection (term C) into the meiyu-baui. They also identify moisture advection into the rain-band by moist low-level southwesterly flow along the western edge of the subtropical high as a moisture source to feed the rain-belt. Rainfall along the meiyu-baui exhibits interannual variability which is related to the interannual variability of the mechanisms identified by Sampe and Xie (2010), namely warm horizontal temperature advection at 500hPa from the Tibetan Plateau (Kosaka and Nakamura 2011).

2.3.2 Caribbean rain-belt

TRMM gridded precipitation estimates during the ERS show evidence of a similar meiyu-baui like seasonal rain-belt pattern in the Caribbean. Like the meiyu-baui, the Caribbean rain-belt stretches from the southwest to the northeast during the late spring and early summer. Interannual variability causes the Caribbean rain-belt to be more pronounced during some years than others (figure 1.4). For example, the Caribbean rain-belt is prominent during 2012 and is indistinguishable in other years (1998, 2001, and 2004). In addition, in some years (2008 and 2011) the Caribbean rain-belt is more zonal, lacking a southwest to northeast slope. Despite its variability, the Caribbean rain-belt is present in the long term precipitation mean in the Caribbean. Remote teleconnections (ENSO, MJO, NAO) have been known to modulate the interannual variability of rainfall during the Caribbean rain season, which may affect rainfall intensity along the Caribbean rain-belt.
2.4 Quasi-geostrophic approximation framework for diagnosing the Caribbean rain-belt

The relationship between synoptic weather activities and the underlying ERS rainfall climatology time-space pattern are addressed from a dynamical perspective in the quasi-geostrophic (QG) approximation framework. Considering that flow in synoptic-scale weather systems is approximately geostrophic, a simplified set of equations are used to diagnose forcing for vertical motion and to describe the geopotential tendency or the processes responsible for the development of weather systems. Time averaging of the seasonal weather systems ultimately shape the Caribbean rain-belt time-space pattern.
From equation 1.2, vertical motion response (term A) giving rise to “weather” is diagnosed by the forcing terms (B and C) in the QG omega equation.

\[
\left( \nabla^2 + \frac{f_0^2}{\sigma} \frac{\partial^2}{\partial P^2} \right) \omega = - \frac{f_0}{\sigma} \frac{\partial}{\partial P} \left(- \nabla g \cdot \nabla \eta_g \right) + \frac{1}{\sigma} \nabla^2 \left[- \nabla g \cdot \nabla \left( \frac{\partial \Phi}{\partial P} \right) \right]
\]

(equation 1.2)

A \hspace{1cm} B \hspace{1cm} C

Term A describes the vertical velocity, while B and C describe the differential vorticity advection and the thermal advection terms respectively. Specifically, term B represents the vertical derivative of the absolute geostrophic vorticity advection by the geostrophic wind and indicates forcing for ascent when positive. Therefore positive values for B are consistent with cyclonic vorticity advection increasing with height. Term C is proportional to thickness advection where warm advection is associated with forcing for ascent. Term C is zero when the geostrophic wind flows parallel to thickness contours. In essence, the QG omega equation describes forcing for ascent to be associated with cyclonic vorticity advection and areas of warm advection where each forcing term can be diagnosed independently. In the QG approximation framework, \( \omega \) is the vertical component of a restoring ageostrophic circulation that acts to decrease horizontal temperature gradients and bring the atmosphere back to thermal wind balance. The QG omega equation does not provide a quantitative estimate of \( \omega \), but instead provides a conceptual understanding of \( \omega \) and its causes.
The QG height tendency equation reveals processes that contribute to the
development or decay of weather events (equation 1.3). The QG height tendency
equation differs from the QG omega equation in the fact that it is a prognostic equation
with a time derivative term that is used to deduce changes in geopotential height fields.

\[
\nabla^2 + \frac{d}{dp} \left( \frac{f_0^2}{\sigma} \frac{d}{dp} \right) \chi = -f_0 \vec{V}_g \cdot \nabla \left( \frac{1}{f_0} \nabla^2 \Phi + f \right) - \frac{d}{dp} \left[ -\frac{f_0^2}{\sigma} \vec{V}_g \cdot \nabla \left( \frac{d\Phi}{dp} \right) \right]
\]

(equation 1.3)

Term A in the QG height tendency equation describes height tendencies resulting
from advection of absolute geostrophic vorticity by the geostrophic wind (term B) and
the vertical variation of geostrophic thickness advection (term C). Advection of positive
absolute geostrophic vorticity at a location is related to falling heights while negative
absolute vorticity advection at a location produces rising heights. Term C describes the
impact of height tendencies by cold and warm air advection. Cold advection decreasing
with height acts to produce height falls and amplify existing troughs, while warm
advection decreasing with height acts to produce a height rise and amplify existing
ridges. The QG height tendency equation is useful at describing “digging” or “lifting”
troughs resulting from either the import or export of cyclonic shear vorticity. Jet streaks
associated with late season mid-latitude cold fronts and the subtropical jet can have major
impacts on trough development and precipitation in the Caribbean during the late spring.
2.5 Caribbean rain-belt climatology

Diagnosing the cause of the Caribbean rain-belt time-space rainfall pattern is examined via the QG approximation framework from the QG-omega equation and the QG-height tendency equation. Together, these equations address the fundamental forcings responsible for the vertical ascent and thickness advections that can either activate or suppress precipitation.

2.5.1. Time mean rainfall

A distinctive Caribbean rain-belt time mean spatial pattern is more prominent during May 15\textsuperscript{th} to June 15\textsuperscript{th} compared to the June rainfall monthly mean. The accumulated rainfall difference between the climatological May 15\textsuperscript{th} to June 15\textsuperscript{th} period and the long term June monthly mean reveals a wetter pattern throughout the central Caribbean during May 15\textsuperscript{th} to June 15\textsuperscript{th}. The accumulated rainfall climatology for June relative to the mid-May to mid-June period highlights a wetter region north of the Caribbean rain-belt into the Gulf of Mexico and southeast USA. Rainfall during the mean mid-May to mid-June period is consistent with the climatological Caribbean rain-belt spatial pattern during the ERS that characterizes the first peak of the Caribbean bimodal rainfall pattern (figure 1.5). Throughout this paper, the ERS period rainfall and its related synoptic-scale dynamics are examined in terms of the May 15\textsuperscript{th} to June 15\textsuperscript{th} long term mean although the exact start and end dates of the rainy season vary from year to year.
Figure 1.5. The difference between the average accumulated rainfall during June from the average accumulated rainfall during the May 15 – June 15 Early Rainfall Season. Average accumulated rainfall was computed from TRMM daily measurements.

2.5.2. Lower troposphere

Poleward flow by low level southerlies out of the tropics at the 850hPa level is observed during the ERS throughout the Caribbean basin. Figures 1.6a-c show the seasonal transition of rainfall, sea level pressure, and 850hPa winds for April (pre-ERS), the ERS, and July (the MSD period) climatology. A relative lack of rainfall is observed during the Caribbean dry season in April despite the presence of low level southerly flow around the western flank of the NASH. The meridional component of the circulation at 850hPa increases during the ERS, thus enhancing southerly flow into the Caribbean rainbelt from tropical latitudes.
Figures 1.6 a-c. Average sea level pressure isobars (1hPa contours > 1014hPa), 850hPa wind vectors (max 10 meters per second), and TRMM rainrate (shaded) for April (a), Early Rainfall Season (b), and July (c).

After the ERS, rainfall along the Caribbean rain-belt diminishes during the MSD period while the circulation at 850hPa becomes more zonal and less favorable for moisture advection out of the tropics. Surface pressure associated with the NASH strengthens during the MSD and extends further into the western Caribbean and may contribute to mid-summer drying (Kelly and Mapes 2011).

2.5.3 Mid-troposphere

Figures 1.7 a-c show the seasonal transition of the 500hPa climatological circulation. A synoptic scale anti-cyclone is centered over southern Mexico in April and dominates the mid-tropospheric general circulation of the region. Northerly winds along the eastern edge of the anti-cyclone advect cooler air from higher latitudes into the warmer Caribbean. The response to cold air advection (thickness advection < 0 in QG omega equation) into the Caribbean represents a forcing that suppresses rainfall by
inhibiting ascent. In addition, anticyclonic differential vorticity advection (differential vorticity advection < 0 in QG omega equation) increases with height from the surface to 500hPa throughout the western Caribbean and represents additional forcing for descent.

Later during the ERS, the anti-cyclonic circulation shifts north and weakens over southern Mexico. At the same time a second anti-cyclone develops over the eastern Caribbean and advects warm air along its western edge from the tropics into the Caribbean rain-belt region. Vertical motion response from increasing thickness over the region supports ascent and precipitation throughout the Caribbean.

The twin anti-cyclones migrate northward during the MSD period and are observed over the southwestern section of the United States and in the north Atlantic. At that time, mid-tropospheric flow throughout the Caribbean becomes more zonal, which results in geostrophic flow parallel to geopotential height contours and a termination of temperature advection from the tropics. In addition, the predominately zonal flow through the Caribbean reduces the chances of ascent by limiting positive vorticity advection into the region.
Figure 1.7 a-c. Average 500hPa temperature and 500hPa streamlines for April (A), the Early Rainfall Season (B), and July (C).
Following the methods used by Sampe and Xie (2010), contributions from mid-tropospheric uplift are diagnosed from the thermodynamic energy equation in pressure coordinates such that the vertical advection term balances the horizontal advection term within the tropics. Zonal temperature gradients are weak throughout the Caribbean, but larger meridional temperature gradients are present. The meridional component of the horizontal temperature gradient therefore dominates the horizontal advection term of the thermodynamic energy equation.
Figure 1.8 a-c. Average horizontal positive temperature advection (contours, 0.03Ks⁻¹ interval) and vertical velocity (shaded) at 500hPa for April (a), the Early Rainfall Season (b), and July (c). Climatology calculated from 1998-2012.

Figures 1.8 a-c show the two balancing terms, horizontal temperature advection and vertical velocity, of the thermodynamic energy equation at 500hPa. To satisfy the thermodynamic energy equation, the positive temperature advection is balanced by upward vertical velocity (−ω). The mid-tropospheric zonal temperature gradients in the Caribbean are weak. However, the combination of a strong meridional temperature gradient combined with southerly flow supports positive temperature advection at 500hPa across latitudes. During the ERS, vertical velocity is collocated in areas of positive temperature advection in the mid-troposphere. Southerly flow, positive temperature advection, and positive vertical velocity trace the region along the Caribbean rain-belt during the ERS. The northerly transition of positive horizontal temperature advection away from the Caribbean during the mid-summer reduced the forcing needed for ascent helps and thus helps to explain the onset of the Caribbean MSD.
2.5.4. Upper troposphere

Here it is important to restate the significance of a westerly jet in maintaining the presence of STCZs (Kodama 1992). The jet stream along the meiyu-baiu steers transient disturbances from the mid-latitudes into areas of elevated low level moisture (Sampe and Xie 2010). A similar westerly jet that steers baroclinic disturbances into areas of high PW is found along the Caribbean rain-belt and is consistent with circulation found along the meiyu-baiu. In both regions, jet streams serve as a wave guide for passing late season mid-latitude synoptic disturbances. A positive relationship exists between upper tropospheric zonal wind and rainfall in the Caribbean during the early summer months (Small et. al 2007; Maloney and Esbensen 2007).

