A Quasi-Global Survey of Precipitation Extremes: Interpretations of Their Dynamics and Thermodynamics

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UNIVERSITY OF MIAMI

A QUASI-GLOBAL SURVEY OF PRECIPITATION EXTREMES:
INTERPRETATIONS OF THEIR DYNAMICS AND THERMODYNAMICS

By
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A THESIS

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A QUASI-GLOBAL SURVEY OF PRECIPITATION EXTREMES:
INTERPRETATIONS OF THEIR DYNAMICS AND THERMODYNAMICS

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Rainfall extremes are caused by a complicated interplay between moisture and dynamics with many challenges and ideas that have yet to be fully understood in the literature. This study examines the dynamics and thermodynamics of extreme precipitation events through a combination of observational data and reanalysis model outputs.

Data from the Tropical Rainfall Measuring Mission (TRMM) 3B42 product, version 7, were retrieved for the time period 1998-2016. The greatest single-day and three-day rainfall accumulations for each grid pixel in several different spatial grid resolutions were identified along with its associated date and time of occurrence. Global maps of these record events were created and statistics were computed over their worldwide occurrence. The mode of precipitation events occur for records of ~100 mm over a single day and ~160 mm over 3-day accumulations, a ratio that is much smaller than 1:3 even for the largest-scale events (4-degree grid cells). Much higher accumulations are observed but are quite rare as expected. The densest concentration of high precipitation accumulation events occurred over east Asia and the Caribbean.
extremes were also found to occur primarily during the local summer months of the northern and southern hemispheres.

Three distinct case studies over the Indian subcontinent were selected based on the global map of 4-degree, 75-hour record rainfall events. These events were selected such that no two events on the map coincided with one another. For each case, an intercomparison of 14 various daily rainfall products (both observation and reanalysis-based) assembled as part of the NASA-NEWS program shows challenges in quantitative estimation. However, the MERRA-2 reanalysis analyzed rainfall totals were comparable to the event-defining TRMM 3B42 data. Various two and three-dimensional dynamic and thermodynamic variables from MERRA-2 were visualized to explore the meteorological story behind these extreme rainfall cases. Two cases involved deep depressions that formed in the Bay of Bengal and subsequently made landfall over India during the 2006 North Indian Ocean cyclone season, while a third case was a quasi-stationary low-pressure system over central India in 2005. In all three cases, a mid-level vortex, defined by its potential vorticity (PV), was embedded in the monsoon shear flow. These sheared cyclones exhibited induced ascent on the downshear-left quadrant (southwest of the center in monsoon depressions). Because PV is centered in the mid-levels between the low-level westerlies and upper-level easterlies, these systems moved slowly which contributed to large rainfall accumulations at fixed locations.
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Chapter 1: Introduction

1.1. Motivation

Rainfall is an essential component to the hydrologic cycle and beneficial to agriculture. However, cases of extreme precipitation accumulations are as variable as they are complex and can have adverse effects on those affected because of the prominent risk of flooding. Flooding is defined in several ways; the American Meteorological Society (AMS) defines flooding as an “accumulation of water caused by surface runoff in low-lying areas not usually submerged” (AMS cited 2018). Flash flooding, regarded as a more rapidly developing event, is summarized by the World Meteorological Organization (WMO) as a rapid increase in water volume in otherwise dry areas “with a relatively high peak discharge” amid excessive rainfall, dam failure, or an ice jam (WMO cited 2018).

When considering extreme rainfall events, their occurrence does not merely rely on just high rain rates or total accumulation. The complexity lies within the interplay of various atmospheric, hydrologic, and topographic features that can vary greatly in space and time. In essence, these scales range from urban canyons and rivers to the larger half-continent catchment basins, each responding in a different manner to different meteorological forcings. Smaller meteorological scales involving the development of convective thunderstorms (e.g., pulse thunderstorms) may shadow any potential for generalized scientific thinking as the global atmosphere is not well-resolved at fine resolutions and can be difficult to predict beyond a single day. Larger-scale extremes involve the understanding and modelling of different hydrological feedbacks (i.e., evaporation, runoff, soil health;
feedbacks which can be non-local in nature) as rainfall is just one component to an event.

For these reasons, this thesis is motivated to examine cases of extreme precipitation within a moderate spatiotemporal scale (cases on the order of tens of kilometers in space; a day to a week in time). In the following chapters, various cases will be studied to survey the dynamic and thermodynamic components that contribute to extreme rainfall events. The principal questions that will be addressed in this thesis include the following:

1. What is the spatial distribution of extreme precipitation cases across different regions and seasons of the world? Are there more cases in specific regions of the world?

2. Are there similarities or differences from multiple case studies of precipitation extremes on such large mesoscales? Furthermore, do the observational and reanalysis products adequately resolve the location and intensity of precipitation adequately?

3. Are common ingredients for precipitation extremes identifiable, and if so, how can they be discerned or inferred upon based on their relative importance? In other words, are there specific dynamic or thermodynamic components that contribute the most to certain events?

In all, this work will attempt to elucidate some common forcing mechanisms contributing to extreme precipitation accumulations. These results will also serve as a platform for future studies that will be highlighted in the future works section of this manuscript.
A brief outline of the content in this thesis is as follows: Chapter 1 introduces the symbolic and important literature pertaining to rainfall extremes. Mechanisms from previous studies on extreme precipitation events are highlighted. Chapter 2 details the methodology in this thesis. The series of datasets and necessary tools that are consulted to perform the analysis are presented, and the rationale behind the data processing techniques is explained. Chapter 3 illustrates the results of quasi-global precipitation extremes where the greatest rainfall totals over a near-global grid are identified. Chapter 4 highlights a series of extreme precipitation case studies from the results obtained in chapter 3, focusing specifically on events over the Indian subcontinent for clarity of comparison. Lastly, chapter 5 offers a summary of this thesis as well as ideas, motivations, and open questions for future works.

1.2. Historical Contributions to Extremes and the Atmosphere

In a strictly meteorological frame, extreme precipitation events leading to flash flooding can arise from a combination of favorable mesoscale and synoptic-scale features and dynamics coinciding over a region. But before examining these features in any detail, definitions for both scales must be addressed. Meteorological features at the mesoscale are defined as having a spatial scale of $2 \text{ km} < x < 2000 \text{ km}$ and a temporal scale of minutes to a few days. However, this definition alone is not descriptive of some of the smaller-scale phenomena that occur within the mesoscale. Orlanski (1975) proposed that the mesoscale could be broken apart into three subclasses, each of which contains meteorological
features unique to them; these are the Meso-$\gamma$ $[O(2-20 \ km); \ e.g. \ thunderstorms]$, meso-$\beta$ $[O(20-200 \ km); \ e.g. \ mesoscale \ convective \ systems]$, and the meso-$\alpha$ $[O(200-2000 \ km); \ e.g. \ frontal \ boundaries]$. The synoptic scale contains features that are larger in space and longer in time, defined as having a spatial scale of $x > 2000$ km and a temporal scale of $O(1 \ week)$. While not an exhaustive list, some features associated to the synoptic scale include surface high and low pressure systems, extratropical cyclones, and upper-level troughs and ridges.

With these subscales, it could then be useful to classify various meteorological features based on their spatiotemporal scales. Extreme precipitation accumulations attributed to these features could become the building blocks for such case studies. But in an example of multiple case studies of similar systems, how can one deduce a set of common ingredients for extreme precipitation?

A classical framework that categorizes extreme rainfall and flash flood events based on their synoptic and mesoscale states was presented by Maddox et al. (1979). In their study, a total of 151 flash flood cases were sampled from 1973-1977 across the United States and were divided into different types of flash flood events based on the synoptic and mesoscale characteristics during each event. Despite the subtle quantitative differences in each of the cases, qualitative commonalities of favorable ingredients between all the individual cases were realized: 1) A lack of vertical wind shear through the troposphere; 2) total column water vapor (CWV) amounts greater than 37.5 mm and >150% of monthly climatology; 3) surface dewpoint temperatures greater than $18^\circ C$. These
ingredients are all fundamentally important in considering case studies of such extreme rainfall.

For extreme precipitation accumulations to materialize, there must be anomalously high precipitation rates for an extended duration of time over a finite area (Doswell et al. 1996) described as:

\[ P = \bar{R}D \]  

where \( P \) is the total accumulated precipitation and \( \bar{R} \) is the average precipitation rate, over a length of time \( D \). The duration of a precipitation event depends on a combination of the total coverage area of precipitation and the storm motion speed (Doswell et al. 1996), but it is important to note that the duration of an event can be poorly defined in cases where precipitation gaps may exist for moments in time. Furthermore, \( R \) can be decomposed to explain the relationship between essential dynamical and thermodynamic components used in describing total precipitation. For this reason, \( R \) can be shown as:

\[ R = Ewq \]  

where \( E \) represents the ratio between precipitation rate and water vapor flux into the cloud (or precipitation efficiency), \( w \) is the air parcel’s vertical velocity of ascent, and \( q \) is the water vapor mixing ratio of the ascending parcel. This upward flux of water vapor through the cloud base associated with \( w \) in (2) can be related to convective instability in balanced flows in single updrafts and/or secondary circulations (Raymond et al. 2015). Intuitively, the potential for a high \( R \) value may exist since \( R \) increases non-linearly as \( E \), \( w \), and \( q \) increase and thus, rainfall extremes are more likely in situations where there is fast vertical ascent of highly
saturated air parcels. Persistence of high rain rates are a pathway for the development of long probability tails in the overall distribution of precipitation (Over and Gupta 1994).

This ingredients-centric approach continues to provide an excellent framework in which to examine case studies. However, this approach is vague as no two extreme precipitation cases are exactly alike. Both the list of ingredients and the types of extreme precipitation cases have been further refined with more data obtained from satellite measurements and sophisticated model reanalysis outputs. Schumacher (2017) captures this approach and elucidates the various mechanisms and ingredients of extreme precipitation events throughout past and current research, including the criteria set forth by Maddox et al. (1979) for classifying extreme precipitation events. Even though detection of flash flood events has become easier with time, the predictability of such extreme events remains a challenge (Schumacher 2017).

1.3. Mechanisms Involved in Extreme Precipitation

1.3.1. Topographic Forcing

Forcing mechanisms and features involved in a case are important to understand, particularly the interactions of features between the mesoscale and synoptic scale. An example of this is land-air-sea interactions in the South Asian Monsoon. To demonstrate the general nature of the South Asian monsoon convective activity, Romatschke and Houze (2011) used eight years of data from the Tropical Rainfall Measuring Mission (TRMM; Kummerow et al. 1998)
Precipitation Radar (PR) to classify different types of convective systems pertaining to the South Asian monsoon and how these systems add to the precipitation budget over south Asia. After identifying all convective systems and classifying them based on their spatial extent and their precipitation intensity, this study determined that the Western Himalayan foothills are dominated by small-to-medium but highly convective systems which peak in development during the early morning or afternoon hours. Eight-year storm-type composites of height anomalies at both the surface layer and at 500-hPa are slightly negative in magnitude over the Himalayan foothills and positive for southern India. Upper-level low pressure systems provide enhanced synoptic-scale forcing as easterly flow transports high moisture content from the Bay of Bengal over the elevated terrain of northern India and Pakistan (Romatschke and Houze 2011).

