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INVESTIGATING ENVIRONMENTAL INFLUENCES BEHIND SUMMERTIME
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NORTHERN GREAT PLAINS

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A THESIS APPROVED FOR THE
SCHOOL OF METEOROLOGY

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Abstract

Extreme rainfall events (EREs) in the United States are known to cause significant and deadly flash flooding events. Despite advancements in extreme rainfall forecasting in recent decades, owing to the increase in numerical weather prediction model skill and resolution, extreme precipitation forecasting remains a significant challenge and motivates NOAA’s Precipitation Prediction Grand Challenge initiative. One particularly difficult forecasting problem involves summertime nocturnal convection, which has been shown to produce a high percentage of summertime (June–August) EREs. This nighttime occurrence poses additional complications when messaging warnings and potential impacts to at-risk communities. One of the primary forecast challenges is understanding how environmental variables control the predictability of nocturnal EREs and their severity, as the events are strongly influenced by interactions between the synoptic-scale and mesoscale processes that can be difficult to accurately represent and predict. To further investigate this, this study investigates the large-scale environmental conditions that favor summertime nocturnal EREs in the Northern Great Plains of the United States using self-organizing maps (SOMs). SOMs are used to cluster EREs of varying severity, i.e., extreme and less extreme, using large-scale synoptic meteorological variables to understand how environments vary across extreme rainfall events and inform the predictability of extreme rainfall.

First, EREs are identified using a high-resolution database, that uses NCEP Stage IV 6-hour quantitative precipitation estimates (QPE) and average recurrence interval (ARI) thresholds from NOAA Atlas 14. Within this database, events are defined by 6-hour Stage IV QPE accumulations exceeding the 10-year and 100-year ARI thresholds. A SOM is then trained on 500-mb geopotential height anomalies associated with

the resulting more-extreme (100-year ARI exceedance) and less-extreme (10-year ARI exceedance) events. Following the SOM training, events assigned to each SOM pattern (node) are then separated into more-extreme and less-extreme events and used to generate composites of additional atmospheric variables, such as total column water vapor (TCWV), 850-mb winds, 850-mb temperature advection, vertically integrated water vapor flux (VIWVF), and VIWVF convergence. By doing this, we can identify atmospheric patterns associated with the different SOM nodes and allow for the investigation of whether certain patterns are associated with more-extreme rainfall events compared to less-extreme events. This thesis will compare the clustered environmental conditions between more-extreme and less-extreme rainfall events and detail a few of the node regimes resulting from the trained SOM. Results from the SOM framework are also compared to full-domain bulk composites of all more-extreme and less-extreme event environments to evaluate the added value of the SOM approach in resolving distinct synoptic environments compared to traditional composite approaches. In addition, the large-scale environmental evolution preceding the events is analyzed.

Results show that both the bulk and node regime composites reveal a common set of environmental features associated with summertime nocturnal extreme rainfall events in the Northern Great Plains that are known to be favorable for extreme rainfall. The bulk composites also identify some key environmental differences between more-extreme and less-extreme events that were consistently present across the evaluated SOM node regimes. However, the bulk composites do not capture the full range of environmental differences between more-extreme and less-extreme events. The implications of these new findings will be shared to advance our collective knowledge of extreme rainfall events and how SOMs can be useful tool in identifying common patterns.

Chapter 1

Introduction and Motivation

Summertime extreme rainfall events (EREs) pose significant societal and economic risks across the United States, often resulting in flash flooding, infrastructure damage and lives lost. Flash flooding has long been recognized as one of the nation’s most dangerous natural disasters. By the 1950s, flash flooding had become the United States’ most significant natural disaster, and a large percentage of annual flood-related fatalities have been attributed to flash floods (e.g., Maddox et al., 1979; Zevin, 1994). Flooding has continued to remain a major weather hazard, ranking as the second-deadliest weather-related fatality across the United States, Puerto Rico, Guam, and the Virgin Islands (National Weather Service, 2026). Because flash floods often develop rapidly with limited warning time, especially during the nighttime hours when public awareness is reduced, improving the understanding and prediction of EREs remains critical for reducing devastating impacts. This need for accurate forecasting of flash floods becomes even more crucial as the impacts of climate change become more imminent. In a warming climate, an increase in extreme rainfall rates is likely as water vapor is projected to increase in the atmosphere (e.g., U.S. Global Change Research Program, 2023; Dougherty and Rasmussen, 2019). Westra et al. (2014) found a clear relationship between atmospheric temperatures and extreme rainfall intensity, particularly for sub-daily extreme rainfall events, as they were found to be more sensitive to local changes in atmospheric temperature. Unable to make direct conclusions based on regional differences, Westra et al. (2014) found that it is very likely that there will be considerable

differences in daily extreme rainfall depending on geographic location. Understanding possible regional differences is important as they could help to determine where flash flooding risks may intensify and how forecasting strategies might need to adapt to regionally distinct extreme-rainfall environments. Therefore, further understanding of the common atmospheric processes associated with regional EREs remains crucial to improving flash flooding forecasts and mitigating future societal impacts.

Flash flooding is a result of prolonged rainfall and high rainfall rates over a given area. Doswell et al. (1996) developed a basic approach to understanding the ingredients and meteorological processes behind forecasting the potential for flash flooding. They advanced the idea that the heaviest precipitation occurs where the rainfall rate is the highest for the longest period. They explain that the duration of a flash flooding event can be affected by system movement speed, system size, and the variations of rainfall intensity within a system. Precipitation rate considers aspects such as precipitation efficiency (the ratio of mass of water falling as precipitation to the influx of water vapor mass into the cloud) and vertical moisture flux (the movement of water vapor in upward or downward direction) (Doswell et al., 1996). Maddox et al. (1979) studied 151 intense precipitation events that all caused flash flooding across most of the conterminous United States and identified three types of flash flooding events: mesohigh, western, and frontal. Despite different event types, they found that certain features were common among all. All events were associated with convective storms and occurred in environments characterized by very high surface dewpoint temperatures, abundant moisture throughout the troposphere, and vertical wind shear that was generally weak to moderate. Maddox et al. (1979) also noted that subtle mid-tropospheric troughs played a role in initiating convection, and the resulting storm systems tended to be located near the position of a broader mid-level ridge. They also found that the highest amount of flash flooding events occurred in the summer months

of June, July, and August.