The wind speed and direction, zonal wind speed, and positive relative vorticity at 200hPa throughout the Caribbean vary from April to July (figures 1.9 a-c).
Figure 1.9 a-c. Average zonal wind (shading), positive relative vorticity (contours, 0.2 x 10^-5 s^-1), and wind vectors (max 40m/s) at 200hPa for April (a), the Early Rainfall Season (b), and July (c). Climatology calculated from 1998-2012.

Circulation at the 200hPa level during April is westerly throughout the Caribbean and diminishes in strength at lower latitudes adjacent to the ITCZ region. 200hPa westerly flow remains unchanged in the central Caribbean into the ERS, but becomes easterly at low latitudes near the equator. A “bulls-eye” of positive vorticity centered on
the Florida peninsula is embedded in the 200hPa westerly flow during the ERS. The presence of a pronounced area of positive vorticity is significant because time averaged positive vorticity during the ERS is difficult to capture in the climatological record due to the transient nature of passing mid-latitude troughs. This area of positive vorticity is positioned to contribute to falling geopotential heights into the Caribbean by advecting positive absolute vorticity (term B of the QG height tendency equation, equation 1.2) into the area by late season upper level troughs. Falling heights related to a net import of cyclonic vorticity may produce “digging” troughs which can extend deeper into the Caribbean to disrupt the thermal wind balance and lead to frontogenesis. If this happens, uplift through ageostrophic vertical velocity develops to weaken the horizontal temperature gradients in order to restore thermal wind balance. If such uplift occurs in the presence of a moist background state, then “weather” will develop and precipitation along the Caribbean rain-belt region will occur.

Upper tropospheric westerlies remain over North America during the MSD period and either reverse or significantly weaken throughout the Caribbean during that time. The emergence of the drier Caribbean MSD period coincides with the weakening or reversal of positive upper-tropospheric zonal flow (figure 1.10).
200hPa zonal wind speed during the MSD period approaches zero or becomes negative throughout the Caribbean rain-belt region while remaining positive in the higher latitudes of the NACZ (latitudes > 30N). The climatological shift from u200 > 0 to u200 < 0 occurs abruptly during early July from about 12N to 30N. Upper level positive vorticity is embedded in easterly flow in the northern Caribbean and is coincident with a weakening Caribbean rain-belt during the MSD.

Rossby waves are unable to transport transient wave energy eastward towards the Caribbean due to the easterly background winds during the mid-summer. The Rossby wave phase speed equation (equation 1.4) requires positive background zonal wind speed (U > 0) in order for troughs and ridges to move towards the east (positive phase speed, c) in a barotropic atmosphere. Therefore, vertical motion and rainfall connected to upper level troughs via frontogenesis shifts away from the Caribbean during the MSD period.
and increases along the NACZ where the background upper level zonal wind remains positive.

\[ c = U - \frac{\beta L^2}{4\pi^2} \]  

(equation 1.4)

In general, the circulation of the upper troposphere is marked by a series of troughs and ridges that aid in geostrophic vorticity advection and differential temperature advection to produce frontogenesis. Frontogenesis, the lagrangian time tendency of an increase in a temperature gradient, acts to move the atmosphere out of thermal wind balance. In order to restore thermal wind balance, a direct secondary circulation develops in association with frontogenesis and causes upward motion. The vertical motion decreases the thermal gradient by adiabatic expansion or compression (promoting frontolysis) and can be diagnosed from the quasi-geostrophic omega equation.

Near surface horizontal density variations are not an essential requirement for ageostrophic uplift to develop. An example of this exception occurs along the subtropical jet in the Caribbean. Upper level westerlies enhance early- and late- summer Caribbean wet spells by steering subtropical jet-stream troughs into the Caribbean (Jury et al. 2007). Forcing for uplift occurs in the area below the right entrance region of a jet streak due to positive vorticity advection in that specific quadrant of the jet. Assuming a simplified 2D zonal jet with no meridional component in wind speed, vorticity is estimated by: \( \zeta = \frac{dv}{dx} - \frac{du}{dy} \approx -\frac{du}{dy} \). In this sense, positive vorticity advection (term B in equation 2) forces rising motion (term A equation 2) near the jet right entrance region. Therefore, upper
level circulation associated with a westerly jet provides a dual purpose: it is a wave-guide to steer transient disturbances and it provides an uplift mechanism (- $\omega$).

The strongest easterly winds in the Caribbean are found at the 925hPa Caribbean Low Level Jet (< -10 m/s) during the mid-summer. There has been considerable research linking variability in the strength of the Caribbean Low Level Jet to the variability in Caribbean summertime rainfall (Muñoz et al. 2008; Whyte et al. 2007; Taylor, Whyte, et al. 2013; Martin and Schumacher 2011). As the summer wears on, the Caribbean Low Level Jet strengthens and westerly winds in the upper troposphere turn over to easterlies during the drier mid-summer period. As a result, the Rossby wave guide is no longer present in the Caribbean during the mid-summer and the Caribbean rain-belt weakens and an area of enhanced rainfall is displaced northward during the MSD period towards areas of positive zonal flow. In essence, the dynamics related to the Caribbean rain-belt retreat from negative 200hPa zonal flow during the MSD period and re-establish at higher latitudes to enhance the NACZ along positive 200hPa zonal flow.

2.5.5. SST and water vapor

The concentration of atmospheric water vapor varies seasonally throughout the Caribbean and is modulated by SST. The daily time series of PW and SST in the Caribbean from 2000 to 2010 are positively correlated, illustrating a seasonal dependence on warm SST for increased PW and rainfall potential (figure 1.11).
Figure 1.11. The seasonal cycle of precipitable water (red, units in mm) and sea surface temperature (blue, units in °C) from 2000 – 2010. Sea surface temperature time series was calculated from the area average between 14.2 - 17.42°N and 278.12 – 298.12°W. Precipitable water time series was calculated from the area average between 13.33 – 17.14°N and 278.44 – 299.06°W.

Warm SSTs found in the AWP (SST>28°C) provide a potential on monthly time scales for a background environment of high PW values (PW > 50mm). The frequency of occurrence of warm SST, not SST alone, may support a regime towards strong atmospheric convection. High values of PW on the other hand vary on a day to week time scale and are sculpted by lower-tropospheric background circulation from the moist tropics. Convection along the Caribbean rain-belt is bound by islands of high PW that advect around the AWP. Strong convection is most tied to the vertically averaged tropospheric temperature, which acts as a point of “self-organized criticality” for a transition to an increased rain rate (Peters and Neelin 2006).

The seasonal variation of SST and PW between the pre-ERS and MSD period is given in figures 1.12 a-c. Suppressed PW observed prior to the ERS (PW < 40mm) is
expected due to cooler Caribbean basin SST. SST in the Caribbean increases into the MSD period as noted by the northward progression of the 28.5°C isotherm. The 40mm PW contour also advances northward during the MSD period when SST is warmer relative to the ERS.

There is a discontinuity in the climatological 43mm PW contour throughout the central Caribbean during the MSD period (figure 1.12d). A reduction in mid-summer PW in the central Caribbean exists despite the presence of a uniform AWP related 28.5°C SST isotherm. The discontinuous 43mm PW contour is also collocated with the western extension of the NASH during the MSD which suggests that SST alone does not modulate PW levels in the Caribbean. The mid-summer pressure increase described by Curtis and Gamble (2008) as the “pressure-bump” may lend support to the existence of a discontinuous PW contour and suppressed rainfall during the MSD. For reference, the Caribbean bimodal rainfall signal is the strongest where the mid-summer 43mm PW gap exists.
Figure 1.12 a-d. Average precipitable water (shaded), 40mm precipitable water contour, and 28.5°C sea surface temperature isotherm for April (a), the Early Rainfall Season (b), and July (c). Climatology calculated from 1998-2012. (D), same as (c), but PW contour is 43mm.

The highest concentration of atmospheric water vapor resides in the lower troposphere and is advected by low level winds. The strongest easterly winds in the Caribbean are found at the 925hPa Caribbean Low Level Jet (< -10 m/s) during the mid-summer. There has been considerable research linking variability in the strength of the

The change in strength of the 850hPa meridional wind from the pre-ERS to the MSD period impacts moisture advection into the Caribbean rain-belt region. Southerly flow at 850hPa arises geostrophically between a heat low towards the west near Central America, known as the Panama Low, and the NASH to the east. Moist low level meridional wind increases towards the Caribbean rain-belt region as the low level zonal geopotential height gradient increases from the pre-ERS to the ERS period (figure 1.13a).

Later, during the MSD period, zonal circulation at 850hPa dominates the flow pattern throughout the Caribbean as the NASH extends further west and weakens the zonal geopotential height gradient (figure 1.13b). The weakening low level geopotential height gradient decreases the meridional component of the 850hPa wind out of the tropics.

Southerly flow at 850hPa is displaced from the Caribbean rain-belt area towards higher latitudes along the NACZ during the MSD period. The moist 850hPa southerly flow that supports rainfall for both the Caribbean rain-belt during the ERS and the NACZ during the MSD period meets the low level poleward flow criteria proposed by Kodama (1992) and supported by Sampe and Xie (2010).
Figure 1.13 a-b. Climatological difference of 850hPa zonal wind between the Early Rain Season minus April (a) and July minus the Early Rain Season (b). Climatological averages are calculated from 1998-2012.

Moisture flux convergence is described from the conservation of water vapor in pressure coordinates (equation 1.5) where S is water vapor storage, which can be represented by the difference between sources (evaporation rate) and sinks (precipitation rate) or S = E - P.

\[
\frac{dq}{dt} = S = E - P \tag{equation 1.5}
\]
Using the continuity equation, $\frac{dq}{dt}$ can be expanded in flux form, which conserves the total mass of water (equation 1.6).

$$\frac{\partial q}{\partial t} + u \frac{\partial q}{\partial x} + v \frac{\partial q}{\partial y} + \omega \frac{\partial q}{\partial p} + q \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial \omega}{\partial p} \right) = E - P$$  \hspace{1cm} \text{(equation 1.6)}

which can be further written as equation 1.7.

$$\frac{\partial q}{\partial t} + \nabla \cdot \left( q \mathbf{V} \right) + \frac{\partial}{\partial p} (q \omega) = E - P$$  \hspace{1cm} \text{(equation 1.7)}

Therefore, we see that the total rate of change in water vapor, $q$, depends on horizontal moisture convergence (term 2 in equation 1.7).