In cases of anomalously favorable environmental conditions, sustained and severe convection can lead to flooding events much like the Indus River flood of July 27-30, 2010 in Pakistan. Studies by both Houze et al. (2011) and Galarneau et al. (2012) examined this event and found several mesoscale and synoptic scale features that contributed significantly to the persistent heavy rainfall over the affected areas. At 500-hPa, negative height anomalies associated with two different surface-based low pressure systems were detected over the Bay of Bengal and the Arabian Sea while a significantly strong 500-hPa anticyclonic blocking pattern was located between Russia and the Tibetan Plateau (as indicated by strong positive height anomalies), providing persistent and strong southeasterly flow to India and Pakistan (Fig. 1.1; from Houze et al. 2011).
addition, an upper-level jet located northwest of Pakistan had intensified to a maximum of near 70 m s\(^{-1}\). This was due to a combination of increasingly positive potential vorticity (PV) advection with respect to time along the cyclonic shear periphery of the jet. Conversely, negative PV advection was observed along the anticyclonic shear side of the jet suggesting that sufficiently strong synoptic-scale ascent was present for continued deep moist convection (Galarneau et al. 2012). The established anticyclonic blocking pattern advected deep tropical moisture of CWV values >70 mm from the Bay of Bengal eastward over the higher terrain of Pakistan by a persistent and robust low-level jet, providing sufficient topographic forcing via lower-level upslope flow for the length of the event (Galarneau et al. 2012). Rainfall accumulations for the period of July 28-30, 2010 in several provinces of northern Pakistan were well above 200 mm which were sufficient to eclipse monthly climatological precipitation values.

1.3.2. Mesoscale Convective Systems

Many extreme precipitation events can arise from weakly forced mesoscale convective systems (MCSs; Houze 2004). Primarily occurring during the warm season, MCSs are formed through the mechanism of upscale growth. Since each thunderstorm within an MCS has a corresponding updraft and downdraft, a “cold pool” of rain-cooled air develops in the wake. This cold pool acts as a pseudo frontal boundary which propagates opposite to the MCS main flow. The leading edge of this cold pool acts to trigger more convection as the differential heating is
strongest ahead of the MCS while older storms entrenched in the cold pool die out due to lack of instability.

MCSs are largely responsible for flooding events that occur in the tropics and subtropics. Two types of structures to MCSs are most prominent: training line/adjoining stratiform (TL/AS) and the backbuilding/quasi-stationary (BB) (Schumacher and Johnson 2005; Figs. 1.2a-b). The development of the former is characterized by a surrounding environment of high moisture content and instability on the relatively cooler side of a slowly propagating boundary (e.g., surface gust front), while the latter depends on the more immediate environment enveloping the system (e.g., instability fed via the wake of a cold pool). These two archetypes of MCSs are not standalone since they can change behavior and shape over the course of the storm lifetime.

Storm motion in either of these MCS types plays a pivotal role in the way that precipitation can accumulate over any given area. In some cases, the storm cell motion and the overall MCS propagation vectors nearly or do in fact cancel each other (Corfidi 2003). The consequence of this is the sustained training of thunderstorms over the same region in a short period of time. Predictability for MCSs continues to be a challenge for forecasters and researchers alike as the responsible forcing mechanisms for such systems tends to be weak compared to the overall system flow and not well understood in the literature.
1.3.3. *Tropical Cyclones*

Tropical cyclone (TC) landfalls are another way that flooding rains can occur over a region, especially if they are slow-moving. A recent example of this was Typhoon Morakot (2009) which brought devastating rain and wind to southern Taiwan. Chien and Kuo (2011) used a combination of reanalysis products from the Modern-Era Retrospective reanalysis for Research and Applications (MERRA, Rienecker et al. 2011), the European Center for Medium-Range Weather Forecasting: Year of Tropical Convection (ECMWF-YOTC), and observational datasets with archived best-track data from the Joint Typhoon Warning Center (JTWC) to examine factors contributing the most to the extensive flooding over southern Taiwan. Forward motion for 19 previous TCs with a similar trajectory were investigated for translation speed anomalies between Morakot and the other 19 storms. Their study showed that a combination of weakening in the upper-level steering flows and presence of Typhoons Goni and Etau to its west and east, respectively, allowed Morakot to decelerate from a forward speed of 6-7 m s\(^{-1}\) at about 24 hours before landfall to a minimum of 1.4 m s\(^{-1}\) at 18 hours succeeding landfall. This deceleration in forward motion was more rapid compared to the other TCs in the sample. Sustained moisture flow from Goni in an east-west orientation provided a conducive environment for sustained convective activity as high levels of moisture flux convergence were advected through the Taiwan Strait and over the Taiwan mainland (Chien and Kuo 2011). The stationarity of convective activity over Taiwan due to the slow propagation of Morakot, while not the most significant
contributor, allowed for topographic areas of southern Taiwan to experience rainfall totals greater than 3000 mm over a period of 4 days.

1.3.4. Subtropical Jet-Front Systems

Low-level jets help to transport large quantities of water vapor across long distances. As a forcing mechanism, low-level jets can act to destabilize the atmosphere enough to amplify convection which may lead to extreme accumulations. East Asia is a prime example of a highly precipitious region of the world. The mei-yu (China), baiu (Japan), or changma (Korea) front is a stationary warm-season baroclinic feature that extends through much of Taiwan and China and to the East China Sea just south of Japan. This feature, which derives its moisture from the South China Sea and the Bay of Bengal, is responsible for most of the precipitation in the warm season. Instability is generated under the right entrance region of a low-level subtropical jet stream as warm air advection processes induce vertical ascent through ageostrophic mechanisms (Sampe and Xie 2010; Fig., 1.3). Given that this feature sits primarily in the warm tropics, there is very little temperature gradient. However, the combination of the inherent slow motion of developing MCSs and high moisture content act to raise the potential for extreme-precipitating systems in the region. This environment and the weak steering mechanisms involved may lead to the persistent development of showers and thunderstorms (Chen and Chang 1980).

Another well-documented area of heavy precipitation events related to a subtropical jet feature is in the Caribbean sea, particularly during the onset of the late spring Caribbean rain belt (Allen and Mapes 2017). Moist tropical air is
advected northward via strong low-level southerly winds and then can be bent in a northwest-southeast orientation depending on the latitudinal shear of the subtropical jet situated aloft of the Caribbean. A strong upper-level trough over the Caribbean helps to induce ageostrophic uplift over a narrow but robust region of moist, conditionally unstable air (Fig. 1.4). Synoptic-scale weather systems can be partly responsible for the development of deep convection along the moist airmass, but the convective and mesoscale processes associated with certain features such as jet streaks and ageostrophic uplift dynamics can lead to the development of thunderstorms with potentially intense rainfall rates.
Figures

Figure 1.1: Geopotential height anomaly maps based on TRMM satellite data and National Center for Environmental Protection (NCEP) reanalysis data. The geopotential height anomalies for the July 28, 2010 case study are in (a). Topographic heights for the entire domain are shaded in (b) with a logarithmic scale to emphasize the relief between the foothills and the Tibetan Plateau and Himalayan mountain chain. Panels (c) and (d) display typical geopotential height anomalies associated with storms originating in the Western Himalayan Foothills and the Bay of Bengal, respectively. From Romatschke and Houze (2011) Fig. 1.
Figure 1.2: Schematic diagram of evolutionary trend in (a) training line/adjoining stratiform (TL/AS) and (b) backbuilding/quasi-stationary (BB) MCSs. Radar reflectivities of 20, 40, and 50 dBZ are given by the white, light grey, and dark grey contours and shading, respectively. For (a), the frontal boundary denotes the low-level and mid-level shear arrows which pertain to the surface-to-925 hPa and 925-500 hPa levels respectively. The dashed/dotted line in (b) details an outflow boundary that is typical of BB MCSs. The length scale at the bottom can vary by a wide margin and is mostly dependent on the number of active convective cells. From Schumacher and Johnson (2005) Fig. 3.
Figure 1.4: Schematic explaining the onset of meiyu-baiu over Eastern Asia during the early summer months. The smaller green arrows indicate the mid-tropospheric flow while the thicker green arrows represent the mean eastward jet stream motion and transient motion of mid-latitude systems (alternating high and low pressure dashed circles). Low-level southerly flow is given by the black arrows. The orange contour over China and Japan indicate the region of convective ascent with precipitating clouds. The typical positioning of the warm low and the North Pacific high are given by the red and blue shaded ovals, respectively. The solid black contour line denotes the 3000-m isohypse. From Sampe and Xie (2010) Fig. 15.
Figure 1.4: Schematic of factors leading to the development of the Caribbean rainfall belt. In (a) the pink region indicates the location of typical rainfall within the belt. The 250-hPa and 850-hPa isobars are given by the black and red contours, respectively. The pink clouds and the blue shading in (b) and (c) denote the typical location of convective development within a plume of moist air (blue). The red arrows show the typical movement of storms within the belt modulated by the 250-hPa trough over the Caribbean. From Allen and Mapes (2017) Fig. 6.
Chapter 2: Data and Methods

Previous studies have evaluated individual observational and reanalysis datasets for their overall performance in detecting precipitation (Bosilovich et al. 2008; Peña-Arancibia et al. 2013), but precipitation estimate accuracy continues to be a statistical and scientific challenge. This is especially prevalent in reanalysis products where model physics and biases contribute to the overall skewness and uncertainty of the data output (Adler et al. 2001). This chapter will focus on the suite of precipitation data products and how the outputs will be visualized.

2.1. Observational Datasets

2.1.1. TRMM 3B42

The principal rainfall observation product to be utilized here is the Tropical Rainfall Measuring Mission (TRMM) 3B42 precipitation product (Huffman et al. 2007), provided by the National Aeronautics and Space Administration (NASA) as a joint mission with the Japan Aerospace Exploration Agency (JAXA). The data is available at a native spatiotemporal resolution of 0.25°, 3-hourly time steps for all grid points covering latitudes 50°S to 50°N. TRMM’s primary coverage area is focused over the latitude belt 38°S to 38°N but extra satellite inputs complete the grid. Although TRMM overpasses are only twice a day, the combination of these swaths along with other space-borne data makes for excellent temporal coverage. To estimate total precipitation, algorithms within the TRMM 3B42 sensor take in data from a combination of high quality (HQ microwave data and estimates derived from infrared (IR) cloud-top brightness temperature data where available.
Analyses from surface station rain gauges are then used to perform final calibrations to the data. Data from version 7 of TRMM 3B42 is utilized due to its improved detection and intensity of rainfall events (Liu 2015). TRMM was effectively decommissioned after 2015, although the 3B42 product will still output data through a portion of 2018.

2.1.2. GLDAS

The Global Land Data Assimilation System (GLDAS; Rodell et al. 2004) product is a pseudo-real-time dataset that assimilates satellite-derived and ground-based precipitation data served up through NASA. Land-surface fluxes are resolved through thoroughly developed modelling and data retrieval techniques while still providing a high spatial resolution of 0.25°.