Despite better understanding of how heavy precipitation forms and recent advancements in convective-allowing numerical modeling, forecasting heavy precipitation and flash flooding remains a persistent challenge for forecasters across the United States (Nielsen and Schumacher, 2016). A past study by Nielsen and Schumacher (2016) showed that the primary error in convective-allowing models is the location of the heaviest rainfall, not rainfall timing. This suggests that models may reasonably capture the evolution of extreme rainfall producing storms but struggle to capture where these storms initiate, travel, and produce the maximum precipitation. Understanding these errors is important because differences in regional characteristics, such as topographic effects, moisture sources, and diurnal cycles, can affect how large-scale forcing translates into extreme rainfall. The study by Nielsen and Schumacher (2016) showed that diurnal cycles in different regions will cause different interactions between the convective and synoptic scale forcings. This implies that forecast errors are not the same across the CONUS but instead are going to be regionally distinct. Therefore, improving the understanding of extreme rainfall synoptic and mesoscale environments is important in strengthening extreme rainfall forecasts and removing regionally biased model errors.

Recent work has been focused on building long-term, high-resolution datasets of EREs to better characterize the environment in which they occur. A recent study by Chiappa et al. (2024) developed a database of short-duration summertime (June-August) EREs over the period of 2003 through 2023 across the Central and Eastern United States. They used hourly Stage IV data to define points where precipitation accumulation exceeded a location’s 10-year average recurrence interval (ARI) threshold. This type of database allows for structured analysis of the synoptic and mesoscale environments behind EREs, which are often difficult to summarize from case studies

alone. Using this framework, Chiappa et al. (2024) found strong connections between EREs and low- to mid-levels of moisture over the Central Great Plains and enhanced moisture flux from the Gulf of Mexico. One of their main conclusions was that a high percentage of the total summertime extreme rainfall events across the studied regions were caused by nocturnal convection. Their conclusion aligned with past studies that found that nocturnal mesoscale convective systems (MCSs) are the most common producers of summertime extreme rainfall events (Stevenson and Schumacher, 2014). However, numerical modeling of nocturnal convection continues to be a major challenge due to the difficulties in accurately representing the initiation, structure, and evolution of nocturnal convection in weather and climate models (e.g., Geerts et al., 2017; Tang et al., 2021). This challenge highlights the continued need to better understand nocturnal EREs and the environments in which they occur to improve the representation of nocturnal convection in numerical models and reduce forecast errors associated with extreme rainfall intensity and placement.

Understanding the mechanisms that drive summertime nocturnal rainfall events across the U.S. Great Plains is critical for improving the prediction of nocturnal extreme rainfall and flash flooding in this region. A key contributor to summertime nocturnal across the Great Plains is the Great Plains low-level jet (LLJ), which develops in the overnight hours in response to differential cooling. A study by Higgins et al. (1997) looked at the influence of the Great Plains LLJ and moisture transport on nocturnal summertime precipitation over the central United States by comparing assimilated datasets to station observations of precipitation, winds, and specific humidity. They found that during summer, rainfall over the central Great Plains peaks and is closely associated with an increase in nocturnal convective activity. They also documented that the nocturnal maximum in extreme rainfall is consistent with the peak of the low-level jet intensity. During the summer months, it was found that in

the vicinity of the LLJ, there is the strongest influx of moisture onto the continent from the Gulf of Mexico. This not only provides storms with a proficient moisture source for convection but also can enhance low-level moisture convergence, promoting upward vertical motion. This study highlights the importance of moisture transport in both storm initiation and rainfall amount. A more recent study by Parish et al. (2020) related the LLJ and its association with nocturnal destabilization in the Great Plains by demonstrating this process through a case study from the Plains Elevated Convection at Night field program. Their results demonstrated that strong LLJ events were not only associated with enhanced moisture transport but also associated with enhanced warm air advection, which contributes to steepened lapse rates and decreased stability above the jet. These processes work together to create a favorable environment for initiation and maintenance of nocturnal convection and nocturnal EREs.

While these studies provide a physical understanding of the environments favorable for nocturnal extreme rainfall, they primarily rely on case studies and composite analysis, which may not fully capture the diversity of synoptic and mesoscale patterns. This motivates the use of clustering techniques that can identify recurring atmospheric patterns, such as self-organizing maps. A self-organizing map (SOM) is an artificial neural network that is often used to cluster events into a predetermined number of nodes with common spatial structures in the parameters that are being analyzed (e.g., Anderson-Frey et al., 2017; Kohonen, 1982). First developed by Kohonen (1982), these maps allow for easier visualization and identification of common patterns in reanalysis data that are hard to detect directly in the full-resolution fields. SOMs preserve the topological structure of high-dimensional data while reducing it to a lower dimension which provides additional detail on environments compared to traditional average composites (Kohonen, 1990). This topology-preserving feature has been shown to be advantageous compared to methods such as k-means or hierarchical clustering as SOMs also preserve

neighborhood relationships between clusters (e.g., Chen et al., 2010; Philippopoulos et al., 2014). Topology preservation provides insight into how atmospheric environments transition between related circulation regimes rather than separating them into unrelated clusters. Past studies have used SOMs to identify and classify similar synoptic atmospheric circulation patterns over different regions (Hewitson and Crane, 2002) and even to analyze extreme weather and rainfall events (e.g., Hong et al., 2005; Laiho et al., 2024).

Despite extensive research on extreme rainfall events, a complete understanding of the processes that distinguish varying degrees of extreme rainfall intensity remains incomplete. While the Great Plains LLJ is a well-known contributor to nocturnal moisture transport and convection initiation, uncertainties remain regarding how synoptic and mesoscale forcing processes ultimately distinguish between more-extreme and less-extreme events. Many studies analyze environmental differences between non-extreme and extreme rainfall events, which often treats extreme events as a single category (e.g., Ding et al., 2024; Veeram et al., 2026). This thesis aims to explore the large-scale environmental and meteorological influences behind nocturnal summertime extreme rainfall events in Northern Great Plains of the United States. This will be done by investigating the environmental differences between more extreme and less extreme rainfall events. While previous studies have identified key ingredients favorable for EREs across the entire central CONUS, the presence of these ingredients does not fully explain the variability in intensity of extreme rainfall. Additionally, focusing on a subregion of the central CONUS is important because the mechanisms that produce extreme rainfall vary spatially. Examining environmental differences between more extreme and less extreme events therefore provides insight into environmental patterns associated with nocturnal summertime EREs in this region and helps identify the environmental controls that distinguish between extreme types.