Horizontal velocity convergence is associated with ascent and can be related to a deepening of the lower-tropospheric layer of moisture. Regions of locally deeper moisture are more favorable for convection initiation than surrounding areas, all else being equal. Updrafts triggered in these areas either through diabatic heating or dynamical forcing for ascent experience larger CAPE. Therefore, while moisture advection alone may not contribute to an overall time rate of change in local water vapor mixing ration, the deepening boundary layer resulting from horizontal velocity convergence can initiate convection.

2.6 The cessation of the Caribbean rain-band and the emergence of the MSD

A relative dry period develops along Japan and throughout the Caribbean as both the meiyu-baiu and Caribbean rain-belt advance poleward during the mid-summer. The poleward migration of the rain-belts and the dryness in its path can be partially explained by variations in local circulation. For the meiyu-baiu, the Bonin High develops during
the mid-summer and its related subsidence helps to suppress rainfall over Japan while the westward extension of the NASH and diminishing upper level westerlies inhibit rainfall throughout the Caribbean during the MSD period (Enomoto et al. 2003; Gamble et al. 2008).

![Graph of rainfall comparison](image)

Figure 1.14. Daily climatological time series of rainfall within the Caribbean Sea (black) and the North Atlantic (red) from 1998-2012. The Caribbean Sea time series is averaged between 14-29N and 80-75W and the North Atlantic time series is averaged between 30-35N and 70-65W.

A noticeable increase in rainfall along the NACZ coincides with a decrease in rainfall along the Caribbean rain-belt during the MSD (figure 1.14). Rainfall increases abruptly along the NACZ in July as the Caribbean rain-belt pattern fades and a drier mid-summer climate is established in the Caribbean. The rainfall difference between the ERS and MSD period shows a distinct decrease in rainfall for the Caribbean rain-belt and an increase along the NACZ during the mid-summer (figure 1.15). The shifting rain-belt
pattern from the Caribbean to the North Atlantic suggests a northward displacement of rainfall towards areas of positive upper level zonal wind during the MSD.

Various experiments have been designed to test the influence of the Gulf Stream on the mid-summer rainfall maximum associated with the NACZ (Kirtman et al. 2012; Minobe et al. 2008). However, model results suggest that a climatological rain-belt in the vicinity of the NACZ exists even in the presence of a smoothed SST field that eliminates the pattern of the warm Gulf Stream (Yoshida et al. 2010). Therefore, it is likely that the northward shifted rainfall away from the Caribbean during the MSD period enhances rainfall along the NACZ while exacerbating dryness in the Caribbean. Similarly, the difference in rainfall between the meiyu and baiu periods also exhibits a discernable northward shift from the early to mid-summer from China to Japan.

Figure 1.15. Average TRMM rain-rate of the Early Rainfall Season minus July (mm/day). Climatology calculated from 1998-2012. Red colors indicate areas where July rainfall is less than rainfall during the Early Rainfall Season.
Accumulated rainfall climatology from May 15th to July 31st shows a continuous rain-belt stretching from the Caribbean rain-belt into the NACZ (figure 1.16). The continuous climatological rain-belt, the CARNA-belt (Caribbean and North Atlantic Rain belt), resembles the meiyu-baiu rain-belt in the fact that it shares distinct rainfall regimes from the early (ERS) to late (MSD) formation periods. Following the nomenclature of the meiyu-baiu, perhaps the generalized North Atlantic Convergence Zone from previous research should be reconsidered as the CAR-belt (Caribbean Atlantic Rain-belt) consisting of part early season Caribbean rain-belt and later season NACZ rainband.

2.7 Case studies – 2 Caribbean floods

Extreme rainfall occurs along the Caribbean rain-belt and can result in the loss of property and life. Dynamics leading to extreme rainfall along the Caribbean rain-belt are analyzed from two case studies that were subjectively chosen to represent flooding events with significant social impacts. These case studies highlight the dynamics of rainfall
along the Caribbean rain-belt so that they can be applied to improve life-saving forecasting.

Heavy and prolonged rains in Jamaica during May 21-27, 2002 initiated widespread flooding and landslides causing 9 deaths, over 500 displaced residents, and more than $10 million in agricultural damage. Seven days of persistent rainfall had the most damaging effect along the island’s southern coastal parishes. In addition, one of the most extensive Caribbean floods in recent history occurred during the ERS between May 18-25, 2004 near Hispanola. Some areas during that period experienced up to ten inches of rainfall, which resulted in over 2,000 deaths and 1,300 lost homes. An improved understanding of the dynamical processes related to the Caribbean rain-belt would be useful for forecasting extreme rainfall during the ERS; especially for hilly to mountainous areas with unstable terrain.

2.7.1 May 21-27, 2002

IR satellite imagery from May 18-20, 2002 highlights a frontal band that moved into the Gulf of Mexico from the mid-latitudes (figure 1.17).

Figure 1.17. GOES-8 infrared satellite imagery at 15:15 UTC for May 18-20, 2002.

Rainfall associated with a related upper level trough persisted from the Gulf of Mexico into the Caribbean until an upper level ridge formed in the region on May 24. Accumulated rainfall during the May 20-24 frontal passage was not uniform across the
Intra-American Sea region. Instead, extensive rainfall (+100mm) fell within a zone from Cuba to Jamaica where PW exceeded 50mm during the 4 day period. Therefore, a combination of a highly moist background environment coinciding with favorable frontal uplift dynamics in the form of positive vorticity advection conspired to produce the bullseye of rainfall in the region. While the passing frontal system played a significant role towards enhanced rainfall, flooding during the May 24-27 period occurred after the trough passage from the initial mid-latitude front. This timing suggests the impact of the upstream dynamics from a front related jet streak.

Figure 1.18 a-d. GOES-8 infrared satellite imagery for May 26, 2002 15:15 UTC (top left). Accumulated TRMM precipitation (mm) between May 24-27, 2002 (top right).
200hPa streamlines, 32m/s windspeed isosurface (yellow shading), TRMM rainfall (rainbow shading), and Q-vector convergence (pink contours) (bottom left). 850hPa geopotential height (black contours), total precipitable water > 40mm (shading) and TRMM precipitation (rainbow shading) (bottom right).

By May 25, 2002 satellite IR imagery shows an absence of any banded front-like cloud pattern within the Caribbean (figure 1.18a). Instead, a mesoscale cloud cluster is observed on May 26 that contributes to additional flooding. The mesoscale system lies within the Caribbean rain-belt region and its accumulated rainfall between May 24-27 is characteristic of the spatial southwest to northeast Caribbean rain-belt rainfall pattern (figure 1.18b). A region of intense rainfall (+10 mm/hr) within the cloud system occurs along the southern coastline of Jamaica on May 26, 2002.

The forcing for ascent that produced the elevated rain-rate at that time can be diagnosed from the QG equations. Ageostrophic uplift dynamics associated with a jet streak above an “island” of high PW > 50mm combine to produce strong forcing for ascent and heavy precipitation (figure 1.18c-d). Upper tropospheric wind speed in excess of 30 m/s resided in the vicinity of the flood region and marked the position of the subtropical jet near the 200hPa level. The flooded region was positioned below the right entrance region of the jet stream where further low level convergence and ascent occurred. Positive temperature advection was observed at the 500hPa level into Jamaica. The resulting differential thermal advection from the surface to the mid-troposphere (term C in the QG height tendency equation) further destabilized the area. A 850hPa zonal geopotential height gradient directs moist low level southerly flow into the region of vertical ascent. PW > 50mm is the primary moisture source required for enhanced rainfall and is advected from the tropics by the lower tropospheric circulation.
Back trajectory analysis for a 36 hour period prior to May 26 reveals southerly 850hPa flow towards Jamaica and a 200hPa circulation along the path of the subtropical jet (figure 1.19). Back trajectory results help to diagnose advection of PW > 50mm out of the tropics and an upper level circulation originating from the subtropics.

Figure 1.19. 36-hour back trajectory output from NOAA Hysplit at 850hPa (left) and 230hPa (right). Stars indicate the origin of back trajectory analysis ending at 1500 UTC May 26 2002.

These results disprove the notion that Caribbean flooding results exclusively from advancing mid-latitude cold fronts. In addition, the back trajectory analysis at each level indicates a destabilizing environment via low and upper troposphere ascent over a moist environment 12-18 hours prior to the May 26th rains.
2.7.2 May 18-25, 2004

Unlike the previous flood of 2002, IR satellite imagery shows no sign of an intruding mid-latitude frontal cloud band for the flood centered on Hispanola in 2004 (figure 1.20).

![Daily GOES-8 infrared satellite imagery from May 18 – 25, 2004.](image)

Instead, clouds emerge from the tropics towards the flood region and ultimately form a rainfall producing mesoscale cloud cluster with peak rainfall intensity of +20 mm/hr on May 24th (figure 1.21a). A relative dry zone exists throughout the Gulf of Mexico during the May 18-25, 2004 period, which indicates the absence of a propagating mid-latitude front (figure 1.21b).
Upper tropospheric uplift dynamics in the form of positive vorticity advection and positive thickness advection along with lower tropospheric tropical moisture transport dynamics are represented in both of the 2002 and 2004 flood cases. An ageostrophic vertical circulation associated with the right entrance region of the subtropical jet near 200hPa contributes to ascent near Hispanola. 200hPa southerly flow over a region of high PW (+50mm) provides a favorable environment for moist ascent leading to heavy
rainfall. Like the 2002 flood event, moist southerly flow from the tropics is supported by zonal gradients in the 850hPa geopotential height contours and low level convergence. Likewise, positive 500hPa meridional flow from the tropics, which results in positive mid-tropospheric temperature advection, is found along Hispanola and encourages moist mid-tropospheric ascent along the flood region (figure 1.21d). Instability found in the upper troposphere (terms B and C of the QG omega equation) and moisture advection by the low level winds both combine to produce rainfall with the characteristic Caribbean rain-belt pattern.