2.1.3. CPCU

The Climate Prediction Center “Unified” (CPCU)-Gauge Based Analysis of Global Daily Precipitation is a high resolution rainfall assimilation product from the National Oceanic and Atmospheric Administration (NOAA). As the name above suggests, this is a land-gauge-based product which assimilates data measurements from other stations, radar and satellite-borne observations, and numerical weather prediction outputs for quality control. Data is available from 1948 to the present day and is provided daily at a 0.5° spatial mesh.
2.1.4. CHIRPS

The Climate Hazards Group InfraRed Precipitation with Station (CHIRPS; Funk et al. 2015) dataset combines very fine resolution satellite imagery with ground-based station data (when and where available) from both NASA and NOAA onto a 0.05° grid. The spatial coverage of CHIRPS is identical to TRMM 3B42 and extends back to 1981.

2.1.5. GSMAP

The Global Satellite Mapping of Precipitation (GSMAP; Okamoto et al. 2005) is another satellite-derived precipitation product by JAXA that uses a blend of microwave radiometer and infrared imagery to deduce precipitation estimates similar to the methods used in TRMM 3B42. The data is collected, calculated, and released to the open web at a spatial resolution of 0.1°.

2.1.6. GPCP 1DD

The Global Precipitation Climatology Project 1-degree daily (GPCP1DD) version 1.2 rainfall product is a NASA satellite product. Data from GPCP1DD is calculated such that the rainfall estimates are obtained via the combination of geostationary IR satellite imagery, the GPCP monthly product, and Television IR Observations Satellite (TIROS) Operational Vertical Sounder (TOVS) to produce daily rainfall totals. Native outputs of GPCP1DD are of 1° daily spatial temporal resolution with data spanning back to 1997.
2.1.7. MSWEP

The Multi-Source Weighted Ensemble Precipitation (MSWEP; Beck et al. 2017a) dataset is one of the newer datasets as it was only released in 2016. It ingests data from observational, gauge-based and reanalysis product sources to assemble a precipitation estimate with full global coverage. The spatiotemporal resolution is a fine 0.1° and data extends back through 1979. Beck et al. (2017b) showed that MSWEP streamflow simulations were superior to other precipitation products in the United States, western Europe, and eastern Australia.

2.1.8. CMORPH

The Climate Prediction Center MORPH (CMORPH; Joyce et al. 2004) technique dataset ingests data from passive microwave sensors onboard several satellites including the Defense Meteorological Satellite Program (DMSP) 13, 14, 15, NOAA-15, 16, 17, 18 (making up the advanced microwave sounding unit – AMSU), NASA Aqua’s advanced microwave scattering radiometer (AMSR), and TRMM microwave imager (TMI). CMORPH combines the series of existing precipitation estimates instead of utilizing a different algorithm to estimate precipitation totals, thus allowing for other microwave satellite precipitation estimates to be included. CMORPH data outputs at a 3-hourly, 0.25° spatiotemporal resolution.
2.1.9. PERSIANN-CDR

The Precipitation Estimation from Remotely Sensed Information using Artificial Neural Networks- Climate Data Record (PERSIANN-CDR; Ashouri et al. 2015) utilizes approximations derived from neural networks to estimate precipitation totals based on infrared brightness temperatures. The PERSIANN dataset assimilates data from geostationary infrared and daytime visible satellite imagery, and an algorithm based on longwave infrared data calculates rainfall. PERSIANN-CDR data is available at 0.25° spatial resolution for multiple temporal resolutions (e.g., 1-hourly to yearly) for a global grid spanning 60°S to 60°N with data available starting in January 1983.

2.2. Reanalysis Products

2.2.1. MERRA and MERRA-2 Reanalysis

The chief reanalysis product that is utilized in this study is MERRA version 2 (MERRA-2; Gelaro et al. 2017) between NASA and the Japanese Aerospace Exploration Agency (JAXA). Data is available beginning in 1980 to the present day at 1-hourly, 3-hourly, or 6-hourly 0.67° X 0.5° grids and is downloadable through a highly-aggregated database known as a GrADS Data Server (GDS). Many meteorological enhancements to the assimilation systems from the original MERRA products are applied such that several over-sensitivities and biases in the land-surface estimates were addressed and corrected. Data from the original MERRA version is consulted for the rainfall outputs that will be shown in various cases in chapter 4.
2.2.2. ERA-Interim

The European Centre for Medium-range Weather Forecasting Re-Analysis Interim product (ERA-Interim; Dee et al. 2011) provides 6-hourly interval data at a T255 resolution which corresponds to a spatial grid of ~0.7° X 0.7°. Unlike its predecessor products like the ERA 40 year which could only assimilate in three dimensions, ERA-Interim’s assimilation technique has been advanced recently to include four-dimensional analysis of meteorological variables (4DVAR).

2.2.3. CFSR

The Climate Forecast System Reanalysis (CFSR; Saha et al. 2010) is a coupled ocean-atmosphere reanalysis product that includes an interactive sea ice model along with multiple satellite observations to create a best-guess depiction of the atmospheric state. Data from CFSR exists for all global grid points at 0.5°, 6-hourly resolution, except for the tropics and near the equator where the resolution is a finer 0.25°.

2.2.4. JRA-55

The Japanese 55-year Reanalysis (JRA-55; Kobayashi et al. 2015) is a reanalysis project from the Japanese Meteorological Agency (JMA). JRA-55 uses a second-generation four-dimensional data assimilation technique that improves on much of the biases and shortcomings from its predecessor product JRA-25. The output of JRA-55 is made up of a combination of observational data used within JRA-25 as well as the newly processed data from METEOSAT and the
Geostationary Meteorological Satellite (GMS; best known as “Himawari”). The dataset served out by NCAR and used here is of 1.25°, 6-hourly resolution, although its native counterpart is a much finer 0.5625°.

Altogether, a total of 14 datasets are consulted for investigating the spatial distribution of rainfall extremes.

### 2.3. Data Analysis Techniques

#### 2.3.1. Development of Extreme Precipitation Record Database

Since a goal of this thesis is to analyze the extremes on spatiotemporal scales that are not too short or too long, the methods in Mapes (2011) were employed to calculate record events at each grid cell. The TRMM 3B42 data was coarsened from the native 0.25° resolution to a global 1°, 2°, and 4° which yield grids of 1440x400, 360x100, 180x50, and 90x25, respectively. This was performed using second-order conservative remapping with climate data operators (CDO). For the finer-scale resolutions in which the size of the dataset is much too large to be stored into memory all at once, the data was broken up into several chunks to ease processing. Of course, smaller grids generally reveal the fine scale of some of the more extreme precipitation events and it is expected that peak precipitation values should decrease on a coarser grid. A coarse grid cell with a high precipitation maximum is surely worth investigating.

To define total accumulations, a 27-point (75-hour) running average was applied to all of the grids with CDO. The objective here is to determine the greatest 75-hour rainfall total for each grid point, and so the maxima for the entire 18-year
time series in each grid point was calculated with Python’s NumPy and xarray libraries. Values for the maximum accumulation as well as the year, month, day, and hour of occurrence for each grid point are stored in arrays.

For the data in the web atlas to be accessed properly as image files with single-byte depth, the data were encoded into bytes such that the arrays were rescaled depending on the highest value in each grid. For instance, precipitation totals were rescaled by a factor of 6/50 (as in Mapes 2011), and then 1/5, 1/4, and 1/2 for the 0.25°, 1°, 2°, and 4° grids respectively. The rescaling limits the dynamical revolution of each maximum value to 1/256th of itself. Years were rescaled by subtracting 1998 (the start of the TRMM 3B42 data life) and multiplied by a factor of 14. Months, days, and hours of record were rescaled by factors of 20, 8, and 12, respectively. These values are indeed arbitrary, but the key in this approach is to compress the data to fit to the byte spectrum of 0-255. PHP code adapted from Mapes (2011) fetched this byte-formatted data and upscaled back to the corresponding integer values in the original data arrays. A duplicate, colored GIF image is superimposed above the black-and-white PNG for user-friendly selection.

2.3.2. Case Study Analysis

To achieve an apples-to-apples comparison between the precipitation products, the datasets described in sections 2.1 and 2.2 have all been coarsened to a common 1° mesh. This re-gridding was performed using first-order conservative interpolation, and only days containing six or more 3-hourly samples
were used to estimate the daily average rain rate. Since precipitation estimates from TRMM 3B42 will be used as reference, the combination of various observational and reanalysis products will help to bracket any uncertainties within the comparison. Any similarities and differences between product outputs will be visible in a series of 16 panels (14 total datasets and an ensemble mean and root mean square) that will aid in this analysis. Based on the user-specified latitude-longitude-time parameters, a script fetches the server-side datasets and computes the accumulated rainfall totals for each dataset in the suite¹.

To interactively visualize the state of the atmosphere, this work makes use of Unidata’s Integrated Data Viewer (IDV). The IDV is a powerful, open-source data visualization software package that interactively allows for the study of both two-dimensional and three-dimensional variables in space and time. The IDV stores the case study information into “bundles” which contain hyperslabs of data and its associated time-based metadata.

Even with the raw power of IDV, studying multiple fields for each individual dataset can be time-consuming and unnecessary as the uncertainties within each product outweigh the lessons that can be learned beyond those that are already documented. For this work, two and three-dimensional variable data from MERRA-2 are utilized to construct bundles of the atmospheric state for each analyzed case. The structural evolution of cases are performed via analysis of static images and animations of various features of interest to determine any significance in

¹ A Python script for which these multiple outputs are computed and visualized can be obtained on https://github.com/bmatilla/JuPyIDV_Notebooks/tree/master/Template.
development of the extreme precipitation event. Variables that are investigated for each case are given in chapter 4.
Chapter 3: Statistics of Global Rainfall Extremes

Because the focus of this thesis is to examine the larger-scale extremes beyond a single-day event, the greatest 4-degree, 3-day rainfall totals across the entire TRMM 3B42 coverage area are shown in Fig. 3.1a with the 18-year climatology for direct comparison in Fig. 3.1b. These figures are presented in the form of a global “atlas” similar to the plots readily available on the web\(^2\) with the longitudinal grid shifted to center over 180°. By accumulating the greatest 3-day rainfall totals, each square pixel corresponds to the event at the top \(0.045^{\text{th}}\) percentile \(\left(\frac{3}{365.25 \times 18}\right) \times 100\%\). Therefore, it is important to note that each value is significant regardless of the magnitude of record. Nevertheless, the color scale is skewed towards higher values in order to better distinguish the higher-value accumulations.

From a global perspective, the densest concentration of the most extreme record rainfalls is located over tropical Asia given by the range of orange to white pixels. Because of the extensive amount of moisture available from warm seas and a multitude of mesoscale and synoptic-scale patterns and features that develop in and around the Indian Ocean (Zhou et al. 2017) and South China Sea (Xu et al. 2009), these areas can yield many different case studies alone. In Oceania, the greatest concentration of heavy rain events is focused over northern Australia and the islands of Polynesia. In particular, Australia’s warm-season rainfall patterns can have significant variability depending on the strength of the El-Niño Southern Oscillation (ENSO) teleconnections and landfalling TCs while cool-season events

\(^2\) http://weather.rsmas.miami.edu/HeavyRains_clickmaps
are dominated by synoptic-scale blocking patterns and subtropical ridging (King et al. 2014).