Traditional composites and SOMs are both used to characterize large-scale environmental characteristics associated with nocturnal summertime extreme rainfall events, with SOMs providing additional insight into distinct and recurring patterns beyond what is captured in mean composites. This thesis strives to answer the following questions:

1. What distinct and recurring large-scale atmospheric patterns are associated with nocturnal summertime extreme rainfall events in the Northern Great Plains?
2. How do large-scale atmospheric patterns differ between more-extreme and less-extreme rainfall events?
3. What additional insights are gained from self-organizing map analysis compared to traditional composite analysis in characterizing nocturnal summertime extreme rainfall events?
4. How does the synoptic environment evolve in the days preceding more-extreme and less-extreme nocturnal summertime extreme rainfall events?

Chapter 2

Data and Methods

2.1 Stage IV Precipitation Data

The National Center for Environmental Prediction’s (NCEP) Stage IV quantitative precipitation estimates (QPEs) are used for this study to identify EREs. This product consists of precipitation accumulation intervals of hourly, 6-hourly, and 24-hourly on a 4-km x 4-km horizontal grid. QPEs are produced by combining Next Generation Radar Data (NEXRAD) radar-estimated rainfall and hourly rain gauge data (Nelson et al., 2016). NCEP uses a mosaicking technique to combine the two rainfall datasets across 12 different River Forecasts Centers (RFCs) in the continental United States. Before the data are interpolated onto a common hydrologic rainfall analysis project (HRAP) grid and mosaicked into a unified 4-km gridded dataset over the CONUS, there is manual and automatic quality controls done at each individual RFC. Because of these quality control methods, this product is widely used for hydrometeorological analysis and comparison to other precipitation datasets. The Stage IV dataset used in this study is the additional quality-controlled version from Schumacher and Hill (2026). This additional quality control includes statistical outlier identification, detection of spatial inconsistencies across precipitation datasets, residual analysis, and manual visual inspection, which makes the data more reliable when identifying extreme rainfall events.

In addition to the quality control methods, the Stage IV dataset has many known strengths that make it a suitable dataset for this thesis. Nelson et al. (2016) provides strong evidence that Stage IV precipitation estimates perform well at high rainfall rates. This is especially important when forecasting hazardous weather events. Notably, this study also shows that Stage IV data does a better job at identifying convective precipitation compared to rain gauge observations alone.

However, there are also known limitations that should be considered alongside the strengths of this dataset. It was shown that with increasing rain rates, there is a higher percentage of missed events. This indicates that there may be challenges in capturing extreme rainfall events. On a broader scale, radar-based precipitation estimates have inherent limitations, due to things such as ground clutter and attenuation, that RFCs work to reduce, but cannot be fully eliminated. Additionally, the quality control of hourly rain gauge observations at each RFC is automated and has been shown to occasionally allow erroneous gauge reports to continue through. Collectively, the Stage IV dataset remains a reliable and widely used product for identifying extreme rainfall events but should be used with caution of these systematic limitations.

2.2 Identification of Extreme Rainfall Events

This work uses the Stage IV 6-hourly precipitation intervals to focus on shorter duration EREs. We identify nocturnal summertime EREs events during the months of June, July, and August (JJA) for the 6-hour periods of 0000 to 0600 UTC and 0600 to 1200 UTC that represent local nighttime hours (7pm to 7am local). The events are identified over a 21-year period from 2002 to 2023, which was the maximum period permitted by the availability of data used in this study. The study domain encompasses the Northern United States Great Plains, defined by latitudes 40°N to 49°N and longitudes 105°W

to 93°W (Fig. 2.1). The western boundary of 105°W is chosen to avoid the Rocky Mountains, which can introduce additional variability in atmospheric fields due to the complex terrain.

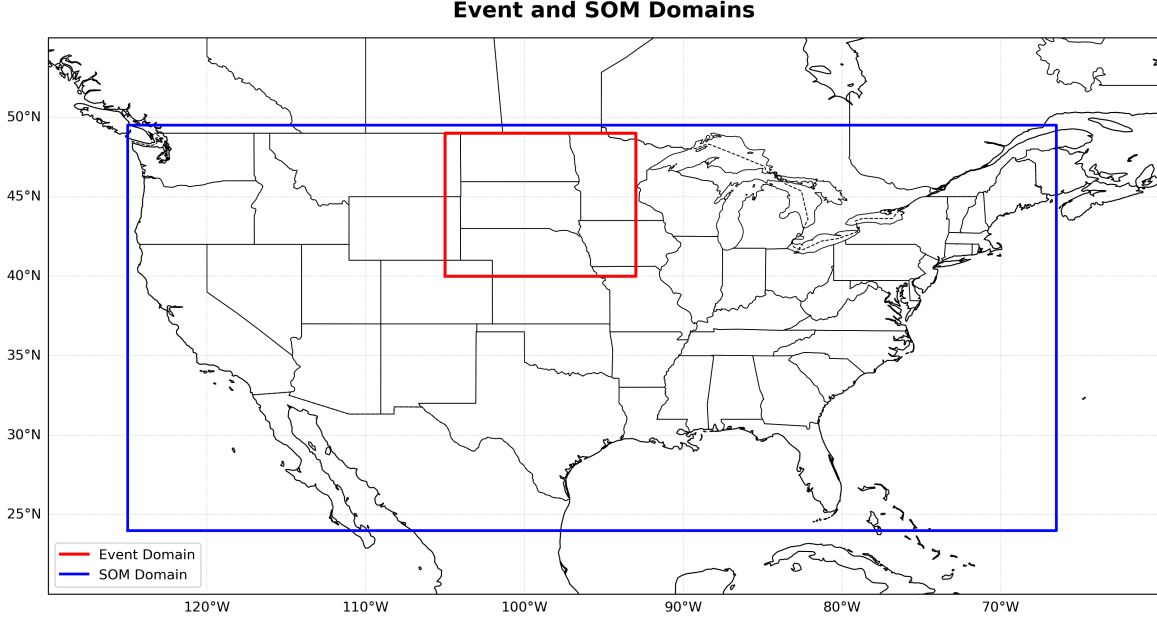


Figure 2.1: Spatial domains used in this study for event identification and SOM analysis. The red box indicates the event domain, used to define and extract nocturnal precipitation events. The blue box outlines the larger SOM analysis domain.

The National Oceanic and Atmospheric Administration (NOAA) Atlas 14 precipitation frequency estimates are used in this study as a reference for identifying areas of extreme rainfall (National Weather Service, 2025). The frequency estimates are determined by using long-term rain gauge observations and provide estimates of rainfall magnitude based on different recurrence intervals. Atlas 14 precipitation frequency estimates are derived using extreme value statistical methods applied to annual maximum precipitation series and represent the expected rainfall depth associated with a given return period and accumulation duration. For example, a 1-hour, 100-year recurrence interval corresponds to the 1-hour rainfall depth with a 1% annual exceeded probability.