Back trajectory analysis at the 200hPa and 850hPa levels show southerly flow into Hispanola from the moist tropics for the 36 hours prior to the May 24, 2004 floods (figure 22). Again, additional evidence is provided that suggests atmospheric circulation at both the lower and upper troposphere is not driven by mid-latitude cold fronts. Back trajectory analysis indicates ascent at the 850hPa and the 200hPa level over a moist environment, which in essence are the basic ingredients for rainfall.
2.8 Summary and discussion

Rainfall climatology during the ERS shows a quasi-stationary band-like rainfall pattern, known as the Caribbean rain-belt, which extends from the Caribbean coast of Central America towards The Bahamas between May 15 and June 15. A new conceptual model illustrates the climatological Caribbean rain-belt and is based on three primary factors: positive upper tropospheric zonal flow, positive mid-tropospheric temperature advection, and tropical moisture advection by the low level winds (figure 1.23). The distinct Caribbean rain-belt pattern results from time averaged weather events that develop within a common path (represented by the grey cloud) during the ERS. Ascent along this path is forced by a combination of vorticity and thermal advection above islands of high PW values that are advected along the AWP by low level winds (thin
arrows). Cyclonic vorticity is advected into the Caribbean from troughs associated with the westerly jet (thick arrows) while warmer temperatures are advection in the mid-troposphere from the tropics (medium arrows).

Figure 1.23. Schematic diagram representing the environmental factors that contribute to the Caribbean rain-belt. Mid-tropospheric winds (medium black arrows) advect warm air from the tropics and induce ascending motion along the jet stream (thick black lines). The ascent favors convection (grey cloudy diagram) in the presence of low level southerly moisture transport (think black lines) along the western edge of the north Atlantic subtropical high. Transient disturbances are denoted by H and L and are steered by the jet stream.

The lower level southerly circulation develops between Central America and the westward extension of the NASH. It is noted that a negative SST gradient exists along the southern Caribbean during the MSD period due to increasing trades related to the CLLJ maximum in July. Therefore zonal flow over a cooler SST surface lowers the specific
humidity of the southern Caribbean and inhibits moisture advection towards the Caribbean rain-belt region.

The dependence on precipitable water Caribbean rainfall is also evident during the drier winter months. Mid-latitude fronts during the winter can advance into the Caribbean, but measured rainfall along their path is low. Southerly low level flow out of the tropics is observed during the dry winter season, but deep and energetic fronts propagate through the Caribbean too quickly for the southerly flow and related moisture flux to persist as it does during the ERS. Low level southerlies during the winter occur in bursts from passing fronts, while sustained southerly flow at the 850hPa level during the ERS is due to a persistent lower tropospheric zonal pressure gradient across the Caribbean. Moreover, winter dryness also occurs because SST throughout the Caribbean Sea is well below the AWP convective threshold. Therefore, rainfall along the Caribbean rain-belt is more sensitive to low level moisture flux than from mid to upper tropospheric uplift dynamics. However, a lifting trough out of the Caribbean is critical for the characteristic southwest to northeast signature of the Caribbean rain-belt.

Rainfall does occur within the Caribbean during the MSD despite the absence of a westerly jet; however scattered rainfall during that time is less than the preceding ERS and may be driven by instability from local diabatic processes and a warm SST. Mid-summer drying throughout the Caribbean results from an easterly shift in the upper level circulation along with a more zonal low level circulation. The disappearance of the Caribbean rain-belt pattern signals the end of the first peak of the Caribbean bimodal rain season.
Precipitation along the Caribbean rain-belt dries during the mid-summer and increases along the North Atlantic Convergence Zone where upper tropospheric flow is from the west. The CARNA-belt represents a time space rainfall pattern from the Caribbean to the North Atlantic. Rainfall is more active throughout the Caribbean rain-belt during the ERS, but later during the MSD, rainfall along the Caribbean rain-belt ends while precipitation along the NACZ increases. The Caribbean MSD partially arises due to the northward relocation of favorable rainfall dynamics during the mid-summer. The CARNA-belt serves as an Atlantic rainfall analog to the Pacific meiyu-baiu in terms of a summertime poleward shift and related rainfall dynamics.

This research identifies and characterizes a climatological rain pattern that can guide parameter selection for statistical modeling, which in turn is motivated by applications that can benefit from forecasting. We can begin to diagnose the interannual variability of the Caribbean rain-belt by borrowing dynamical ideas from the well-studied meiyu-baiu rainband.
Chapter 3. STATISTICAL PREDICTION OF MULTIPLE TIME SERIES: LAGGED REGRESSION AND LIM

Recall that the purpose of the study above was to choose time series to use in multivariate statistical hindcasting exercises, in hopes of contributing to genuine forecasting. Because the boxes and indices chosen above are themselves provisional or experimental, it is useful to illustrate the statistical forecasting apparatus with a more familiar system: the eastward propagating MJO.

3.1. Illustration with daily MJO data, using longitude boxes as the channels

The Madden Julian Oscillation (MJO) is an intraseasonal large scale zonally propagating atmospheric signal in tropical rainfall and related fields (Madden and Julian 1971). Tropical convection initiates over the Indian Ocean and propagates eastward as the MJO. The MJO signal resembles an atmospheric Kelvin wave with Rossby wave lobes and exhibits propagating westerly and easterly wind anomalies along the equator. Convection associated with the MJO may be amplified as it approaches the east Pacific through interactions with background convection during the Northern Hemisphere summer season (Maloney and Hartman 2000). Upon reaching the Americas, the MJO impacts the precipitation patterns associated with both the North and South American monsoons (Higgins and Shi 2001; Jones and Carvalho 2002). In addition, the influence of the MJO wind anomalies has been shown to modulate Caribbean rainfall variability and hurricane genesis (Martin and Schumacher 2011, Klotzbach 2010, Maloney and Hartman 2000).
Recently, intercomparisons have been applied between statistical models and dynamical models to predict the MJO (Xavier et al. 2014, Klingaman et al. 2014, Newman, et al. 2009). Statistical forecasts have also been applied. For example, Wheeler and Weickmann’s (2001) Fourier filtering method to forecast the OLR signal of the MJO, produces a 15 to 20 day forecast skill (0.5 to 0.6 correlation compared with diagnostically filtered OLR), which they conclude to be comparable to other empirical forecasts. Jones et al. (2004) also forecast filtered OLR anomalies with positive forecast skill out to 5 pentads using multiple linear regression. Cavanaugh et al. (2014) use a suite of linear inverse models (LIM) to hindcast the OLR anomaly signal of the MJO with comparable forecast skill to other modern dynamical methods and Klingaman et al. (2014) include both the Cavanaugh LIM and the results of the methods described below.

The goal of this section is to illustrate to workings of LIM and LR, explore the linear predictability of the MJO, and to make methodological comparisons. Our work builds upon the linear inverse methods that have previously been applied to other geophysical applications (e.g. Shin et al. 2010; Newman et al. 2003), and extends these by introducing channel randomization methods to eliminate number-of-channels dependence and make skill tests truly comparable. In this way, the information value of non-orthogonal but non-redundant channels (time series) can be evaluated cleanly. Wheeler and Hendon’s (2004) MJO index is based on an EOF analysis using tropical belt averages of u850, u200, and OLR. For familiarity, our LIM uses time longitude sections of those same tropical fields.
3.2. MJO data and Methodology

3.2.1. Data

Daily outgoing long wave radiation (OLR) observed from satellites (Liebmann and Smith 1996) along with zonal wind observations at the 850hPa and 200hPa levels derived from the NCEP-NCAR Reanalysis project (Kalnay et al. 1996) between 15S to 15N from 1979 – 2011 are used in this study. Each variable was averaged from 15S to 15N while maintaining an initial 2.5 degree spatial resolution along the zonal direction. A 25 year composite annual cycle (1979-2004) was removed from each variable to produce anomalies. A 120 day prior mean was subtracted to remove low frequency signals, following Wheeler and Hendon.

3.2.2 LIM

For each LIM considered, a multivariate anomaly state vector was built from daily data: A time-longitude section, with repeated longitudes if u850 and u200 are applied to OLR. The time evolution of this system is modeled as a linear stochastically forced model in the form of equation 3.1. The evolution of the anomaly state vector that LIM predicts, \( \mathbf{x} \), is assumed to follow:

\[
\frac{d\mathbf{x}}{dt} = \mathbf{Bx} + \text{noise} \tag{3.1}
\]

where \( \mathbf{B} \) is an \( m \times m \) dynamical operator matrix. How can we estimate \( \mathbf{B} \) from data? Following Penland and Sardeshmukh (1995), if the noise is white and Gaussian, then the expectation value solution to (3.1) takes the form of (3.2) which provides a forecast at any given lead time (\( \tau \)) from an initial condition \( \mathbf{x}(t) \):
\[ \dot{x}(t + \tau) = \exp(B\tau)x(t) = G_\tau x(t) \]  

\( G_\tau \) is known as the propagator matrix that evolves initial anomalies, \( x(t) \), forward by any desired lead time (\( \tau \)). The dynamical operator matrix, \( B \), is found by solving (3.2):

\[ B = \left( \frac{1}{\tau} \right) \ln[C(\tau)C(0)^{-1}] . \]

\( C(0) \) and \( C(\tau) \) are the observed simultaneous and \( \tau \)-lag covariance matrices for \( x \). All linearly predictable dynamical interactions among the system variables are represented in the dynamical operator, also known as a deterministic linear feedback matrix or the system sensitivity matrix (Shin et al. 2010). All nonlinear system dynamics are approximated by the stochastic forcing or white noise that is uncorrelated in time with the linear system, although it is correlated in “space” (i.e., across the time series of variables that make up the columns of the state vector, \( x \)). Specific nonlinearities, if postulated to be important, can be included by using the square or cube of a state vector column as another state variable. With channel randomization (below), we can evaluate the information content or information value in such a postulated channel in a clean way.

Lagged regression and LIM predictions are constructed using both single-parameter (OLR only) and multi-parameter (OLR, u850, and u200) state vectors. The LIM trains on data from 1979 - 2000 and is verified from 2000-2011 so that hindcasting is not influenced by its own training data, thus eliminating a source of artificial skill (DelSole and Shukla 2009). Seasonal masks are also optionally applied to the training period to seek an optimal LIM construction for verification in that specific season. The score to be minimized is a global sum of squared OLR’ hindcast errors. One baseline of
hindcast skill is comparing the forecast error to the global climatological variance, which is the skill of a forecast of zero anomaly every day (climatology used as a forecast).

In abstract terms, LIM works on a set of time series or signal “channels” with no special ordering or meaning. Traditionally, principal components have been used to maximize variance represented. Shin et al. (2010) is one exception, valuable for its clarity of meaning. Here we also make our channels spatial and furthermore ordered: they are adjacent longitudes. This is most like Hakim (2013) who use radius-time sections in a tropical cyclone. Each channel consists of a time series of more than 7,000 daily observations so that the full 72 channel LIM uses 504,000 daily observations.