Shifting the view towards the North and South American continents, the number of significant total accumulation pixels are not as great as those over Asia. In mid-latitudes, there are a few areas of relatively elevated record accumulations of precipitation such as near the Cascade Mountain range in the Pacific Northwest, where most of the heavy rainfall may be attributed to the winter-time presence of onshore and upslope flow commonly referred to as “atmospheric rivers” (Warner et al. 2012). In subtropical areas such as the southeastern US, swath-shaped patches are indicative in some areas of the passage of moving TCs and weak baroclinic systems during the warm season (e.g., Atallah and Bosart 2003; Schumacher and Johnson 2006, Schumacher and Johnson 2008), but also in others the passage of moving midlatitude cyclones with strong integrated vapor transport during boreal winter and spring (Moore et al. 2015). In deep-tropical areas such as the western Caribbean Sea, heavy rains can be attributed to more frequent passages of strong TCs as well as small-scale convective bursts associated with the onset of the late spring early rainfall season (Allen and Mapes 2017). South American rainfall events are of more convective-scale nature and are not well-resolved at larger spatial scales. However, annual rainfall within the tropical Amazon contribute ~20% to the global total freshwater discharge budget (Richey et al. 1989) and record events are still significant to analyze as the presence of back-building MCSs (Schumacher and Johnson 2005; Rasmussen et al. 2014) make for productive rainfall events across the South American continent.
Records over the Rio de la Plata basin (including Brazil, Paraguay, Uruguay, eastern Bolivia, and Argentina) constitute only a small fraction of total rainfall events. Within this small fraction, “approximately 95% of the total summer rain in the southern La Plata basin region is accounted for by contributions of storms with extreme convective and stratiform elements” (Rasmussen et al. 2016).

Tropical Africa’s record rains have similar fine-scale spottiness, but also with somewhat larger mesoscale size patches (<100 km), especially in the Sahel where organized storms can occur in environments that can be too dry for ordinary convective cells to produce heavy rains (Mathon et al. 2001). Over Europe, records are much lower in magnitude compared to other regions of the world, but sporadic record rains are seen over the northwestern Iberian Peninsula suggested by the development of atmospheric rivers across the North Atlantic Ocean (Ramos et al. 2015). Records over the Mediterranean region are also sparse, but Mariani and Parisi (2014) indicate that the development of cut-off lows and MCSs as a few of the principal driving mechanisms.

To emphasize the significance of extreme totals in these 3-day record events, the ratio between extremes and climatology is displayed in Fig. 3.1c. This ratio is calculated by dividing each pixel’s maximum value from the climatology in Fig. 3.1b to yield the approximate amount of months of rainfall in any 3-day record event. In heavily precipitating areas such as southeast Asia and Oceania, the records only reflect totals that are on the order of 1-2 months’ worth of rain (up to 1/6 of the annual mean). Likewise, the near-equatorial belt of the Pacific Ocean has records that do not contribute beyond a single month’s worth of rain. However,
subtropical desert areas as in the Sahara Desert in Africa, Arabian Peninsula, and ocean-land areas bounded by eastern boundary currents (e.g. Pacific coast of the North and South American continents; Atlantic coast of west Africa) have record events that easily amount to many months in climatology, even where the magnitude of record rains is quite low.

The months in which these record 3-day precipitation accumulations are set are represented in Fig. 3.2. While it is a rather noisy statistic, seasonality of these extremes can still be interpreted. Most of the records are set during the austral and boreal summer months, especially in India, China, and Australia. Other locations such as the North Atlantic Ocean and Caribbean Sea have their record events observed during the TC-active Fall months. Still, it is possible that some record events diverge from climatological peaks.

The histograms that summarize these record values are represented in Figs. 3.3a-d for the 0.25° (576000 pixels), 1° (36000 pixels), 2° (9000 pixels), and 4° (2250 pixels) resolutions respectively. The 27 and 75-hour precipitation totals are collocated in each histogram to reflect the different running totals. Going from the finer to coarser resolutions in each panel, the amounts decrease only slightly. The single-day modal rainfall record amount is ~100 mm while the 3-day records are most frequently ~160 mm, values that hold across all scales which are consistent with the findings in Mapes (2011). While the precipitation axes’ values drop from 600 to 300 mm in decrements of 100 mm, these declines in coarser grids are due in part to how heavy rainfall events are more than just isolated convective events. Rather, as evidenced by the right-skewness of these histograms, one
would be able to discern that extreme precipitation events are in fact a rare occurrence. Higher 3-day records are indicative that mesoscale features play a role in both the distribution and intensity as some organized events such as TCs or MCSs can produce significant rainfall totals over a region. In wet regions (e.g., Southeast Asia), the adjacent days to the central date contribute <30% to the total rainfall in the event. On the contrary, drier areas (e.g., African Sahara; eastern Pacific subtropics) have the adjacent two days contribute a greater percentage of the total.

The distribution of the yearly count of precipitation maxima in each spatial grid is shown in Figs. 3.4a-d. Each pixel's year-of-maximum value is captured and stored in arrays after being converted from the standard “seconds from” notation. Across all grids, the peak year of maximum precipitation was in 1998, but this is largely due in part to the clumping of maximum precipitation values over the eastern Pacific “cold tongue” during the 1997-1998 El-Niño event even as the accumulation values were very low (Fig. 3.1a). 2016 had the fewest total cases across all grids. This is a surprising result considering the development of a robust negative signal in the ENSO southern oscillation index during 2015-2016. One possible explanation for the reduced case count involves the decommissioning of the TRMM satellite in April 2015. Since then, an algorithm based on the Global Precipitation Mission’s (GPM) integrated multi-satellite retrievals for GPM (IMERG) was used to compute the 3B42 estimates.
Figure 3.1: Color-pixel plots of precipitation amounts in the TRMM 3B42 product domain in 1998-2016 at the coarsened 4-degree resolution for (a) record 3-day accumulation (units mm), (b) climatological annual rainfall divided by a factor of 12 (units mm month$^{-1}$), and (c) the ratio of maxima to climatology taken from a/b.
Figure 3.2: As in Fig. 3.1, but for the month of record rainfall. The solid white pixels on the International Date Line indicate that no month data was present for those grid cells.
Figure 3.3: Histograms of the spatial structure of extreme TRMM 3B42 accumulations (units mm) over 27 h (blue lines) and 75 h (orange) sliding time windows for (a) 0.25-degree, (b) 1-degree, (c) 2-degree, and (d) 4-degree spatial resolutions.
Figure 3.4: Histogram of total pixel count for 3-day precipitation maxima in TRMM 3B42 from 1998 – 2016 for (a) native 0.25-degree, (b) 1-degree, (c) 2-degree, and (d) 4-degree spatial grid.
Chapter 4: Meteorological Analysis of Extreme Precipitation Events

The work so far in this thesis has identified the total maximum precipitation in any single-day (27-hour) and three-day (75-hour) period over the length of the 18-year precipitation dataset from TRMM 3B42 on various spatial grids. Understanding events in which these extremes occurred adds another nugget of knowledge regarding the dynamic and thermodynamic interplay leading to extreme precipitation events.

In this chapter, the moisture and dynamics at the forefront of heavy rain events are analyzed from a triplet of case studies subjectively picked to represent systems responsible for significant social impacts. More specifically, these case studies represent heavy rain events limited to a single geographical region for clarity of comparison instead of arbitrarily selecting the most extreme of extremes from vastly different regions.

4.1. Background and Case Selection

4.1.1. Key Variables Considered

Potential vorticity (PV) is a conserved quantity and is a useful diagnostic tool of rotation in the atmosphere, particularly in the larger synoptic-scale circulations. Raymond and Jiang (1990; hereafter RJ90) postulated that mesoscale convective vortices (MCVs) can be idealized as balanced mid-tropospheric PV anomalies sandwiched between two isentropic ($\theta$) surfaces (Figs. 4.1a and 4.1b). In the presence of vertically sheared background flow, air moves...
below the PV anomaly, lifting the isentropes downshear of the PV anomaly and
descending the isentropes upshear of the PV anomaly. Parcels in the lower levels
of the atmosphere would then undergo adiabatic ascent as they follow the
isentropic surfaces. This low-level adiabatic ascent can potentially act as a catalyst
for the growth and sustainability of secondary convection downshear of the PV
anomaly (RJ90; Trier et al. 2000). In essence, the ensuing latent heat release
contributes to vortex stretching. However, one of the caveats in RJ90 involves the
uncertainty behind the shorter-term, more transient motions on the narrower
mesoscale which could not be resolved in their model. The general framework in
which RJ90 define this sustenance of convective activity via θ-PV interactions is
applied to determine if these potential mid-level instabilities leading to the heavy
rains exist in these case studies.

The quasi-geostrophic (QG) framework can also provide abundant insight
into the dynamics involved with heavy rain events. Vertical behavior of air parcels
can be diagnosed through an understanding of the QG omega equation given by

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

(3)

where vertical velocity (left-hand side term) is \( \omega \). The first term on the right-hand
side is the vertical differential of vorticity advection and the second term describes
the horizontal Laplacian of thermal advection, respectively. Recalling that the
Laplacian is proportional to the negative of its associated term (for simple
sinusoidal patterns) near its extreme values, it can be inferred that vertical ascent
(descent) will occur when the sum of the vorticity and thickness advection terms is
positive (negative). While specific quantitative estimates of \( \omega \) are limited in (3)
because ageostrophic observations are not considered (i.e., in the frame of a geostrophic atmosphere), interpretations can still be made to better understand the sign of $\omega$. Primarily, vertical ascent of air parcels is coupled to more rainfall and the horizontal surface convergence is a forcing for low-level cyclone intensification (Grimes and Mercer 2015).

To the first order, the atmosphere heats and cools by radiative, latent, and sensible heat fluxes. Diabatic heating is another way to diagnose the state of the convecting atmosphere, primarily as it is a proxy to moisture convergence via latent heating processes. The total diabatic heating tendency product due to model physics in MERRA can be expressed in component form such that:

$$\frac{dT}{dt}_{\text{tot}} = \frac{dT}{dt}_{\text{fri}} + \frac{dT}{dt}_{\text{gwd}} + \frac{dT}{dt}_{\text{rad}} + \frac{dT}{dt}_{\text{mst}} + \frac{dT}{dt}_{\text{trb}} \tag{4}$$

In (4), the total diabatic heating is dependent on the temperature tendencies due to the frictional dissipation of kinetic energy through processes such as surface friction (fri), dissipation of kinetic energy via gravity wave drag (gwd), radiative processes (rad), moist processes including latent heating via condensation and evaporation within the convective parameterization scheme (mst), and turbulence due to sensible heat fluxes (trb). In regions that are convectively active such as the tropics, MCSs exhibit the greatest radiative and latent heating in the midtroposphere with a net cooling concentrated in the lower troposphere, while stratiform regions have a net heating just above the melting level (Houze 1982). This knowledge of the vertical structure in diabatic heating and cooling can provide an insight to the development of positive PV anomalies as alluded to in RJ90, which can then cascade into the aforementioned growth of secondary convection.
leading to heavy rain events. This makes vertical diabatic heating profiles another key variable to analyze in the case studies, especially in events involving organized circulations.