This work compares more-extreme events and less-extreme rainfall events to investigate the environments behind nocturnal summertime rainfall. Because of this, 100-year and 10-year average recurrence intervals (ARIs) are used for identification. Events are identified using consistent point-based thresholding of Stage IV QPEs within the event domain. More extreme events are identified as 6-hour periods where there is at least one point that exceeds the 100-year ARI (Fig. 2.2a), while less-extreme extreme events are those where at least one point exceeds the 10-year ARI (Fig. 2.2b) over the same duration, but the 100-year ARI has not be exceeded. For periods that have both 10-year and 100-year events, the period is considered a more-extreme event period. This event identification approach is similar to that of Chiappa et al. (2024), which also used Stage IV data and the NOAA Atlas 14 dataset to define extreme rainfall events. However, Chiappa et al. (2024) used derived 12-hour accumulation precipitation and only considers exceedance of 10-year ARIs.

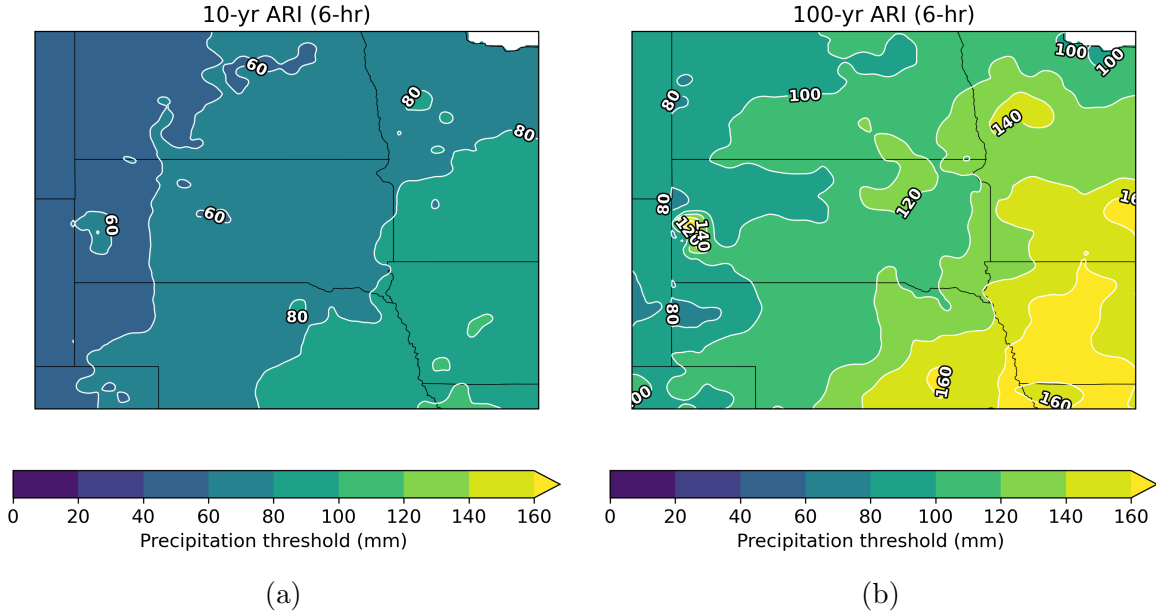


Figure 2.2: Threshold precipitation amounts (mm) associated with the (a) 10-year and (b) 100-year 6-h average recurrence intervals, shown over the Northern Great Plains (40–49°N, 105–93°W). Filled contours indicate the precipitation depth required to exceed each ARI threshold.

2.3 ERA5 Reanalysis Data

The European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis v5 (ERA5) is used in this study as input data for atmospheric composites and self-organizing maps. ERA5 integrates observational data with model simulations to produce a comprehensive and internally consistent global atmospheric dataset (Hersbach et al., 2020). This makes ERA5 particularly useful for composite and SOM analysis, which require gridded, physically consistent data cross time and space. ERA5 provides hourly atmospheric data at $0.25^\circ \times 0.25^\circ$ horizontal grid spacing on multiple pressure levels which makes it ideal for the investigation of synoptic environments. ERA5 is chosen because it is well suited for examining the large-scale atmospheric conditions associated with extreme rainfall events, as no equivalent reanalysis product is available at the mesoscale and storm-scale.

From ERA5, several atmospheric variables are selected to characterize the synoptic environments associated with nocturnal extreme rainfall events, with 500-mb geopotential heights used as a primary descriptor of the large-scale synoptic pattern. Building on this synoptic framework, additional variables are chosen to represent key physical properties that have been shown in previous studies to influence extreme rainfall.

The nocturnal low-level jet (NLLJ) is represented using 850-mb wind fields and is used to assess the strength and positioning of low-level wind speeds at the time of the events relative to the event domain. This feature is also commonly linked to enhanced moisture transport into the central Great Plains and is expected to influence differences in moisture availability and transport between extreme types (Higgins et al., 1997). Total column water vapor (TCWV), which represents the amount of water vapor there is in column extending from the surface of the Earth to the top of the atmosphere, is used to investigate moisture variability during events and how

this differs between extreme types. 850-mb temperature advection is used as a proxy for a lower tropospheric destabilization method, synoptic-scale forcing for ascent, and transportation of warm air by the LLJ. Lastly, vertically integrated moisture flux and vertically integrated moisture flux convergence is investigated to quantify the magnitude and pathways of moisture transport into the event domain, as well as regions of moisture accumulation associated with dynamically forced ascent.

2.4 Composite and Composite Difference Analysis

To obtain an initial understanding of what the synoptic environments look like during nocturnal summertime extreme rainfall events, more extreme and less extreme event environments are composited separately using multiple variables introduced in section 2.3. For each 6-hour event period, this analysis uses the middle hour of the period to represent the conditions at the time of events. All event periods are then combined into a single composite for each event category. Composite differences are then calculated to evaluate the synoptic differences between the two categories. Statistical significance is then assessed to identify robust differences between composites. Significance is determined using non-parametric bootstrap resampling method at the 95% confidence level. This method is chosen because it accounts for unequal sample sizes, which we have, and does not rely on assumptions of normality.