3.3 Skill score sensitivities to changes in LIM

It is helpful to define a LIM baseline or control case: all 3 variables, in 15 degree longitude bins, using $\tau = 2$ days. Each of the 24 longitude bins contain three channels consisting of a daily time series of anomalous MJO index variables (OLR, $u_{850}$, and $u_{200}$). Thus, the baseline LIM has a total of 72 input channels, 3 for each longitude bin (Figure 3.1).
For clean comparisons to this baseline, we need a score that can be repeated for reduced channel selection (e.g. OLR only). For this reason, we choose to score the hindcasts based on 24 OLR bins only. When other channels (u850, u200) are used, their impact is evaluated only in terms of the OLR prediction skill. Likewise, when additional longitude fine structure is included, we score its effect only on 15 degree scale OLR.

LIM is designed to approximate observed lag covariability at a range of selected lag times. The ‘‘tau test’’ (Penland and Sardeshmukh 1995; Newman et al. 2009) asks whether LIM is able to reproduce similar observed lag covariance statistics at all lags other than which the LIM is trained. To pass the tau test, and to confirm that the system dynamics are a stochastically forced linear system, LIM should reproduce similar lagged covariance statistics at all lags given by equation 3. The 72 channel LIM described above
exhibits similar hindcast skill score for the first four lags ( $\tau = 1, 2, 3, 4$), which indicates that our LIM passes the tau test and implies that variations in OLR can be approximated as a stochastically forced linear system (figure 3.2).

![Normalized OLR skill test tau0 < 5 values](image)

Figure 3.2. Normalized global hindcast skill score for a multivariate 72channel LIM built with the first four tau0 values. $\tau = 1$ (red), 2 (blue), 3 (green), 4 (grey).

### 3.3.1 Variations in input channels

The impact of state vector input channel quantity on skill score is examined by doubling the number of longitude bins from 24 to 48, which increases the zonal resolution of each longitude bin from 15 to 7.5 degrees, doubles the number of channels from 72 to 144, but it quadruples the number of coefficients in G and B matrices estimated from the doubled data. Doubling channel quantity has an undetectable effect on 15 degree OLR anomaly skill score (figure 3.3).
Figure 3.3. Normalized global OLR hindcast skill score of a 72 channel (red) and 144 channel (blue) LIM with the range of individual skill score for each longitude bin of the 72 channel (pink shading) and 144 channel (blue shading) LIM. Mauve shading occurs where pink and blue shading overlap.

Eliminating wind information from the anomalous state vector reduces the number of LIM input channels. Excluding u850 and u200 from the LIM serves a dual purpose. In addition to testing the sensitivity to a reduction in state vector channels, it also provides physical insight towards the impact of anomalous winds on the prediction of anomalous OLR associated with the MJO. For our LIM, the exclusion of wind information results in a 24 channel LIM. The 24 channel OLR-only LIM has a lower OLR global skill score relative to the 72 channel baseline LIM described above (figure 3.4). The difference in global skill score diminishes beyond a two week lead time.
Figure 3.4. Normalized global OLR hindcast skill score of a multi-parameter all MJO index 72 channel (red) and a uni-parameter OLR only 24 channel (blue) LIM with the range of individual skill score for each longitude bin of the 72 channel (pink shading) and 24 channel (blue shading) LIM.

Does the beneficial information provided from the zonal winds arise from the zonal mean only or from the zonal component at each longitude? A 26 channel LIM built from an anomalous state vector using OLR for each longitude bin and the zonal mean zonal winds at both the 850hpa and 200hPa level has a higher global OLR anomaly hindcast skill score compared to a LIM using OLR anomaly alone (figure 3.5). The skill resulting from the 72 channel LIM that uses all of the MJO index variables at each longitude is even better than the 26 input channels that use only the zonal mean of the zonal winds.
Figure 3.5. Normalized global OLR hindcast skill score of an OLR only 24 channel LIM (red), a multivariate all MJO index 26 channel LIM (black), and a multivariate all MJO index 72 channel LIM (green). The 26 channels represent OLR at each longitude (24 channels) combined with the zonal mean 850 zonal wind (1 channel) and the zonal mean 200 zonal wind (1 channel). The 72 channels represent u850, u200, and OLR at each of the 24 longitudes.

These 48 wind input channels are randomized in the multi-parameter 72 channel LIM (control LIM) to compare the effect of observed winds against 48 channels of randomly ordered data with identical values. Differences between the hindcast skill score from the control LIM and the LIM with scrambled wind channels isolates the information value inherent in using actual winds. Zonal winds at each longitude for the control LIM were scrambled in time to produce a randomized zonal wind time series. The control LIM exhibits higher OLR anomaly hindcast skill score up to two weeks lead time compared to the same LIM with randomized wind channels (figure 3.6). Randomizing data channels demonstrates that observed winds and not just increased data channels improves the LIM prediction skill.
Figure 3.6. Normalized global OLR hindcast skill score of the multi-parameter 72 channel LIM (red) and the multi-parameter 72 channel LIM with randomized wind channels (black).

We can further examine the value of wind information by comparing three independent LIMs built with a) full wind observations at each longitude bin, b) uniparameter OLR only, and c) zonal mean zonal wind channels for u850 and u200. At 10 days lead time, the LIM built with the full wind information at each longitude bin has the lowest hindcast prediction error, while the LIM built with zero wind information has the greatest hindcast prediction error (figure 3.7). The full wind LIM is 1.5% better than the LIM built with zonal mean wind information and is 2% better than the LIM that omits wind information.
Figure 3.7. Normalized global OLR hindcast skill score of the multi-parameter 72 channel LIM (grey), the multi-parameter channel LIM that substitutes the full wind information with the zonal mean value (green), and the uni-parameter LIM with zero wind channels (blue).

3.3.1.a Seasonal dependence

Understanding how OLR hindcast skill score varies over seasons may help us identify months of increased predictability. It is of interest to understand if a difference exists between summer and winter MJO events and if OLR anomalies in one season are more predictable than in the other. The control 72 channel LIM of figure 3.2 was trained on all days of the year from 1979 to 2000 and verified on all days of the year between 2000 and 2011. Separating the boreal summer (JJA) and boreal winter (DJF) training and scoring only reveals LIM skill score sensitivities for related seasonal hindcasts. OLR anomaly hindcast skill score improves during DJF training and scoring compared to the 72 channel all season LIM. In contrast, training the LIM on the JJA season reduces
hindcast prediction skill (figure 3.8). A 0.85 OLR anomaly hindcast skill score (normalized squared error achieved from JJA training and scoring) is achieved after a 3 day lead time for the JJA LIM, 7 days for the all season LIM, and 9 days for the DJF LIM. In relative terms for a 1-week lead time prediction, the control LIM has a 7% increase in OLR anomaly hindcast error compared to the DJF trained LIM while exhibiting an 18% decrease in error relative to the JJA trained LIM from a one week hindcast lead time. Like most of the other LIM variations, the seasonal OLR anomaly hindcast skill score tends to converge as it asymptotes to 1 after a 14 day hindcast lead time. Overall, the LIM is sensitive to training during winter and summer seasons. The LIM prediction error decreases during reduced DJF specific training and testing days versus a larger training set including all days from all seasons. In this case, it is preferable to train on less data, but on the proper season (DJF) rather than a using a larger set of training data including data from the wrong season (JJA).

![Figure 3.8. Global normalized hindcast skill score for a LIM trained and tested with DJF seasonal (red), JJA seasonal (blue), and all season (black) multi-parameter data.](image-url)
3.3.1.b Regional dependence

LIM skill can also be displayed at each longitude rather than as the zonally summed error (figure 3.9). Regional differences in correlation skill are evident. The highest correlation at all lead times is found in the region of the maritime continent between 110E - 130E. Here the correlation skill remains above +0.6 for 6 days and remains above +0.5 for 13 days. By contrast, the east Pacific region (longitude 230 in Figure 3.8) has the lowest one week hindcast correlation skill (r < +0.1). Summarizing figure 3.8, three hindcast skill hot-spot peaks are identified within the central Indian Ocean, the maritime continent, and central Pacific with areas of low predictive skill from the east Pacific to the Pacific coast of Central America.
Figure 3.9. OLR anomaly hindcast skill score for each of the 24 longitude bins from a three parameter 72 channel multivariate LIM.

3.4 Comparison between LIM and lagged linear regression

Lagged linear regression (LR) uses the dynamical operator equation,

\[ B = \left( \frac{1}{\tau} \right) \ln[\mathbf{C}(\mathbf{T}_0)] \mathbf{C}(0)^{-1} \],

...
curve based on a single fixed $\tau_0$ in equation 3.3. LIM applies the exponential fit calculated from the dynamical operator ($B$) for some fixed $\tau_0$ to create the auto-correlation curve from an initial value and some chosen lag value. In essence, an anomaly for LIM evolves as multiplication by an exponential fit to the historical auto-correlation rather than from multiplication of the auto-correlation itself at each lag. Global hindcast skill results from the LR technique are compared to the 72 channel 3-parameter LIM in figure 3.10.

![Graph of 24bin_72chan](image)

Figure 3.10. Global normalized hindcast skill score for the 72 channel control LIM (red) and lagged regression (blue).

Despite the differences between LR and LIM, they both share two characteristics for predicting the anomalous OLR signal from the MJO. Both methods identify an eastward propagation of a decaying anomaly signal over time (figure 3.11). Figure 3.10 illustrates
a period of an active MJO signal from late 1996 into 1997. The eastward propagating
OLR anomaly (shaded colors) is predicted by both LIM (solid curve) and LR (dashed
curve) methods. Longitudes ranging from the central Indian Ocean to the maritime
continent exhibit similar OLR anomaly predictions between LIM and LR because of the
high predictive skill found within that region. Differences between LIM and LR
predictions in areas outside of the high skill region may result from a non-exponential
structure to the autoregressive curve. If there were a systematic non-exponential structure
to the autoregression curve, LR would capture it and produce better forecasts. On the
other hand, LR is more subject to over-fitting and could have spurious correlations at
long lead times for example.

Figure 3.11. Observed OLR is plotted in grey scale. Positive (negative) anomaly contours
in red (blue) are 15, 10, 5 (-15,-10,-5) Wm^2. LIM (solid line) and LR (dashed line)
hindcasts were trained on data from 1979-2000, excluding 1996, using a 72 channel
multivariate state vector.
3.5 Discussion and conclusions

This chapter presents anomalous OLR hindcast prediction skill for a variety of LR and LIMs composed from daily OLR and optionally 200hPa and 850hPa zonal wind from 1979 to 2011. Results show some positive prediction skill (relative to climatology) up to three weeks. These results are consistent with results from Cavanaugh et al. (2014) that show a positive forecast skill for OLR anomalies up to three weeks from a LIM built with a reduced EOF analysis, and Klingaman et al. (2014) shows that this approach is more skillful for the Year of Tropical Convection intercomparison case.