4.1.2. Domain of Study and Case Selection

The Indian Summer Monsoon (ISM) is responsible for a large portion of the total annual rainfall over the Indian subcontinent. This feeds directly into the agricultural yield and back into the economy of India. The unique topography of the Indian subcontinent makes the region susceptible to frequent flooding events. Fig. 4.2 shows that while most of India is relatively flat, the country has rolling hills and relatively high mountain peaks just inland of both coastlines. Furthermore, high terrain from the Tibetan Plateau exists to the north of India and has been previously shown to cause implications in ISM transitions and ISM rainfall patterns (Sato and Kimura 2007). Extreme precipitation events compound the risks in places that are already prone to flooding from seasonal rains, especially in areas where distinct topographic features could play a role in the genesis and regrowth of convection leading to significant precipitation.

Three case studies over the Indian subcontinent will be described in this chapter and the geographical locations for each case are presented in Fig. 4.3. The first event is an organized low-pressure system over the Bay of Bengal that was eventually classified as Deep Depression BOB 02 (hereafter BOB02) by the Indian Meteorological Department (IMD) (RSMC 2007). The second event is a mid-summer quasi-stationary rain event over the state of Madhya Pradesh in central India, and the third event is another summer heavy rain event in the eastern
Maharashtra relatively close to Nagpur associated with another Bay of Bengal deep depression. Recalling the most critical ingredients to the development of convection (e.g., moisture, instability, and lift), a concise meteorological analysis for each case is described, capturing the essence of the moisture and dynamics involved with each case. A quick overview of the metadata regarding the three investigated cases are provided in Table 4.2.

Monsoon depressions that typically form over the Bay of Bengal and propagate towards the India or Bangladesh coastline are characterized by a low-level cool core that extends through much of the lower levels with high levels of near-surface convergence (Fig. 4.4, from Hunt et al. 2016). Conversely, a warm-core is typically located in the middle and upper levels, and the region of highest PV typically resides between 650 and 350 hPa where peak vertical ascent rates are also expected. The strongest winds are located on either side of the center of circulation (Fig. 4.4). The upper-troposphere is characterized by the presence of divergent outflow and another moist cloud cover deck similar to the one located in the lower-troposphere but only broader.

The hypothesis presented in this chapter is based on the growth and sustenance of a mid-level PV vortex associated with monsoon low pressure systems and its effect on secondary convective initiation (e.g., Raymond and Jiang 1990; Trier et al. 2000) downshear. In essence, if a positive PV anomaly exists in the mid-levels (similar in structure to that presented in Fig. 4.4) and is in combination with a highly moist and unstable atmosphere, low-level adiabatic ascent leading to secondary convection may be favored downshear of the PV
anomaly and thus leading to significantly high rainfall totals over a short period of
time.

4.2. Case 1: Bay of Bengal Deep Depression Two (2006)

BOB02, the third classified tropical system of the 2006 North Indian Ocean
cyclone season, began as a disorganized area of convective activity over the
southern Bay of Bengal. Peak winds for BOB02 were estimated at 18 m s\(^{-1}\) by July
2 at 1200 UTC based on the Joint Typhoon Warning Center (JTWC) best track
analysis. Landfall was only a few hours beyond peak intensity as the center of the
system crossed the Odisha coastline between the cities of Paradip and Chandbali
(RSMC 2007). BOB02 claimed 131 lives via landslides, collapsing infrastructure,
and floods from the heavy rains; 77 of those deaths were from the Odisha and
Vidarbha states (India Meteorological Department 2007). On the 4-degree scale
from the web atlas (box 1 from Fig. 4.3), the TRMM 3B42 precipitation estimate
exceeded 358 mm with a central date of June 30, 2006 at 0900 UTC.

Fig. 4.5 presents the reanalysis and observational rainfall accumulations
described in sections 2.1 and 2.2 for intercomparison. Since most of the
accumulations in an event are antecedent and subsequent of the central date in
the TRMM 3-day record window, the accumulations for 5 days is shown. This
allows for the full understanding of the growth of the event and to safely bracket
any precipitation uncertainties in time. The 5-day TRMM 3B42 precipitation
accumulation was estimated to be >800 mm, which aligns well with the 3-day atlas
estimate when considering much finer spatial scales and a slightly longer time
period. Most of the heavy precipitation totals occurred just offshore of Odisha state. GSMAP and CMORPH (Figs. 4.5j and 4.5m) also estimated high precipitation totals of ~700 mm and ~650 mm, respectively. MSWEP (Fig. 4.5f) resolves high precipitation totals of ~650 mm and a peak location similar to that of TRMM 3B42. Meanwhile, GLDAS, CPCU, and CHIRPS (Figs. 4.5a-c respectively) only contain ground-based data and thus completely miss any peak in the proper location. PERSIANN and GPCP (Figs. 4.5e and 4.5k) also miss the peak location and intensity with most of the higher values shifted to the south. In the reanalysis datasets, MERRA-2 (Fig. 4.5h) resolves the precipitation intensity and location fairly decently, albeit shifted just slightly to the north unlike MERRA (Fig. 4.5g) which blurs out any defined location. Meanwhile, JRA-55 (Fig. 4.5i) precipitation is less clearly resolved yet still features a slightly defined area of relatively heavier rains, and ERA-Interim (Fig. 4.5l) depicts much lower totals across a broader area. The non-land-confined mean estimate (Fig. 4.5o) peaks at ~350 mm while the ensemble root mean square (RMS; Fig. 4.5p) is ~500 mm. This sizable difference is implied by the datasets with very high rainfall totals such as TRMM and CMORPH, even with a majority of the remaining datasets blurring out high rainfall values as an offset.

The domain of interest for this case study is displayed as a white box in Fig. 4.6a with the total TRMM 3B42 precipitation and MERRA-2 (Fig. 4.6b) enclosed in it. Estimates within the native-resolution TRMM accumulations peak above 1,100 mm towards the northeast quadrant of the “raining core” (cluster of pink pixels) over the 5-day period, but are otherwise above 600 mm in the vicinity where
BOB02 traversed. On the other hand, MERRA-2 resolves the highest precipitation just to the north of TRMM (on the order of 1-2 grid points), but peak rain totals are comparable at ~900 mm, consistent with the observations in Fig. 4.5.

GridSat IR satellite imagery and TRMM 3B42 rain rates from June 28 to July 2, 2006 illustrate the gradual development of BOB02 in the Bay of Bengal (Fig. 4.7). Deep convection associated with a mesoscale cloud cluster was located just east of India. As convection became more organized and consolidated on June 30 and July 1, intense TRMM 3B42 rainrates in excess of 15 mm hr$^{-1}$ were contained within the deep convection of the system (Fig. 4.7c-d). While most of the intense rain was offshore by July 1, thunderstorms along the western periphery of the storm crossed into Odisha. By July 2, peak rainrates of 20 mm hr$^{-1}$ were observed over the mainland with some of the convection. This coincides with the spike in rain totals over the immediate Odisha coast (Fig. 4.7a) beginning on July 1 09:00 UTC through 21:00 UTC in the animations (not shown).

The synoptic-scale setup during the development of BOB02 is displayed in Fig. 4.8. MERRA-2 rain rates are now displayed to provide clarity along with IR imagery on precipitating locations relative to the location of the monsoon depression. Nearing landfall, MERRA-2’s estimated MSLP in the defined center of circulation was ~986 hPa which is close to the 982 hPa record from JTWC best track analysis, along with the heaviest convection along the storm’s westward periphery (Fig. 4.8e). Also, as the storm develops over time, further organization leads to a westward mid-tropospheric axial tilt evidenced by the westward displacement in the 500-hPa heights relative to the surface. The presence of most
of the heaviest rainfall to the southwest of the cyclone center and the southwestward tilt aloft are typical characteristics inherent of the nature typically found of monsoon depressions (Fig. 4.4; Hunt et al. 2016; Boos et al. 2017).

Snapshots of daily-averaged MERRA-2 CWV values are shown in Fig. 4.9. Moisture along the Odisha coastline was estimated at ~70 mm with even higher values surrounding the center of the storm (> 75 mm). High CWV content is also present well ahead of the depression, with CWV values > 60 mm over inland areas of Odisha. Prior to landfall on July 2\textsuperscript{nd}, CWV values as high as 85 mm were observed as the system approached the northeast Indian coastline.

Fig. 4.10 shows that, while developing, the system was under the influence of modest westerlies of ~10-20 m s\textsuperscript{1} between the near-surface up to 500 hPa, then by easterlies between 300 hPa and 250 hPa downshear of the system (Fig. 4.10). However, the steering flow in the middle levels was generally weak with values < 10 m s\textsuperscript{1} around the vortex. As time progressed, the system encountered south-southeasterly flow in the upper-levels with diffluent (confluent) flow to the south (north). In the lower levels, the strongest winds are located on either side of the center of circulation, consistent with the findings in Hunt et al. (2016).

High atmospheric CWV content is a key component to the makings of an extreme rain event. Understanding the distribution of this CWV throughout a column can help to identify where this CWV is concentrated. To illustrate this, Fig. 4.11 shows the vertical profile of static energies as a function of mass. The usefulness in using this vertical profile structure is visualizing the troposphere in linear-pressure form as opposed to the traditional logarithmic-pressure found in
skew-t diagrams\textsuperscript{3}. Because of the lack of data provided by the Bhuaewar station, only a brief interpretation of the lower and middle troposphere can be facilitated. A very moist profile is observed where an estimated 56.8 mm of CWV is contained through the top of the sounding. Also, parcel buoyancy is enhanced beginning at a point just above the boundary layer where the parcel path is nudged towards $h_{\text{sat}}$ (blue curve). This suggests that while surface-based CAPE is low, available instability well above the boundary layer may be much higher.

Continuing with the vertical assessment of this case, a vertical cross-section of time-averaged PV and theta contours is given in Fig. 4.12. Data for PV has been capped at 200 hPa. PV is greatest in the mid-levels of the atmosphere, especially during the later stages of the storm’s development where an average of 3 PVU is observed. The time-averaged isentropic structure does not reveal a blatant bulge in the theta surfaces as the horizontal temperature gradient was generally weak. However, three-hourly snapshots show that a positive PV tower forms on July 30, 09:00 UTC which exists for several hours (Figs. 4.13a-c). A sharper isentropic gradient was observed closer to the PV maximum indicating an increase in static stability in the lower levels. But the isentropes downstream of this PV maximum bulge upward, indicating lower static stability. The maximized mid-level region of positive PV and raised isentropes with the gradual northwestward propagation does in a way support the RJ90 theory as thermal wind balance holds.

A cross-section of omega (vertical ascent/descent) is shown in Fig. 4.14. Near the surface, weak ascent is observed while vertical velocity values were the

\textsuperscript{3} Information regarding the technical specifications of the calculations to create the plot are found in Mapes (2017).
greatest at around 4 km above MSL, peaking at \( \sim 1.0 \) Pa s\(^{-1}\) (estimated as \( \sim 0.2 \) m s\(^{-1}\)). Vertical ascent was also found to exist in the upper troposphere which, coupled with the modest buoyancy aloft (Fig. 4.11), indicates the presence of convective instability.

Fig. 4.15a reintroduces the vertical PV profile used in Fig. 4.12 but now the time-averaged vertical profile of diabatic heating and cooling rates is juxtaposed. Overall, the diabatic heating profile has the greatest values rooted with the region of maximized PV between 500-300 hPa. A shift in the location of maximized diabatic heating was observed on June 30\(^{th}\) between the hours of 00:00 and 12:00 UTC when peak heating ascended from 5 km to 8 km while becoming more top-heavy in shape (not shown).