2.5 Self-Organizing Map (SOM) Methodology

500-mb geopotential heights anomalies are used to train the self-organizing maps in this study to investigate the mid-tropospheric flow patterns and synoptic patterns associated with summertime nocturnal extreme rainfall events. The ERA5 reanalysis

provides geopotential (Φ , $\text{m}^2 \text{s}^{-2}$) at 500 mb and not geopotential height directly. 500 mb geopotential heights are calculated by dividing the 500 mb geopotential by acceleration due to gravity (9.80665 m s^{-2}). The analysis domain spans over the entire continental United States, defined by latitudes 24°N to 49.5°N and longitudes 125°W to 66.5°W , to show a full synoptic picture of the environments occurring during these events (Table 2.1). Previous studies have shown that training on 500 mb geopotential heights anomalies as the middle-level atmospheric pressure patterns provides crucial information for identifying conditions in the low-level atmosphere and at the surface (Ford and Schoof, 2017).

To construct the input dataset, 6-hour analysis periods are defined as 00–06UTC and 06–12UTC based on Stage IV QPE timestamps. To reduce repetitive sampling of similar environments while capturing the central tendencies of each period, only the midpoint hours, 03UTC and 09UTC, are included in the analysis. The-500mb geopotential height climatology is computed by averaging all available 03UTC and 09UTC time steps across the full 21-year record, not just those corresponding with rainfall events, resulting in a climatological mean field representative of nocturnal conditions. Anomalies are then calculated by subtracting the climatological mean field from the 03 UTC and 09 UTC ERA5 500 mb height fields at each grid point during EREs.

The SOM is trained by repeatedly presenting the 500-mb geopotential height anomaly fields to the map. For each input field, the SOM identifies the node that most closely matches the pattern (the best-matching unit). That node, along with nearby nodes, is then adjusted to become more like the input field. This process is repeated many times over the full dataset. As this is done, the SOM slowly organizes similar patterns into neighboring nodes, preserving the structure of patterns in the data. The number of patterns is predetermined by the layout chosen.

The self-organizing map is trained using the open-source python package MiniSom

(Vettigli, 2018), which has been used in past studies to relate synoptic patterns to different types of weather phenomena (e.g., Wang et al., 2022; Xie et al., 2024). To train the SOM, different parameters need to be set which include the learning rate, the neighborhood radius (sigma), iterations, and layout. For the SOM trained in this study, several combinations of learning rate, neighborhood radius, iterations, and layouts were tested. An analysis of the SOMs quantization error (QE) and topological error (TE) is conducted using all variations to assess the performance of the trained SOM. The QE measures how well the nodes represent the data and the TE measures how well the SOM preserves the neighborhood relationships. When determining the optimal SOM configuration, a balance was sought between the two error types because minimizing QE and TE simultaneously is not always possible. For this SOM, improvements in one error metric led to a decline in the other. The radii tested were 0.01, 0.1, 1, and 2 and the learning rates tested were 0.4 and 0.5. For the SOM trained in this study, the best learning rate was found to be 0.4 and the best radius was found to be 1. After finding this, a table was made to look at different variations of layouts and iterations using this radius and learning rate. Table 2.1 shows QEs and TEs calculated for every variation. The layouts tested were 3x3, 4x4, and 5x5 and the iterations tested were 5000, 10,000, and 20,000. A 4x4 SOM trained with 5,000 iterations was found to provide the best compromise quantitatively. However, the 4x4 layout produced some nodes with nearly identical synoptic patterns, limiting its utility for distinguishing forecast-relevant environments. The 3x3 configuration, while coarser, better separated the dominant regimes and therefore provided a more practical framework for interpreting and communicating synoptic patterns associated with extreme rainfall. Some other parameters set were neighborhood function and the random seed. The gaussian neighborhood function is applied which helps preserve topological relationships by ensuring that updates to the Best Matching Unit (BMU)

of one node are smoothly propagated to the other nodes. The random seed (42) helps to assure that the SOM is reproducible to prevent variability in SOM initialization and differences in node organization and patterns between runs.

map_size	learning_rate	sigma	iterations	quantization_error	topographic_error
(3, 3)	0.400	1.000	5000	108.839	0.194
(3, 3)	0.400	1.000	10000	109.640	0.263
(3, 3)	0.400	1.000	20000	109.667	0.266
(4, 4)	0.400	1.000	5000	100.803	0.183
(4, 4)	0.400	1.000	10000	100.575	0.464
(4, 4)	0.400	1.000	20000	100.976	0.352
(5, 5)	0.400	1.000	5000	95.793	0.324
(5, 5)	0.400	1.000	10000	94.848	0.466
(5, 5)	0.400	1.000	20000	94.925	0.411

Table 2.1: Quantization and topographic errors for Self-Organizing Map (SOM) experiments across varying map sizes and training iterations.

2.6 SOM Composite and Composite Difference Analysis

Once the SOM is trained, all events (both more-extreme and less-extreme) are individually assigned to a BMU based on the similarity of its 500-mb anomaly field to the patterns that the SOM found. By doing this, each event is assigned to one of the 9 nodes. Node-based composites are then created for each variable using the events assigned to each SOM node. Like the SOM training, only the midpoint hours, 03UTC and 09UTC are used for this analysis to avoid repetitive data. The variables being used for this were discussed previously in section 2.3.

This analysis is also performed for selected 500-mb geopotential height anomaly fields from the days preceding events, including the day prior to event onset. Data are used at the same times, 03 UTC and 09 UTC, a day before and three days before

depending on the event timing. Composites were then constructed and compared to each other and with the 500-mb anomalies at event time to evaluate the evolution of the 500-mb large-scale setup.

After making composites for all events assigned to a node, the more-extreme and less-extreme events are separated and composited independently. Composite differences are calculated to evaluate differences between the two extremes. Statistical significance is assessed using the non-parametric bootstrap resampling method discussed in section 2.4. These composites and composite differences are then compared to the full dataset composites in section 2.4 to evaluate the added value of the SOM approach in resolving and reorganizing of distinct synoptic patterns that may not get picked up in traditional composites.

Similarly, 500-mb geopotential height anomaly field composites are made for select days preceding more-extreme and less-extreme events individually. Composite differences are also calculated to compare the evolution of the 500-mb large-scale set up between extreme types. In addition, statistical significance is assessed using the same non-parametric bootstrap resampling method.

2.7 Event Distributions Within SOM Nodes

To isolate structural differences in more-extreme and less-extreme event composite structures, the distribution of less-extreme events and more-extreme events are plotted separately within the event domain. Because of multiple ARI exceedance points existing within an event period, one representative point is selected per period and per extreme type. To avoid bias, one point is randomly selected from each event period using random sampling without replacement.