The dependence of skill score on changes in the dynamical state vector was evaluated, using a clean comparison technique. LIM predictive skill is not necessarily sensitive to the quantity of input channels. Appending randomized channels also shows that channel number is not excessive in the range of 24-72 as used here. Using 15 degree longitude bins for the 3 parameter LIM results in a 72x72 lagged covariance matrix consisting of 5184 coefficients fitted to a training period between 1979-2000 (7670 days or 7670 degrees of freedom). The number of channels must not exceed the square root of the number of data values (daily observations). In other words, the number of elements in the lagged covariance matrix must not exceed the number of training days for the model. In this case, 72 channels do not exceed the square root of 7665 training days. If we used 12 degree longitude bins rather that 15, then the total number of channels for our multi-parameter system would exceed the square root of the number of training days and result in possible over-fitting.
Fine-scale information derived from doubling longitude resolution does not impact the global LIM hindcast skill score, suggesting it does not get over-fitted. Therefore, the overall quantity of longitude bins is not a principle factor towards maximizing the LIM skill score. On the other hand, LIM prediction skill is sensitive to variations in data quality, or the information contained within each channel that describes a dynamical system. For example, a one parameter anomalous state vector, OLR only, reduces the information content and is a poorer predictor of the anomalous OLR signal. Much of the value of wind channels lies in the zonal mean zonal wind. Evidently, clouds alone are not complete at describing the predictable features of clouds.

Additional predictive skill score sensitivities are revealed from adjusting the LIM training seasons. LIM hindcast prediction error is lowest during DJF seasonal masking and in regions of high total variance.

Adding state vector channels has a negligible impact on LIM skill, but can negatively impact lagged regression skill due to possible over-fitting. If the training data are few enough the lagged covariance matrix is noisy (i.e. more coefficients than days in historical training data), LR would use the noisy structure in its forecasts while LIM would fit a smooth curve that might actually reflect the reliable part of the structure just as well, or better. This example of over-fitting due to excess input channels can be seen in figure 3.12. The lowest prediction skill score at all lead times is from a multivariate LR using 48 longitude bins (144 channels). Unlike LIM, multivariate LR prediction skill is noticeably worse due to possible over-fitting when increasing the zonal resolution of the longitude bins from 15 to 7.5 degrees.
Results from a thoroughly tested LIM allow me to now construct a similar “space” based LIM to predict ingredients related to the evolution of Caribbean rainfall anomalies. Physical reasoning rooted in chapter two will inform my decision in the following chapter to select a state vector that best represents the ingredients that govern Caribbean rainfall anomalies.

Figure 3.12. Global OLR anomaly hindcast skill score for both LIM (reds) and LR (blues) from both a 72 and 144 channel combination. 72 (144) channel models are represented in darker (lighter) shades.
Chapter 4. APPLICATIONS AND PREDICTION OF CARIBBEAN EARLY SEASON RAINFALL

The dryness during the MSD is known to exert a socio-economic impact on farmers throughout the Caribbean (Magaña et al. 1999; Gamble et al. 2010; Campbell et al. 2011). However, variations in regional impacts and response related to mid-summer drying are due to the non-uniform strength of the MSD across the Caribbean. Some progress has been made to understand how farmers in the Western Caribbean respond to the strong MSD signal, but attention towards farmers in the Eastern Caribbean, where the MSD signal is weaker, remains sparse. Understanding the differences in potential crop impact from the intra-seasonal rainfall variability throughout the Caribbean is helpful for designing relevant and area specific rainfall forecasts. Integrating farmer perceptions and responses to changes in rainfall are a vital component to successful agro-climate services.

4.1 Field activities

Campbell et al. (2011) examined the impact of the MSD on Jamaican farmers through an extensive interview process and concluded that local subsistence farmers in Jamaica prepare for the MSD in advance because they expect drying to occur every year in July. Farmers in St. Elizabeth, the bread-basket of Jamaica, take measures to minimize economic loss related to the MSD drying by preparing their fields immediately after the end of the ERS rains. A popular method to do this involves layering the fields with guinea grass during early July to reduce soil moisture evaporation. In addition, farmers also prefer to buy water before the MSD for mid-summer irrigation. However, a farmer’s ability to purchase extra water often depends on the economic success from their early season crops. Heightened income from successful crop performance during the late
spring increases the opportunity to acquire additional resources (water, fertilizer, etc.) to offset the vulnerability of July drying.

Approximately 30% of the 282 farmers interviewed responded that July is the most severe dry month in terms of potential crop damage. Only half as many farmers suggested that the dry winter months were the most severe. Farmer perception of rainy months is bimodal and matches the data record of rainy months with peaks in the late spring and late summer. The close connection between farmer perception of monthly rainfall patterns and the observational time series builds confidence towards developing tailored forecasts to support farmer’s needs.

Allen et al. (2010) described the bio-physical response of the MSD on vegetation in Jamaica and concluded that a 2-4 week lag exists between mid-summer drying and a decrease in remotely sensed NDVI in Jamaica. This study quantitatively examined lagged vegetation response related to the MSD in the Caribbean and physically verifies farmer perception about July drying and its adverse response on vegetation. These results are important because they provide a reference to mid-summer drying and an associated vegetative response in the Caribbean. But, the examples described above represent only one island in the Western Caribbean and may not serve as a general representation for the entire basin.

In 2011 I conducted a field experiment in the Eastern Caribbean that investigated the theme of the MSD and farmer perception in the windward islands. The bulk of my research was based on the island of Dominica where local farming comprises a large portion of their economy. My motivation to interview farmers in Dominica was to discover if their reactions to the MSD were similar to farmer reactions from Jamaica. My
hypothesis was that farming response to the MSD along the eastern Caribbean was mild compared to the response in the western Caribbean because the overall difference between early season rains and mid-summer rains in the eastern Caribbean was smaller (fig. 4.1).

![Figure 4.1](image)

Figure 4.1. Annual cycle of observed monthly rainfall averages for Camaguey, Cuba from 1950-1995 (red), Kingston, Jamaica from 1950-1995 (blue), and Melville Hall, Dominica from 1969-2010 (green).

The Center for Latin American Studies and the Marine Science Graduate Student Organization at The University of Miami provided funding to test my hypothesis. With their support, I conducted 26 farmer interviews throughout Trinidad, Barbados, and Dominica during August 2011. The majority of my responses, 20, were recorded from Dominica.

All interviewed farmers in Dominica were asked what month was the most difficult for potential crop damage and if they sensed any difficulties or uncertainties for mid-summer growing. Every farmer unanimously stated that the MSD does not impact their crops; which is in stark contrast to the responses gained from Jamaica. I concluded through farmer interviews from the agriculturally productive Calibishie region in
Dominica that mid-summer rainfall is sufficient to offset farmers’ vulnerability to the MSD on the island. The field results in Dominica combined with previous results from Jamaica allow me to reject the broad generalization that the MSD uniformly impacts farmers throughout the Caribbean.

Despite the vulnerability to the MSD in the western Caribbean, dialog with Jamaican farmers reveals that they are reluctant to adopt any sort of mid-summer rainfall forecast due to the high historical risk of mid-summer farming. They anticipate a dry mid-summer and have adapted to avoid the low rewards of farming during July. A MSD forecast for Jamaican farmers has no perceived utility. Instead, Jamaican farmers highlight a strong need for an early rainfall season forecast since rainfall is likely and farming is productive during that period. Any chance to exploit an agriculturally productive period to grow more crops is a welcome opportunity that all of the farmers embrace. Knowing more about early season rainfall onset, duration, and forecasted early season rainfall totals could help farmers maximize their profits. This is in contrast to the MSD period when profits are unexpected and unlikely.

I returned to Dominica in 2012 to test my theory that an MSD forecast may be undesired by farmers in the East Caribbean. Earlier discussions with the Dominican National Meteorological Service described a new initiative to develop an agro-climate division that would strengthen the national economy by supporting farming practices. To assist this goal, I administered a climate data analysis and visualization workshop with members of the Dominica Meteorological Service. As suspected, their agro-climate division did not feel the need to mention the MSD during training or development. This
contrasts with previous experiences with the Jamaican Meteorological Service who identified the MSD to be a significant agro-climate threat.

I further examined regional variations in Caribbean climate during the “2012 Capacity building workshop on Data Rescue and Climate Change Indices: A Contribution to the implementation of the Global Framework for Climate Services in the Caribbean” in Kingston, Jamaica. Results from the workshop state that since 1960 all regions in the Caribbean are experiencing significant warming trends. In addition, despite regional variations in the MSD, all regions in the Caribbean are experiencing a significant increase in rainfall since 1986 and extreme rainfall since 1990 (Stephenson et al. 2013). As part of the “Expert Team on Climate Change Detection and Indices” I met representatives from all Caribbean national meteorological stations (Dominica being the sole absentee) and discussed their agro-climate needs related to forecasting mid-summer drying. Anecdotal evidence reinforced previous experiences that conclude mid-summer drying more heavily impacts agriculture throughout the Western Caribbean than the Eastern Caribbean. However, this evidence does not suggest that agricultural vulnerability to overall anomalous seasonal drying is divided between east and west. In addition, discussions from workshop participants further echoed the need to better understand early season rainfall dynamics and its predictability. The capacity building workshop along with results from farmer interviews shifted my initial interest in forecasting sub-seasonal rainfall variability from the mid-summer and towards the ERS when hydrometeorological forecasts have more utility across the Caribbean.
4.2 The rainfall time series in Jamaica and Dominica

The cause for different perceptions of mid-summer rainfall between the two islands can be gleaned from analyzing station rainfall records. Rainfall in the two locations is triggered by two different mechanisms. Jamaica lies within the Caribbean rain-belt and is susceptible to the rainfall dynamics discussed previously, while Dominica’s rainfall pattern is forced more by orographic uplift of the moist trade winds over the interior mountains (Smith et al. 2012). A rain-shadow on the drier western side of Dominica results from the trade wind forced orographic uplift in Dominica (figure 4.2).

Farmer response may also be different between the two islands because of the difference in intra-seasonal variability in rainfall around each island. The standard deviation of ERS rainfall is much greater around Jamaica than Dominica, which may suggest that rainfall is more consistent and expected throughout the ERS in Dominica (figure 4.3).