Similarly, the time-averaged diabatic heating is juxtaposed onto \( \omega \) in Fig. 4.15b. This mid-level diabatic heating gradient coincides with the peak vertical ascent that is observed. It is in this region of peak PV, modest observed ascent, and net positive diabatic heating that the greatest rainrates are observed. As BOB02 approaches the east Indian coastline near Bhubaneswar, observed slow storm motion due to a lack of steering flow and over the highly moist environment conspired to generate convection.

4.3. Case 2: Madhya Pradesh

The flood event of July 2-6, 2005 in Madhya Pradesh was the precursor to a more prolonged event that extended through the middle of July. In all, 62 people were killed and over 1 million people were greatly impacted due to the ensuing
floods. Telecommunications and infrastructure were also impacted severely, including two districts within Madhya Pradesh that were completely disconnected according to relief reports.

TRMM 3B42 (Fig. 4.16d) once again shows the highest total precipitation estimates at ~450 mm compared to the 3-day web atlas estimate of 226 mm. There is generally good consensus of the location and intensity between the other observational products. GSMAP and CPCU (Figs. 4.16j and b, respectively) were far superior in revealing the highest accumulated rain totals (~400 mm) while products like CMORPH and MSWEP (Figs. 4.16m and f, respectively) were more accurate in the spatial confines of heavy rain, given by the sharp gradient from high to low accumulations as in TRMM. Meanwhile, GLDAS blurs out a peak and shows a relative maximum further to the west.

MERRA (Fig. 4.16g) once again resolves rainfall to a lesser degree with maximum values of ~150 mm. MERRA-2 (Fig. 4.16h) is much better in resolving total accumulations with a peak value of ~350 mm that is further north than TRMM 3B42. MERRA-2 does show a sharp precipitation gradient, much like TRMM. JRA-55, while resolving a peak rainfall total of ~300 mm, displaces the most intense rain well to the northwest. ERA-Interim (Fig. 4.16l) shows its greatest rainfall totals to the east. The ensemble mean and RMS (Figs. 4.16o and p) for this case show peak values of ~235 mm and ~300 mm, respectively, suggestive of how these different datasets did not have much spread in overall intensity.

The compact coverage area and location of heavy rainfall is elucidated further in Fig. 4.17. Clearly there is an offset to both maximum values and areal
coverage of precipitation between the two products, as TRMM 3B42 sports higher totals over a greater area compared to MERRA-2 in which significant rainfall is only 2-3 pixels wide on the grid. TRMM 3B42 peak total estimates are ~550 mm while MERRA-2 is ~160 mm when juxtaposed over the same area, even as the absolute maximum is closer to ~400 mm just to the northwest. Lower model-resolved precipitation totals do not always indicate a lack of dynamics, but the discrepancy compared to observations still warrant an understanding.

IR satellite imagery from June 30- July 4, 2005 shows the presence of a weak cyclone-like feature being advected westward across the region (Figs. 4.18a-d). This cyclonic feature moves rather slowly and appears to be disorganized and convectively shallow. The overall cloud pattern appears more stratiform than convective, but some areas of deep convection developed over the same areas in Madhya Pradesh. Most of the rainfall remains southwest of the center while the northern periphery remains relatively rain-free, again consistent with the common rainy location on the southwest side of monsoon low pressure systems described in Hunt et al. (2016). Intense rainfall (>15 mm hr⁻¹) was found in the deepest convection as well as in the secondary convection to its west.

A center of circulation at the surface is resolved well by MERRA-2 with a minimum central pressure fluctuating between 990 and 994 hPa (Figs. 4.19a-e), and the low is closed and is vertically stacked through 500 hPa. Compared to the satellite imagery in Fig. 4.18, the storm appears to be sheared to the west as the center of the storm remains to the east of the dense cloud cover over central India. Upper-level troughing associated with the summertime subtropical jet is found over
the Tibetan Plateau which acts to cap significant meridional motion seen at 300 and 250 hPa (not shown).

A highly moist environment over most of the Indian subcontinent was observed over the length of the episode with CWV values exceeding 70 mm in northern India (Fig. 4.20), although this is confined to the immediate region around the system. Still, this amount of CWV highly encourages moist ascent especially over the rolling terrains of Madhya Pradesh. Relatively drier air is found to surround this narrow band of 70+ mm CWV particularly to the north and northwest where the sharpest CWV gradient exists.

Fig. 4.21 shows the five-day steering flows, all taken at 09:00 UTC consistent with the times in Fig. 4.19. An area of confluent flow was situated to the north of the low pressure system, particularly evident beginning on June 30 above 300 hPa and at 500 hPa on July 1. Interestingly enough, the system sits just to the southeast of an upper-level deformation zone which separates the very moist easterly flow from the Bay of Bengal from the much drier air flowing southeastward from Pakistan and the very dry air from the Tibetan Plateau to the north (Fig. 4.20). Deformation zones are more closely associated with frontogenesis in mid-latitudes, but there is a void in knowledge with regards to the development of these features influencing subtropical systems. Based on previous mid-latitudinal studies, deformation zones can be associated with steady precipitation that may or may not lead to considerable rainfall totals (Steigerwaldt 1986), especially if the deformation zones dominate the synoptic-scale flow patterns (Gao et al. 2008). Steering flows of $< 10 \text{ ms}^{-1}$ were observed above 500 hPa through July 4 while
lower levels were dominated by confluent westerlies particularly to the south and west of the storm center.

Given the lack of sounding stations in India and how spatially removed they are from one another, it can prove challenging to find a location that can more precisely resolve the true behavior of the environment during the event. Though the station in Bhopal is well-removed to the northwest from the actual rain event, the data from there may be useful to understanding the thermodynamic profile of the surrounding environment. Bhopal was under the influence of a highly moist environment with a total CWV content of 67.7 mm (Fig. 4.22), much of which was concentrated in the lower and middle troposphere. While there was no appreciable CAPE at the time of the sounding, the pre-conditioning of ample moisture provided a scenario for convection. CAPE also varies greatly with atmospheric heating during the course of the day, and an early morning sounding may not be indicative of the overall thermodynamic environment.

The vertical cross-section of PV for June 30 – July 4 shows a positive maximum in the mid-levels with a slight westward tilt aloft (Fig. 4.23). Isentropic surfaces bend towards the region of PV maximum but once again, the time-averaged PV and isentropic surfaces do not show a discrete bulge in the theta surfaces. It can only be best visualized in the instantaneous 3-hourly PV-theta analysis. As in case 1, the signature is a bit more robust on the sub-daily scale. The region of highest PV rose from ~3 km to ~8 km and tilted with time. High static stability is rooted perpendicular to the PV maximum on the east and west flanks while lower static stability was found much closer to the surface and ahead of the PV maximum (Fig. 
This result is more in line with the idealized PV-theta structure from RJ90 as the sloping of the isentropic surfaces around the PV maximum is far more evident. Moreover, the associated cool core of this system was advected over the highly moist environment, which likely induced vertical ascent critical to the recurring rainfall.

Figure 4.25 shows a cross-section of omega. This case, unlike BOB02 from section 4.3, shows that the peak vertical ascent was contained to the lowest 4 km, although weak ascent still does occur through much of the troposphere. Satellite imagery from Fig. 4.18 showed cloud types that resemble shallow convection as opposed to some of the deeper, more towering convection seen in the first case with BOB02.

Diabatic heating profiles in Fig. 4.26 confirm that most of the net diabatic heating takes place within the lower half of the atmosphere. There were two discrete peaks in the vertical diabatic heating gradient, one beginning on July 1 at 21:00 UTC and another one on July 3 at 12:00 UTC. Each burst of positive diabatic heating lasted for roughly 6-12 hours but showed the greatest precipitation rate and subsequent increase in total rainfall (not shown). In conjunction with the PV field, a connection can be established between PV and diabatic heating such that the mid-tropospheric axial tilt as mentioned previously can possibly steer the low pressure system. Boos et al. (2017) allude to the possibility that the IMD can be advected adiabatically via the diabatic heating processes. Furthermore, the sharp vertical diabatic heating gradient is potentially due to the model's (MERRA-2)
inability to adequately resolve grid-scale condensation and latent heat processes that are largely responsible for the generation of convection.

Based on the synoptic setup that was present during the life of the storm, it suggests that this is an example of a system that was under little influence of any significant steering flow. Stationarity is plausible as convection regenerated over the same area for several days under the presence of an upper-level deformation zone located to its north and weak steering flows in the surrounding environment. This case is consistent with the findings in Boos et al. (2017) in that this is a system in which the smaller-scale processes within the convective environment dominate the overall flow pattern.


During the first few days of August 2006, a broad low pressure system formed over the Bay of Bengal and slowly drifted westward. According to the JTWC, high vertical wind shear was present over the northern Bay of Bengal and northeastern India and Bangladesh. After gradually organizing itself, this system was officially classified by the IMD as Bay of Bengal Deep Depression 03 (BOB03 hereafter) on August 2. However, the maximum sustained winds only peaked at 15.4 m s\(^{-1}\) and observed proximity to the coastline hampered future intensification. The system was determined to have moved inland through the southern Orissa coast between Puri and Gopalpur by August 3 (RSMC 2007).

TRMM 3B42 (Fig. 4.27d) showed that the greatest rainfall totals were located over an area near 20N, 76E, with some relatively high totals spread over a greater area to the southeast of the maximum. The other ground-based datasets
offered mixed estimates; CHIRPS and GLDAS (Figs. 4.27a-b) blurred out the peak in the correct location while CPCU showed a maximum that was centered more to the east. PERSIANN showed precipitation over a broader area but missed the peak to the southwest. MSWEP and GSMAP both did well in resolving high accumulations of precipitation in a reasonable area closer to the TRMM estimates. GPCP, while decently resolving high precipitation, missed the peak to the southeast. CMORPH also showed higher precipitation values but peaked well to the southeast.

MERRA-2 is seen to be the reanalysis product that closely resembled the location and intensity of rainfall, with a peak of ~400 mm. In fact, MERRA-2 resolved more precipitation than what was observed. In contrast, MERRA showed high rain totals to the south of the TRMM peak but were not as impressive closest to the correct location. CFSR also showed respectable rain totals of ~200 mm, but resolved less rainfall than in the areas where TRMM 3B42 observed the highest values. JRA-55 showed very high precipitation totals of ~400 mm and a location similar to MERRA-2. ERA-Interim disappointingly missed any maximum and was unable to resolve a distinct area of high precipitation totals.

Most importantly, inspecting Fig. 4.27 for the ensemble mean (panel o) and RMS (panel p) reveals that a data challenge is present with this case study not so much because of total precipitation estimates but rather for location. With most of the rainfall estimates skewed to the south and east of the TRMM reference, it is inferred that this serves as an example of a poorly-analyzed case across multiple products. Poorly analyzed cases are not uncommon as each product has varying
calibration techniques and approaches. Nevertheless, data challenges can arise as a result. The investigation of why various product outputs diverge from one another is beyond the scope of this particular work, but it does help to raise questions for future studies that will be highlighted in chapter 5.