Chapter 3

Results

3.1 More-Extreme and Less-Extreme Bulk Composites and Composite Differences

3.1.1 500-mb Geopotential Heights and Anomalies

First, traditional composites of geopotential height anomalies of 500-mb are evaluated for events more-extreme and less-extreme (Fig. 3.1a and 3.1b). The large scale circulation associated with both extremities is characterized by a mid-level trough over the western CONUS and a mid-level ridge over the central and eastern CONUS. This positions the event domain downstream of the 500-mb trough axis and just upstream of the 500-mb ridge axis, which supports southwesterly mid-level flow over the event domain. This 500-mb setup is consistent with the setup in Chiappa et al. (2024), who found that active ERE summers, which favored an increased frequency of events over the central CONUS and Ohio Valley, are commonly associated with enhanced mid-level ridging over the eastern CONUS.

One notable difference between the environments is the magnitude of the 500-mb geopotential height anomalies (Fig. 3.1c). More-extreme events exhibit stronger negative and positive geopotential height anomaly values, indicating a more amplified large-scale mid-level flow pattern compared to less-extreme events. The stronger

trough-ridge amplification suggests a more dynamically active synoptic environment during more-extreme events. Additionally, these geopotential height anomaly differences are statistically significant, indicating that the stronger amplification is robust and not due to random variability.

500-mb Geopotential Height Anomalies Composites and Composite Difference

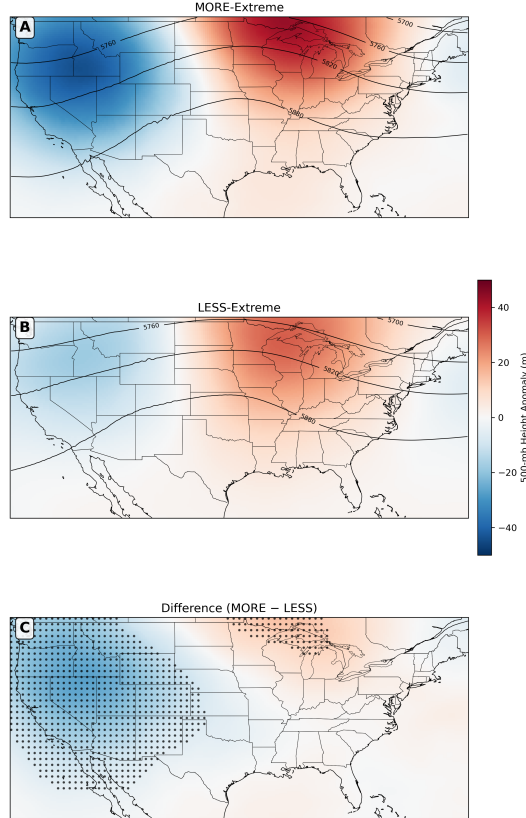


Figure 3.1: Composite 500-mb geopotential height anomalies (shaded; m) and corresponding 500-mb geopotential height contours (black lines) for (a) more-extreme and (b) less-extreme nocturnal extreme rainfall events. (c) Composite difference (more-extreme minus less-extreme) with stippled regions indicating statistically significant differences at the 95% confidence level.

3.1.2 850-mb Wind Speeds

Initial examination of the 850-mb wind composites reveals that more-extreme and less-extreme events exhibit broadly similar spatial patterns, making clear visual differences

between the two extreme events difficult to immediately distinguish (Fig. 3.2a and 3.2b). Both environments exhibit enhanced 850-mb NLLJ extending across the Great Plains, transporting air northward from the Gulf of Mexico. The event domain is located near the nose (area of maximum wind speeds) of the 850-mb NLLJ. This positioning is important to note as the NLLJ is often associated with enhanced low-level convergence near the nose, which can promote upward vertical motion and convection development (Zhong et al., 1996). This enhanced area of the NLLJ weakens and decelerates over the event domain for both event types, suggesting the transportation and accumulation of warm-moist air into the region, as well as potential low-level convergence as a result of this accumulation. This interpretation is consistent with findings from Walters and Winkler (2001), who showed that convergence downstream of the LLJ contributes to convective precipitation.

To identify if there are any significant differences between the two event types, an 850-mb wind composite difference is calculated (Fig. 3.2c). A broad area of positive 850-mb wind speed differences is present in the regions where the 850-mb NLLJ is located in both composites, indicating that 850-mb winds during more-extreme events are approximately 0.5 m s^{-1} than during less-extreme events. In addition, there is an area of enhanced positive 850-mb wind speeds in the more-extreme events over most of the event domain. Most of this area is also statistically significant, indicating that over the event domain, 850-mb wind speeds are consistently stronger during more-extreme events than during less-extreme events. These results suggest a stronger advection of warm, moist air up from the Gulf of Mexico, likely increasing moisture availability within the event domain. The stronger 850-mb flow may also contribute to a more favorable environment for ascent through enhanced moisture transport and downstream convergence for more-extreme events within the event domain.