![Figure 4.2](image1.png) Figure 4.2. Monthly observed rainfall climatology from Melville Hall (green) and Canefield (orange) on the island of Dominica. (1982-2010)

![Figure 4.3](image2.png) Figure 4.3. The standard deviation of rainfall during the Early Rainfall Season calculated from 1998-2012.
4.3 Predicting the early season rainfall for the western Caribbean

Here we focus on predicting the ERS rains in the western Caribbean rather than the eastern Caribbean for a number of reasons. For one, the western Caribbean is more susceptible to sub-seasonal rainfall variability during the late spring. Planning for crops is more challenging for western Caribbean farmers when rainfall is more irregular or erratic. By contrast, farmers in the eastern Caribbean benefit more from regular or more expected rainfall during the ERS and into the mid-summer. In addition, detrimental mid-summer drying can be combatted more in the western Caribbean from farming success during the ERS, whereas mid-summer drying has little impact in the eastern Caribbean and thus no emphasis exists there to mitigate it. Moreover, farmers in the western Caribbean rely upon the early season rains for the success of the crops whereas the reliance on a single month of rainfall is less for eastern Caribbean farmers because rainfall is equally distributed during the entire rainy season in the eastern Caribbean. Insufficient rainfall during the ERS for a Jamaican farmer is difficult to overcome for the remainder of the growing season while insufficient rains during the ERS for Dominican farmers can be overcome by regular mid-summer rainfall. For these reasons, progress on rainfall prediction aims to benefit the western Caribbean the most and thus, is the primary focus here for a regional ERS rainfall forecast.

4.4 A LIM to predict western Caribbean OLR anomalies

Previous approaches have been applied to predict rainfall anomalies over the Caribbean (Ashby et al. 2005; Stephenson et al. 2007; Taylor et al. 2013). I build on this work by using the LIM method to focus on weekly prediction in order to model the frequency of weather events that produce the time average Caribbean rain-belt.
The LIM anomaly state vector, \( \mathbf{X} \), is built from 5 prognostic variables, or channels, that are hypothesized to be responsible for the evolution of the Caribbean rain-belt pattern. These 5 channels are extracted from the Caribbean rain-belt conceptual model from chapter 2 and represent the dynamical system whose linear evolution is postulated to be relevant to the Caribbean rain-belt time-space pattern. The evolution of the dynamical system is given by (equation 4.1) where the anomaly state vector is provided by (4.2).

\[
\frac{d\mathbf{X}}{dt} = \mathbf{Bx} + \text{stochastic forcing.} \quad (4.1)
\]

\[
\begin{bmatrix}
\text{SST} \\
\text{OLR} \\
u_{850} \\
u_{200} \\
v_{200}
\end{bmatrix}
\]

(4.2)

The LIM channels are built from 5 area average historical daily anomaly time series from 1982-2010 and are tested from 2011-2013. The area average for each channel is calculated from a specific region detailed in chapter 2 that captures the effect of that channel on the Caribbean rain-belt (fig. 4.4 a-b). For example, the 850hPa zonal wind channel is produced from an averaged area from the southern Caribbean where moisture is advected from the tropics. Negative \( u_{850} \) anomalies in this region may indicate periods of strong horizontal pressure gradients that favor moisture advection from the tropics into the Caribbean. The OLR time series is averaged throughout the western Caribbean in order to capture the area that encompasses the Caribbean rain-belt region. 200hPa zonal wind is averaged throughout the northern Caribbean and Gulf of Mexico where jet streaks associated with late season polar fronts and the subtropical front are prominent, while \( v_{200} \) is averaged over a region where upper level troughs are observed. “Wavy-
“ness” or oscillations of the meridional wind component indicate areas of propagating Rossby waves with associated troughs (Enomoto 2004). Lastly, SST is averaged in the western Caribbean where spring warming first appears and persists throughout the summer. Spatial averaging of each channel helps to coarse grain the anomaly state vector so that it can be linearly represented by equation 4.1 (Newman et al. 2003).

Figure 4.4a. Averaged areas for each channel parameter in the Caribbean LIM.
Figure 4.4b. Averaged area anomaly daily time series for each channel parameter in the Caribbean LIM.

Interactions between each channel of the dynamical system are evaluated from the linear dynamical operator matrix \( \mathbf{B} \) from equation 4.3 (figure 4.5).

\[
\mathbf{B} = \left( \frac{1}{\tau} \right) \ln[\mathbf{C}(\tau_0)\mathbf{C}(0)^{-1}] 
\]  

Recall that the \( \mathbf{B} \) matrix is constructed from the observed simultaneous and lagged covariance statistics of each channel from the LIM. The diagonal values along the dynamical operator matrix have units of inverse time (in this case the unit is days\(^{-1}\)) and off diagonal values have units of the tendency of one channel and the anomaly units of another. For example, in our LIM the influence of SST on OLR is 4.23 Wm\(^{-2}\) per day of
OLR tendency per Kelvin of SST anomaly. The diagonals of the dynamical operator are negative and imply an internal dampening effect, which is intuitive considering the autocorrelation of each channel decays in time (figure 4.6).

Figure 4.5. The dynamical operator matrix, $B$, calculated from the Caribbean LIM. The real part of $B$ is printed in red and is shaded in grey scale to show relative values.

Figure 4.6. Autocorrelation as a function of lead time for each of the Caribbean LIM channels: SST (a), OLR (b), $u_{200}$ (c), $u_{850}$ (d), $v_{850}$ (e).
The time lag autocorrelation of the SST channel suggests that it is the least damped of all 5 variables in the state vector. The dynamical operator confirms that SST is indeed the least damped of the 5 channels because the simultaneous and lagged covariance statistics of SST (equation 4.3) is the highest (-.04) along the diagonal of the dynamical operator matrix. On the other hand, the value along the diagonal for OLR is the lowest (-0.26), which is expected given the faster decay rate of the OLR 21 day lagged autocorrelation. Therefore, the diagonals of the dynamical operator matrix can be used to assess the rate of decay of the historical autocorrelation for each channel. But, more importantly, information on predictability can be gleaned from the dynamical operator matrix. Highly damped variables tend to lose prediction skill faster than variables that are weakly damped. The rapid loss of prediction skill indicates that either the non-linear forcing term of the time evolution of the system is large or that other descriptive channels are missing from the state vector.

Shin et al. (2010) diagnose the dynamical operator matrix by stating that columns of the dynamical operator matrix can be interpreted as having the “influence of” the particular column variable. The elements along first column of the dynamical operator matrix from figure 4.5 are interpreted as the influence of the time tendency of SST anomalies on the variables from the rows of the matrix. Changes in SST have the highest impact on changes in anomalous OLR (+4.23) within our dynamical system represented by our 5 channel anomalous state vector. The lowest values found in the dynamical operator indicate that none of the wind channels influence SST anomalies or that SST anomalies evolve from forcing from channels not included in the LIM.
Physically interpreting the off diagonal dynamical operator elements with each other is difficult because a state vector built with mixed physical quantities does not have common units. For example, comparing $\text{Wm}^{-2}/\text{day}$ per K of SST anomaly with $\text{ms}^{-2}/\text{day}$ of $\text{Wm}^{-2}$ of OLR anomaly does not reveal physically meaningful information. Therefore, elements in the same column should be compared in order to make meaningful statements regarding the effect of one variable on another. An alternative method to intercompare off diagonal elements from the dynamical operator matrix is to standardize each channel so that each element of the dynamical operator is unit-less. Standardizing the anomaly channels causes the dynamical operator matrix to have the same relative pattern, but with different values. The dynamical operator matrix built from standardized anomaly channels confirms SST to be the most damped LIM channel and shows that SST anomalies have the highest influence on OLR anomalies.

The dynamical operator matrix highlights a strong interaction between OLR and SST relative to other channel interactions. To test whether or not this high value was true, the dynamical operator matrix was recalculated multiple times using a combination of different SST regions. Average area SST calculated from multiple area combinations also determines whether or not the influence of SST on OLR is uniform throughout the Caribbean (figure 4.7). The results from spatially varying the anomalous SST channel show that SST anomalies influence OLR anomalies the most in the western Caribbean. SST anomalies have only a minor influence on OLR anomalies throughout the east Caribbean and the Gulf of Mexico. The dynamical operator from the Caribbean channel LIM confirms that observational estimates of simultaneous and lagged covariance statistics of anomalous SST and OLR show that SST anomalies have the highest impact.
on OLR anomalies where the SST is warmest. Analysis of the dynamical operator also shows that the upper level circulation (u200 and v200 channels) is the least influenced by SST, which suggests that SST anomalies have little influence on the appearance of propagating Rossby waves and upper level troughs in the Caribbean.

![Figure 4.7](image)

Figure 4.7 Various calculations for the influence of multiple SST anomaly area averaged regions (colored outlined boxes) on OLR anomalies from a single area averaged region (shaded rectangle with dashed outline).

### 4.5 LIM prediction skill

Analyzing hindcast skill score is necessary because prediction skill is an important test of any model’s usefulness as a forecasting tool. Testing the LIM supports the hypothesis that variability in the Caribbean rain-belt pattern can be approximated as a linear system defined by the time evolution from the dynamics of the 5 channel anomaly
state vector. Following equation 1, the time evolution of the 5 channel anomaly dynamical system is approximated by

\[
\frac{dx}{dt} = B (x, t) + F
\]

Forecasts of the state vector, \( x \), are made as

\[
x(t + \tau) = \exp(B \tau) x(t)
\]

Hindcast skill score for each channel is calculated as a function of lead time (days) over a 10,585 day training period and is tested over 1,095 days. The LIM is seasonally independent in the fact that it is trained on all days of the year. However,
predictive skill may exhibit seasonal dependence as evident from the previous longitude space LIM (figure 8.3). The hindcast skill score is calculated from two methods, error squared variance and correlation between hindcast (prediction) and verification (observation) from the 2011-2013 verification period. Skill score results indicate that initial error growth is the slowest for SST and is fastest for OLR (figure 4.8). The LIM predicts SST anomalies with positive prediction skill, relative to climatology, throughout the entire 21 day lead time hindcast period. OLR prediction skill loses all prediction skill after only 3 days. Positive LIM prediction skill for the remaining 3 variables (u200, u850, and v200) falls off after one week of lead time. For LIM, positive prediction is simply the measure of whether the signal from each channel is large enough to be distinguished from the noise of the system at daily scales with $\tau_0 = 0$.