A closer inspection reveals that MERRA-2 did well in representing the overall coverage of precipitation across much of central India compared to TRMM 3B42 (Fig. 4.28). Even then (as alluded to in Fig. 4.27), MERRA-2 estimated more precipitation in some areas compared to TRMM especially in the north and east sections inside the indicated white bounding box.

Satellite observations show how BOB03 advanced slowly westward across the Bay of Bengal and onto the Indian subcontinent (Fig. 4.29). However, bursts of deep convection developed and propagated west-northwestward across Maharashtra and Madhya Pradesh. Intense rainrates (>15 mm hr\(^{-1}\)) were observed underneath of some of the deepest convection to the northwest, contributing to the secondary peak of total precipitation closest to the northwest edge of the precipitation coverage area in Fig. 4.28. These rainfall bursts also occurred as convection propagated downslope from the elevated terrain. Rainfall is mainly concentrated over the southwestern quadrant while areas to the north of the center of circulation remained relatively rain free.

BOB03 was vertically tilted towards the southwest for the 5-day period examined (Fig. 4.30). By August 4\(^{th}\) at 00:00 UTC, the center of the system was analyzed by MERRA-2 as being mostly inland with a reanalysis-estimated central
pressure of ~988 hPa. Slow weakening was observed as the system proceeded further inland and by August 5th, the central pressure rose by 8 hPa.

Daily-averaged CWV values were rather high at around ~55-60 mm (Fig. 4.31) in the vicinity of the cyclone. This is lower than the previous two cases described in this chapter, but this elevated CWV does present the potential for heavily precipitating convection especially if enhanced by inland topography.

Fig. 4.32 shows the snapshots of environmental flow. Once BOB03 moved inland, a broad region of > 20 ms\(^{-1}\) westerly winds was located to the southwest of the center of circulation while not as strong just to the north (Fig. 4.32) between 850 and 500 hPa. Easterly flow was observed above 500 hPa which suggests that the inflection in background flow took place at the middle levels, following RJ90. However, an interesting feature to note is in the confluent flow at 850 hPa to the south of the center. The confluent westerlies south of the densest cloud cover and deepest convection suggest that an airmass boundary existed to the south. This airmass boundary was observed to separate the much drier air over southern India from the high moisture airmass encircling the cyclone (not shown).

The sounding taken at Nagpur was located along the rain-free periphery of BOB03 and while its location was not ideal, it still revealed total CWV of 42.7 mm (Fig. 4.33). A relatively drier layer existed near the surface before near-full saturation was noted between 750-550 hPa. CAPE at the time was recorded at 2254 J Kg\(^{-1}\) rooted just above a small boundary layer inversion.

The highest PV is shown to once again be most concentrated in the mid-levels (Fig. 4.34), especially early in the evolution of the case. The time-averaged
field yields a PV structure that is quasi-uniformly positive across the cross-sectional area. Likewise, isentropic surfaces do not slope around the region of PV maximum. Examining the instantaneous PV-theta cross-sections yields some isentropic ascent ahead of a region of enhanced PV in the lower levels (Fig. 4.35).

Modest ascent exists throughout much of the cross-sectional area where vertical ascent rates are between 0.25 and 0.5 Pa s\(^{-1}\) (Fig. 4.36). However, weak subsidence was noted near the surface which suggests that some stability may have been present in the environment around the system.

Most of the latent heating takes place above the boundary layer, specifically between 4-10 km (Fig. 4.37). A net positive diabatic heating gradient is observed in the middle levels above 4 km with peak time-averaged diabatic heating rates above 1 K hr\(^{-1}\). Near the surface, a net diabatic cooling is observed. This is most likely due to evaporative cooling of rainfall.

For this case, any synoptic-scale features that could have influenced the system were not well defined. However, there were several other factors on both the convective-scale and mesoscale that contributed. A moist environment where CWV was greater than 55 mm was present over Orissa and Maharashtra coupled with the advected cool-core system over this moist region may have helped to grow convection in the area leading to the large accumulations.
Figure 4.1: Diagram of balanced lifting of a positive potential vorticity anomaly in an east-west-oriented shear environment. Vertical motions in (a) are related to the sheared background flow along the PV isentropic surfaces. Vertical motions in (b) are related to the PV anomaly flow along the isentropic surfaces in the south-north direction (dashed lines). From Trier et al. (2000) and adapted from Raymond and Jiang 1990.
Figure 4.2: Topographic map of the Indian subcontinent. State provinces are outlined in the narrow grey lines and country borders are outlined in the black lines.
Figure 4.3: Map of selected cases from the 4-degree precipitation maxima grid as in Fig. 3.1. The map is zoomed in over the Indian subcontinent with the case studies denoted numerically and by white boxes around each respective pixel.
Figure 4.4: Schematic of the overall structure of the composite Indian monsoon depression. The abbreviated terms are as follows: CC – cloud cover; CLWC – cloud liquid water content; PV – potential vorticity. From Hunt et al. (2016).
Figure 4.5: Five-day (June 28 to July 2 2006) daily rainfall accumulation estimates for all observational and reanalysis datasets on a coarsened 1-degree grid for the 4-degree 75-hour record event centered at 19 N, 87 E from the web atlas. The ensemble mean and ensemble RMS of all non-land-confined datasets are in panels o and m, respectively. Ensemble RMS utilizes a blue color scheme and color scale in order to differentiate from the rest of the datasets. Red contours are spaced at every 50 mm.
Figure 4.6: Total rainfall accumulation (units: mm) from June 28- July 2, 2006 for native resolution (a) TRMM 3B42 and (b) MERRA-2 products during BOB02. The white box corresponds to the region of interest.
Figure 4.7: GridSat IR satellite imagery and TRMM 3B42 rain rate (green pixels under clouds) at 09:00 UTC for June 28-July 2, 2006. Each panel represents 24-hour intervals from the start date of June 28.
Figure 4.8: MERRA-2 mean sea level pressure (red contours; interval 2 hPa), 500-hPa heights (cyan contours; interval 20 m), GridSat IR satellite imagery, and MERRA-2 rain rate (green pixels) at 09:00 UTC for June 28- July 2, 2006.
Figure 4.9: MERRA-2 daily-averaged column-integrated water vapor (colored shading; units mm) and contours (interval 5 mm) from June 28- July 2, 2006.
Figure 4.10: MERRA-2 wind speeds (yellow contours; interval: 10 m s$^{-1}$), wind streamlines (cyan contours) and GridSat imagery in sequential order from June 28- July 2, 2006 at 09:00 UTC.
Figure 4.11: Static energy-mass profile for Bhubaneswar, India (station code 42971; VEBS) for July 1, 2006 at 00:00 UTC with the station location indicated by the arrowhead in the inset (top right). Curve descriptions are as follows: Dry static energy (leftmost solid red line), virtual dry static energy (broken red line), moist static energy (solid blue line), and saturation moist static energy (rightmost solid red line). The total estimated column water vapor is given in the solid blue fill area. Isohumes are contained within the column water vapor fill and are denoted by the dotted black lines. CAPE is displayed as the solid green fill, which is proportional to the area integral of the parcel curve closer to saturation. Diagonal notches at 100 hPa intervals on the dry static, moist static, and saturation moist static energies represent the hypothetical adiabatic displacements on the order of ±100 m. Temperatures at the bottom represent the dry static and saturation static energy relative to the molecular boundary layer over surface water. On the left, equal and opposite-sign changes due to radiational interplay with static energies are given by the cyan column (approximated by the 1.3 K day\(^{-1}\), 10\(^7\) J m\(^{-2}\) cooling via radiation) and changes due to surface fluxes are in the orange square (equivalent to 116 W m\(^{-2}\) day\(^{-1}\), 10\(^7\) J m\(^{-2}\) warming), staging radiative convective equilibrium.
Figure 4.12: Cross section of time-averaged Ertel potential vorticity (units: K kg m$^{-2}$ s$^{-1}$) and potential temperature (units: K; contour interval: 2 K) from 0 to 12 km (1000 to 200 hPa) from June 28 – July 2, 2006. The cross-sectional area is depicted by the red in the inset with TRMM 3B42 time-averaged rainrates shown in green pixels.
Figure 4.13: Cross section of 3-hourly instantaneous Ertel potential vorticity (units: K kg m$^{-2}$ s$^{-1}$) and potential temperature (units: K; contour interval: 2 K) from 0 to 12 km (1000 to 200 hPa) for June 30, 2006 at (a) 09:00 UTC, (b) 12:00 UTC, and (c) 15:00 UTC. Cross-sectional location as in inset in Fig. 4.12.
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Figure 4.15: Cross section of time-averaged (a) Ertel potential vorticity and (b) vertical velocity from 0 to 12 km (1000 to 200 hPa) from June 28 – July 2, 2006. PV and omega values are represented by the blue-red color shading. Diabatic heating rate contours are superimposed on the two cross sections; blue contours represent a net diabatic cooling and green represents a net diabatic heating (contour intervals: 0.5 K hr\(^{-1}\)). Cross section location as in inset in Figs. 4.12 and 4.14.
Figure 4.16: Five-day (June 30 to July 4, 2005) daily rainfall accumulation estimates for all observational and reanalysis datasets on a coarsened 1-degree grid for the 4-degree 75-hour record event centered at 23 N, 78 E from the web atlas. The ensemble mean and ensemble RMS of all non-land-confined datasets are in panels o and m, respectively. Color intervals are 47 mm for all plots except for the ensemble RMS (30 mm). Color schemes and red contours are as in Fig. 4.5.
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Figure 4.18: GridSat IR imagery and TRMM 3B42 rain rate (green pixels) from June 30 - July 4, 2005 09:00 UTC centered over Madhya Pradesh. Panels a-e represent 24-hour intervals from the start date.
Figure 4.19: MERRA-2 mean sea level pressure (red contours; interval 2 hPa), 500-hPa heights (cyan contours; interval 20 m), GridSat IR satellite imagery, and MERRA-2 rain rate (green pixels) at 09:00 UTC for June 30 - July 4, 2005. Panel timeline is as in Fig. 4.18.
Figure 4.20: MERRA-2 daily-averaged column water vapor (shaded pixels; units mm) and contours (interval: 5 mm) from June 30- July 4, 2005.
Figure 4.21: MERRA-2 wind speeds (yellow contours; interval: 10 m s$^{-1}$), wind streamlines (cyan contours) and GridSat imagery in sequential order from June 30- July 4, 2005 at 09:00 UTC.
Figure 4.22: Static energy sounding with description as in Fig. 4.10, but for station 42667 (Bhopal, India; VABP) on July 4, 2005 at 00:00 UTC. Because CAPE is either 0 or negligibly low, it does not appear in the figure.
Figure 4.23: Cross section of time-averaged Ertel potential vorticity (units: K kg m\(^{-2}\) s\(^{-1}\)) and potential temperature (units: K; contour interval: 2 K) from 0 to 12 km (1000 to 200 hPa) from June 30 – July 4, 2005. The cross-sectional area is depicted by the pink line in the inset with TRMM 3B42 time-averaged rainrates shown in green pixels.
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Figure 4.25: Cross section of vertical velocity pixels and contours (units: Pa s$^{-1}$; contour intervals: 0.25 Pa s$^{-1}$) from 0 to 12 km (1000 to 200 hPa) from June 30 – July 4, 2005. Dashed contours indicate rising motion while solid contours indicate sinking motion. Cross section inset as in Fig. 4.23.
Figure 4.26: Cross section of time-averaged (a) Ertel potential vorticity and (b) vertical velocity from 0 to 12 km (1000 to 200 hPa). Diabatic heating is superimposed on the two cross sections; blue contours represent a net diabatic cooling and green represents a net diabatic heating (contour intervals: 0.5 K hr\(^{-1}\)). Cross section location as in inset in Figs. 4.23 and 4.25.
Figure 4.27: Five-day (August 1-5, 2005) daily rainfall accumulation estimates for all observational and reanalysis datasets on a coarsened 1-degree grid for the 4-degree 75-hour record event centered at 20 N, 78 E from the web atlas. The ensemble mean and ensemble RMS of all non-land-confined datasets are in panels o and m, respectively. Color intervals are 44 mm for all plots except for the ensemble RMS (30 mm). Color schemes and red contours are as in Figs. 4.5 and 4.16.
Figure 4.28: Total rainfall accumulation (units: mm) from August 1-5, 2006 for native resolution (a) TRMM 3B42 and (b) MERRA-2 products during the east Maharashtra flood. As before, the white box corresponds to the region of interest.
Figure 4.29: GridSat IR imagery and TRMM 3B42 rain rate (green pixels) from August 1-5, 2006 at 18:00 UTC. As in Figs. 4.7 and 4.18, panels a-e represent 24-hour intervals from the start date.
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Figure 4.31: MERRA-2 daily-averaged column water vapor (shaded pixels; units mm) and contours (interval: 5 mm) from August 1-5, 2006.
Figure 4.32: MERRA-2 wind speeds (yellow contours; interval: 10 m s$^{-1}$), wind streamlines (cyan contours) and GridSat imagery in sequential order from August 1-5, 2006 at 18:00 UTC.
Figure 4.33: Static energies diagram as in Figs. 4.11 and 4.22, but corresponding to the station in Nagpur, India (sounding code 42867; VANP).
Figure 4.34: Cross section of time-averaged Ertel potential vorticity (units: K kg m\(^{-2}\) s\(^{-1}\)) and potential temperature (units: K; contour interval: 2 K) from 0 to 12 km (1000 to 200 hPa) from August 1-5, 2006. The cross-sectional area is depicted by the red in the inset with TRMM 3B42 time-averaged rainrates shown in green pixels.
Figure 4.35: Cross section of 3-hourly instantaneous Ertel potential vorticity (units: K kg m$^{-2}$ s$^{-1}$) and potential temperature (units: K; contour interval: 2 K) from 0 to 12 km (1000 to 200 hPa) for August 3, 2006 at (a) 12:00 UTC, August 4 at (b) 00:00 UTC, (c) 12:00 UTC, and (d) August 5 at 00:00 UTC. Cross-sectional location as in inset in Fig. 4.34.
Figure 4.36: Cross section of vertical velocity pixels and contours (units: Pa s\(^{-1}\); contour intervals: 0.25 Pa s\(^{-1}\)) from 0 to 12 km (1000 to 200 hPa) from August 1-5, 2006. Dashed contours indicate rising motion while solid contours indicate sinking motion. Cross section inset as in Fig. 4.34.
Figure 4.37: Cross section of time-averaged (a) Ertel potential vorticity and (b) vertical velocity from 0 to 12 km (1000 to 200 hPa). Diabatic heating is superimposed on the two cross sections; blue contours represent a net diabatic cooling and green represents a net diabatic heating (contour intervals: 0.5 K hr$^{-1}$). Cross section location as in inset in Figs. 4.34 and 4.36.
### Tables