850-mb Winds Composites and Composite Difference

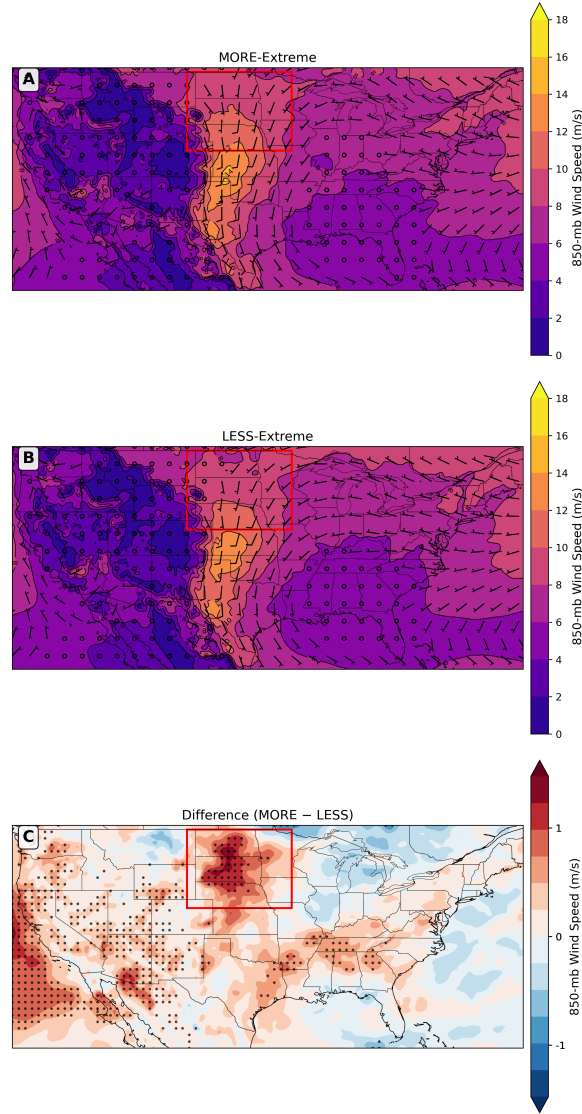


Figure 3.2: Composite 850-mb wind speeds (shaded; m s^{-1}) and wind barbs for (a) more-extreme and (b) less-extreme nocturnal extreme rainfall events. (c) Composite difference (more-extreme minus less-extreme) with stippled regions indicating statistically significant differences at the 95% confidence level. The red box denotes the event domain.

3.1.3 850-mb Temperature Advection

Both the more-extreme and less-extreme events exhibit warm air advection extending over most of the event domain (Fig. 3.3a and 3.3b). This is consistent with the findings

in the 850-mb wind speed composites, which suggests that an enhanced 850-mb NLLJ may be transporting warm air northward from the Gulf of Mexico. This temperature advection pattern suggests widespread low-level ascent for both event types, as warm air advection at 850-mb is commonly associated with synoptic-scale ascent through thermal advection processes. In addition, warm-air advection can contribute to the destabilization of the lower to mid troposphere, supporting the development and/or intensification of convection. This pattern suggests a possible source of synoptic forcing for the extreme rainfall events located in the event domain.

When looking at the 850-mb temperature advection composite differences within the event domain, there are identifiable patterns between the two extreme types (Fig. 3.3c). The northern portion of the event domain exhibits stronger warm air advection during more-extreme events. This pattern suggests that more enhanced warm air advection associated with more-extreme events may extend farther north compared to less-extreme events. This pattern also aligns with the stronger 850-mb wind speeds shown Figure 3.2c which could be helping to enhance the warm air advection farther north. There are also smaller areas of localized warm air advection and cold air advection throughout the rest of the event domain, possibly reflecting areas of smaller-scale variations of low-level flow and thermal gradients. However, none of the differences are statistically significant, indicating that although there are identifiable patterns of mean differences between the extreme intensities, these differences may be caused by random variability and not a consistent pattern.

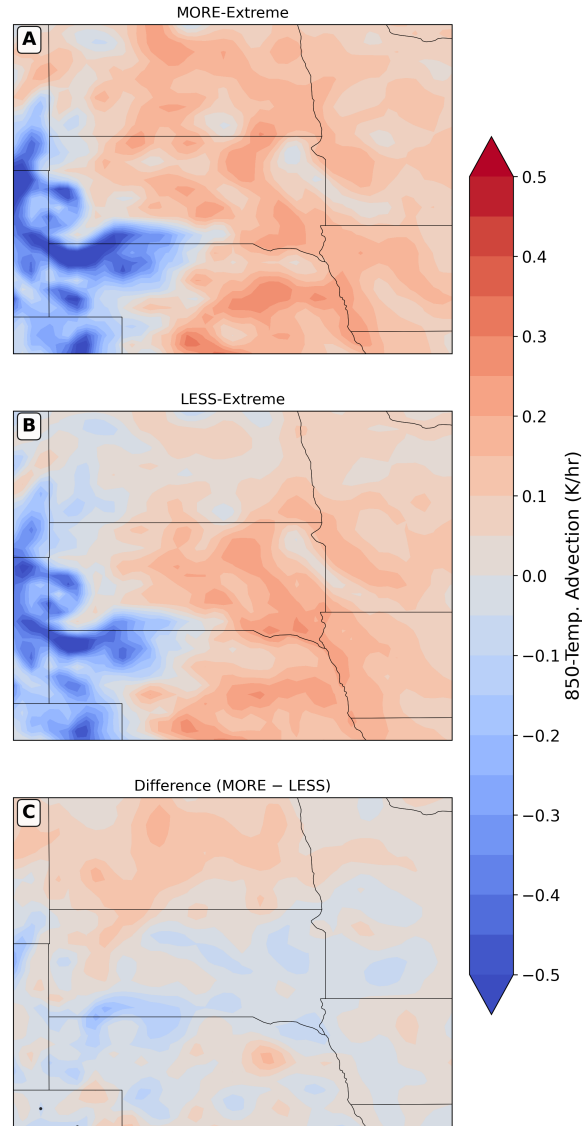


Figure 3.3: Composite 850-mb temperature advection (shaded; K h^{-1}) for (a) more-extreme and (b) less-extreme nocturnal extreme rainfall events. (c) Composite difference (more-extreme minus less-extreme), with stippling indicating statistically significant differences at the 95% confidence level.

3.1.4 Total Column Water Vapor

Both the more-extreme and less-extreme events exhibit very similar TCWV patterns across the eastern and central regions of the CONUS (Fig. 3.4a and 3.4b). For both extremities, an area of broad enhanced TCWV values extends northward from the Gulf

of Mexico into central and eastern CONUS, with values exceeding 25 kg m^{-2} inside the event domain. Since abundant atmospheric moisture is a key ingredient for heavy rainfall and EREs, this finding is expected and highlights the importance of moisture transport from the Gulf supporting these events. This result is also consistent with Chiappa et al. (2024), who identified statistically significant higher values of TCWV over the central and southern regions of the CONUS during active ERE summers.

Differences in TCWV between more-extreme and less-extreme events are generally small over most of the Great Plains Region (Fig.3.4c). Although, positive TCWV composite difference values can be seen over the northern portion of the event domain, indicating higher values of TCWV values for more-extreme events in the area. This pattern may imply a more efficient northward transport of moisture into the event domain for more-extreme events, likely associated with an enhanced 850-mb NLLJ supporting the transport of moisture up from the Gulf. This enhanced area of TCWV would provide more vertical moisture content for more-extreme storms occurring in the northern region of the event domain. This area of positive TCWV values is also statistically significant which shows that these findings are unlikely a result of random variability.