The autocorrelation of each LIM channel input further demonstrates the inherent predictability of each channel and can serve as a measure of predictability. Recall from figure 4.6 that the autocorrelation for OLR drops to zero around 10 days of prediction lead time while the autocorrelation of SST remains relatively high at +0.45 after a three week lead time period.
By construction, a LIM models the interaction of its channels as a stable, linear system that dissipates in time and is driven by noise that is white in time, but not necessarily in space and across parameters (i.e. the noise can be correlated across channels). The Caribbean channel LIM successfully predicts a dissipative process in which most often an observed anomaly decays to zero over time. The anomaly decay rate modeled by LIM can be strongly dependent on the initial condition. If the initial conditions are correlated with, or project strongly onto, an optimum structure of the state matrix, then predictions may be more skillful over a longer period than forecasts initialized at other times and may include anomaly growth rate. Model error can be described as a measure of the system nonlinearities or as an incomplete dynamical state vector that may not best represent the total system dynamics. Seven day OLR LIM
hindcasts from the 2011-2013 validation period are shown in figure 4.9.

Figure 4.9. OLR anomaly time series for validation (black) and 7 day LIM hindcast (red) for 2011 (top), 2012 (middle) and 2013 (bottom).

Each 7 day prediction is initialized every seven days so that the end of one 7 day hindcast period is immediately followed by the beginning of another. As expected, the LIM anomaly prediction decays to zero over the hindcast period with few exceptions. Producing a 7-day OLR LIM hindcast at every day during the ERS shows greater detail of the anomaly predictions and their dissipation (figure 4.10). The slope of the LIM anomaly hindcast describes the rate of anomaly decay. A LIM hindcast with a flatter slope relative to one with a steep slope may suggest the persistence of an anomaly while a LIM hindcast with a steeper slope may indicate a more rapid decay of the anomaly. The
slope of the LIM hindcast can provide utility in an operational sense so that the persistence of the observed anomaly over time may be better understood.

Figure 4.10. Same as figure 4.9 except for the Early Rainfall Season. 7 day LIM hindcasts are initialized at every day during the Early Rainfall Season.

The predictability of a observed anomaly from a LIM is dependent on the linear signal embedded within the historical anomaly time series. The predictable signal of the anomaly time series diminishes over time and eventually can no longer be distinguishable from the noise of the system. Predictability is lost when the noise and signal are of the same strength. The signal-to-noise ratio has been used in the past to assess seasonal predictability (Compo and Sardeshmukh 2004). For the LIM, the signal-to-noise ratio can be calculated by partitioning the state vector covariance matrix into a forecast signal covariance matrix, $F$, (equation 4.4) and a forecast noise variance matrix, $E$, (equation 4.5) (Pegion and Sardeshmukh 2011). The actual signal-to-noise ratio, $S$, is calculated as the square root of the ratio of the forecast signal variance to the noise variance.
F(τ) = diag[G(τ)C(0)G^T(τ)] \quad \text{(equation 4.4)}

E(τ) = diag[C(0) − G(τ)C(0)G^T(τ)] \quad \text{(equation 4.5)}

\[ S = \sqrt{\frac{F}{E}} \quad \text{(equation 4.6)} \]

The signal-to-noise ratio is calculated from the 5 channel Caribbean LIM following the methods from Pegion and Sardeshmukh (2011) (table 2).

<table>
<thead>
<tr>
<th>Channel</th>
<th>Forecast Signal Variance (F)</th>
<th>Forecast Noise Variance (E)</th>
<th>signal-to-noise ratio (S)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SST</td>
<td>0.018</td>
<td>0.132</td>
<td>0.365</td>
</tr>
<tr>
<td>OLR</td>
<td>6.906</td>
<td>393.848</td>
<td>0.132</td>
</tr>
<tr>
<td>U200</td>
<td>0.715</td>
<td>87.789</td>
<td>0.090</td>
</tr>
<tr>
<td>U850</td>
<td>0.037</td>
<td>9.113</td>
<td>0.064</td>
</tr>
<tr>
<td>V200</td>
<td>0.345</td>
<td>35.790</td>
<td>0.098</td>
</tr>
</tbody>
</table>

Table 2. Signal to noise ratio calculated from the 5 channel Caribbean LIM.

The SST channel has the highest signal-to-noise ratio of all of the Caribbean LIM channels, which suggests that the averaged area SST anomaly channel contains the strongest linear signal compared to the other LIM channels. The highest signal-to-noise value for SST reaffirms the results from both the diagonal analysis of the dynamical
operator matrix and the lagged autocorrelation analysis of SST that suggest SST is the
most predictable of the LIM channels. The strength of the noise is greater than the
strength of the signal for each channel, which may indicate that our subjectively chosen
LIM state vector is dominated more by non-linear processes than from a linear signal.
The weak signal-to-noise ratio of each channel may prove to be problematic for the LIM
with respect to potential operational use, but despite the discouraging prognostic outlook,
LIM proves to be an effective method towards diagnosing the linearity and predictability
of a dynamical system.
Chapter 5. CONCLUSION

The initial motivation to pursue this dissertation spawned from applications oriented questions arising from the bimodal pattern of the Caribbean annual rainfall cycle. Work from other researchers lured me into believing that the bimodal pattern was not spatially uniform throughout the Caribbean and that the mid-summer drought may have regionally varying impacts. Focusing on the agricultural impacts of mid-summer drying upon various Caribbean farming communities proved to be instrumental towards developing and completing my dissertation research.

Conclusions related to both the physical and socio-economic impacts of the mid-summer drought were drawn from farmer interviews throughout the Caribbean. Farmers overwhelmingly stated that crop vulnerability to mid-summer drying is perceived to be the greatest in the western Caribbean and is relatively unnoticed by farmers in the eastern Caribbean. However, despite strong vulnerability to mid-summer drying, Jamaican farmers were for the most part uninterested in possible improvements to mid-summer rainfall forecasting. Instead, like the farmers in Dominica, the Jamaican farmers were most interested in forecast improvements during the more productive late spring growing season. These regional conclusions not only closed the loop on my initial research motivation, but also paved the way to isolating the early rainfall season (ERS) as the primary period of focus for the dissertation.

The ERS, characterized by the time average Caribbean rain-belt pattern, is defined from the climatological record of accumulated rainfall between May 15 to June
15. The Caribbean rain-belt appears as a quasi-stationary rain-belt during the ERS and is responsible for the first peak of the Caribbean annual bimodal rainfall cycle.

The Caribbean rain-belt is not a geographically unique feature; other seasonal rain-belt patterns are observed at similar latitudes. The onset and supporting dynamics of the Caribbean rain-belt were compared to the meiyu-baiu, a well-studied subtropical rain-belt along the western Pacific. The atmospheric dynamics that support rainfall in the Caribbean rain-belt are diagnosed from the quasi-geostrophic omega equation. Under this dynamical framework, positive zonal wind at 200hPa and 500hPa temperature advection were identified as two primary forcings for ascent within the Caribbean rain-belt. Moisture along the Caribbean rain-belt is modulated by SST and is supplied by advecting islands of high precipitable water from the tropics. The Caribbean rain-belt that precedes the emergence of the North Atlantic Convergence Zone can be viewed as the analog to the meiyu component of the meiyu-baiu rain-band.

The Caribbean rain-belt exhibits inter-annual variability in terms of onset, strength (amount of rainfall), duration, and location. In some years the Caribbean rain-belt may be zonally elongated or may be absent from the rainfall record entirely. Inter-annual variability of the Caribbean rain-belt impacts farmers who depend on the regular cycle of rainfall during the ERS. Outreach activities and capacity building workshops throughout the Caribbean revealed the need to predict ERS rainfall in an effort to minimize farmer’s vulnerability to inter-annual rainfall variability. To support this need, statistical rainfall prediction for the Caribbean was developed from a linear inverse model (LIM).
The dynamical system that the LIM models was built from state vector channels based on the Caribbean rain-belt conceptual model. Outgoing longwave radiation, a proxy for Caribbean rain-belt cloudiness, has positive LIM prediction skill up to about 3 days of lead time while the LIM wind channels have positive prediction skill up to about one week lead time. While the LIM built from area averaged anomalous time series data may not have instant operational forecast value, the LIM does provide meaningful information about the system dynamics of the Caribbean rainfall system during the ERS. The rates of internal dampening and dynamical relationships between each channel can all be deduced from analyzing the linear dynamical operator.

A surprising outcome from this research has been an alternative explanation for the onset of the Caribbean “mid-summer drought”. Suggesting a physical mechanism to explain mid-summer drying was never a primary objective of this research. However, the discovery of a northward shift of the forcings for ascent during the mid-summer provides a plausible theory for Caribbean mid-summer drying and the intensification of the North Atlantic Convergence Zone during July. Changes in the local circulation could play a larger role for the onset of the “mid-summer drought” rather than the influence of transient features entering the region.

The results from this dissertation provide insight into a regional rainfall phenomenon that is ripe for further research attention. It is critical to understand how the first peak of the bi-modal rainfall cycles develops before we can begin to understand its variability and how rainfall in the Caribbean may evolve under climate change. Future research activities can be directed to address the variability of the Caribbean rain-belt and
to refine statistical prediction methods to forecast either the rainfall related to the
Caribbean rain-belt or the actual position of the Caribbean rain-belt.

Research directed towards addition social applications can also be drawn from
this dissertation. Human health concerns related to summertime climate variability, like
dengue fever, represent an additional motivation to better forecast rainfall during the
summer. Dengue fever can thrive within the Caribbean when high surface temperature
and moisture are present (Amarakoon et al. 2007; Jury 2008). Anomalously high
precipitation during the hot mid-summer period produces ripe environmental conditions
that can favor a dengue fever outbreak. This occurred in epidemic proportions throughout
the Caribbean during the 2010 La Nina when July rainfall was anomalously high (A.P.
2010; CDC Dengue Update 2011). Predicting anomalous summer rainfall prior to July
can alert the medical field towards potential dengue outbreaks.

Anomalously high rainfall may also bolster environmental conditions that favor
landsides in areas with rolling to mountainous terrain. Many of the mountainous
Caribbean islands are enriched with clay soils that can easily become unstable after
extended rainfall. This instability coupled with topographic relief is a recipe for
destructive landslides that can ruin valuable farmland as well as destroy entire
communities. It is not uncommon for landslides to accompany heavy rains in the
Caribbean. From 2007 – 2012 there were 133 reported precipitation related landslides
within the Caribbean (NASA global landslide index).

Forecasting summertime rainfall anomalies could also support food security early
warning systems for populations that rely on subsistence farming. Anomalous rainfall and
vegetation conditions in Haiti for example are monitored through the Famine Early
Warning System (FEWS-net) for food security purposes. Food security early warning also contributes to national security interests by highlighting periods of potential civil unrest.

The results from this dissertation help to fill in the blanks towards explaining Caribbean rainfall while also identifying multiple topics worthy of additional research attention.