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<th>Variable</th>
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<td>*</td>
<td>Potential temperature</td>
<td>K</td>
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<td>*</td>
<td>Vertical pressure velocity</td>
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<td>*</td>
<td>Diabatic heating due to physics</td>
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<tr>
<td>*</td>
<td>Total column integrated water vapor</td>
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</tr>
<tr>
<td>*</td>
<td>Atmospheric Pressure</td>
<td>hPa</td>
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<tr>
<td>*</td>
<td>Geopotential height</td>
<td>m</td>
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<td>*</td>
<td>Isobaric wind speeds</td>
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<tr>
<td>TRMM 3B42**</td>
<td>Instantaneous rain rate</td>
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</tr>
<tr>
<td>**</td>
<td>Accumulated rain</td>
<td>mm</td>
</tr>
<tr>
<td>GridSat</td>
<td>IR imagery</td>
<td>K (cloud top temperature)</td>
</tr>
</tbody>
</table>

Table 4.1: 2-D and 3-D variables considered for study within the IDV for each case study. The first column states the datasets used. The second column states the variables from each respective dataset. Variables from the same overarching dataset are starred accordingly in the first column (e.g., MERRA-2 is *, TRMM 3B42 is **). The third column represents the variables’ respective units.
<table>
<thead>
<tr>
<th>Case coordinates</th>
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<th>3-day total rainfall (mm)</th>
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<td>August 3, 2006, 18:00</td>
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Table 4.2: Information on the selected case studies from the web atlas, contingent with Fig. 4.3.
Chapter 5: Conclusions and Future Studies

5.1. Summary

Extreme rainfall events, and the complex interplay between the moisture and dynamics involved, is an ongoing challenge for predicting them. In this thesis, a two-pronged approach was utilized in order to investigate the frequency of precipitation extremes as well as to examine a sample set of extreme precipitation cases.

Using 18 years of precipitation data from the TRMM 3B42 dataset, the greatest single-day and three-day precipitation records were identified for each pixel on the TRMM 3B42 grid which spans from 50°S - 50°N for all longitudes. The results found in this thesis are consistent with the results in Mapes (2011) for the following points:

- The mode of record precipitation events occur for rainfall totals of ~100 mm during a single-day record and ~160 mm for a three-day record.
- The densest concentration of high three-day precipitation totals occur over southeast Asia, the Caribbean Sea, and over the Indian Ocean. Records in these areas are roughly equivalent to just one or two months’ worth of rainfall. Over drier areas such as Saharan Africa and the subtropical Pacific Ocean, record rainfalls are equivalent to several months or more of rainfall, even when the magnitude of three-day record is small.
- Most of the record events occurred during the northern and southern hemisphere-respective summer months.
While 1998 was the year in which the greatest total of record events occurred, the bulk of those events were over the near-equatorial Pacific associated with the onset of the 1998 El-Niño phase where observed precipitation was minimal. 2016 saw the lowest number of cases for the whole dataset.

From the global grid of precipitation records, a series of three case studies were selected from events over the Indian subcontinent. Each event was carefully selected so that their times did not coincide with one another. A combination of two and three-dimensional variables obtained from the MERRA-2 dataset were used and visualized in the IDV such that key atmospheric features leading to the development of the highly precipitous event could be elucidated. The most striking results from these case studies are described as follows:

- TRMM 3B42 consistently observed the highest rainfall totals out of all the observation-based products. For all three cases, MERRA-2 was closest in both spatial coverage and total precipitation while other reanalyses showed the most intense precipitation well removed from the ensemble mean location.

- The three cases exhibited similar structures to the typical structure of monsoon depressions as described in Boos et al. (2015) and Hunt et al. (2016). Although some of the synoptically driven mechanisms were not well defined in these cases, a highly moist environment where CWV exceeded 60 mm and lack of steering flows helped to facilitate an environment that was favorable for heavy precipitation in convection. This was especially
evident in BOB02 where CWV exceeded 70 mm and slow storm propagation existed for several days during the case.

- The time-averaged PV-theta cross-section structure does not appear to reveal a discernable bulge in the theta surfaces around a PV maximum, although it is best seen in the lower to middle levels of the atmosphere. The instantaneous cross-sections offer a more robust signature in the lower levels, suggesting that the low-level adiabatic ascent may have been responsible for the growth of secondary convection, consistent with RJ90 and Trier et al. (2000).

- Vertical ascent (omega) and diabatic heating profiles resembled each other for each respective case. BOB02 was a case that exhibited strong vertical ascent with the greatest diabatic heating rooted above the boundary layer.

5.2. Suggestions for Future Studies

The causes of rare events are always more than singular. In order to truly understand the essence of the combinatorics behind these factors, future work may begin with analyzing the climatological mean and standard deviation of relevant variables studied in this thesis. For instance, calculations of the information behind such anomalies and standard deviations will highlight the actual magnitude of these extremes.

An overarching goal of future work is to examine in much greater detail the mechanisms that drive extreme precipitation events across various regions of the world. Since the dynamics and thermodynamics are slightly different by
geographical region, one prong of study would involve separating the precipitation records by geographical region. Some relevant questions behind this thinking involve the following:

- Recalling the use of the observations-reanalysis intercomparison plot from the cases in chapter 4, *how would the distribution of the rainfall totals look for cases where existing records were set over the sea?* Or, *how would the rainfall distributions look like for cases where the existing records were set over land?* These questions help to characterize the composite distribution for land and sea cases. In further detail, cases of records involving defined mesoscale and synoptic-scale features such as tropical cyclones or mid-latitude frontal systems will characterize these records by feature, adding more depth to the analysis of records.

- *What are the key differences between different continents or regions? Does one reanalysis model consistently perform better or worse compared to a reference product such as TRMM 3B42 or another?* It was found that there were various differences between each product and some products resolved more precipitation than others or resolved a peak in precipitation total in different locations. But this thesis only examines a set of three cases specific to one geographical region, so it is premature to speculate on the overall reliability of each product based on such a small sample size. Regionalizing all of the 4-degree record events (Fig. 3.1) should offer more information on the quality of precipitation estimation between products.
Another prong of study involves modeling of these extremes in order to address the ongoing challenge of predictability of extreme rainfall events. Since many of the most important processes governing extreme rainfall events are difficult to observe, using numerical modeling experiments will be a useful approach in further closing the gaps in understanding the key mechanisms involved. Some of the open questions that future projects could entail include the following:

- Under the guiding light of the RJ90 MCV theory, how can a model-simulated PV anomaly affect the genesis and sustenance of deep, moist convection responsible for intense rainfall? Even small perturbations (both symmetric and axisymmetric) to the control PV field may lead to a significant change in the system’s development through the model run.

- Beyond the PV-theta argument raised by RJ90 (Fig. 4.1), what other factors govern the growth and sustainability of such rainfall extremes? As mentioned previously, extreme rain events are governed by a complex interplay of ingredients that coincide at the same time. Analyzing factors that are consistently responsible for heavy precipitation in convection deserves to be investigated further.

Future works pertaining to the work presented here will continue on the path of investigating the primary mechanisms involved with the genesis and sustenance of deep convection leading to extreme rainfall totals. The guiding aim will be to improve on the predictability of such events so that life and property may potentially be spared.
References


King, A. D., N. P. Klingaman, L. V. Alexander, M. G. Donat, N. C. Jourdain, and P. Maher, 2014: Extreme Rainfall Variability in Australia: Patterns, Drivers,


Appendix

Figure A1: Color-pixel plots of precipitation amounts in the TRMM 3B42 product domain in 1998-2016 at the native 0.25° resolution for (a) record 3-day accumulation (units mm), (b) climatological annual rainfall divided by a factor of 12 (units mm month⁻¹), and (c) the ratio of maxima to climatology taken from a/b.
Figure A2: As in Fig. A1, but for the coarsened 1° resolution.