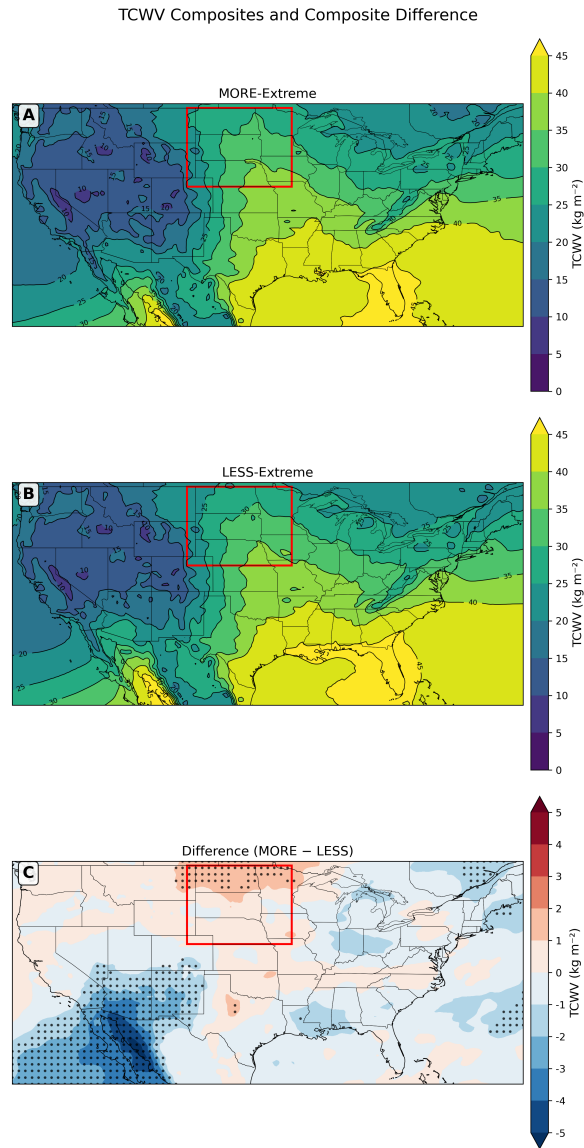


Figure 3.4: Composite total column water vapor (shaded and contoured; kg m^{-2}) for (a) more-extreme and (b) less-extreme nocturnal extreme rainfall events. (c) Composite difference (more-extreme minus less-extreme), with stippling indicating statistically significant differences at the 95% confidence level. The red box denotes the event domain

3.1.5 Vertically Integrated Water Vapor Flux and Vertically Water Vapor Flux Convergence

To synthesize the results presented in the earlier sections, VIWVF composites were constructed for both more-extreme and less-extreme events (Fig. 3.5a and 3.5b). A coherent spatial pattern was identified, characterized by enhanced VIWVF moving northward/northeastward from the western Gulf Coast, consistent with previous results from studies of heavy rainfall and flooding events across the central CONUS (e.g., Chiappa et al., 2024; Mo et al., 1997). These enhanced VIWVF values are directed into the event domain, where it then decelerates. The largest VIWVF magnitudes are located over the central Great Plains with stronger VIWVF occurring during more extreme events. These results are further supported by the positive values of VIWVF composite differences extending northward through the Great Plains region up from the Gulf of Mexico and across most of the event domain (Fig. 3.5c). A broad portion of this area is also statistically significant, indicating a robust result that more-extreme events have stronger values of VIWVF across the Great Plains region. This area of statistical significance also covers most of the western portion of the event domain and aligns with areas of statistically significant stronger 850-mb winds for more extreme events.

VIWVF Composites and Composite Difference

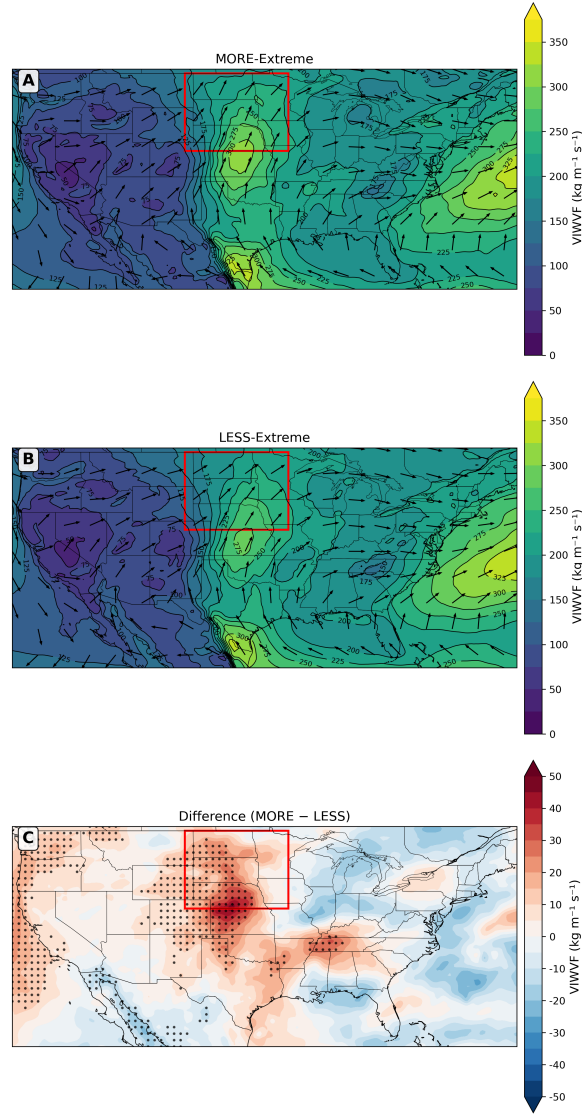


Figure 3.5: Composite vertically integrated water vapor flux (VIWVF; shaded and contoured; $\text{kg m}^{-1} \text{s}^{-1}$) and flux vectors for (a) more-extreme and (b) less-extreme nocturnal extreme rainfall events. (c) Composite difference (more-extreme minus less-extreme) with stippled regions indicating statistically significant differences at the 95% confidence level. The red box denotes the event domain.

Composites for VIWVF convergence were also analyzed to assess whether the deceleration of VIWVF within the event domain is associated with enhanced VIWVF convergence. When comparing the more-extreme events and the less-extreme events,

there are clear signals of localized stronger VIWVF convergence occurring across the event domain during more-extreme events (Fig. 3.6a and 3.6b). In addition, these enhanced localized convergence values are spread out over most of the domain, while the convergence values for the less-extreme events are more concentrated in the eastern portion of the event domain and are less localized. The VIWVF convergence composite difference plot further highlights this pattern. There are statistically significant positive anomalies in the western part of the event domain, suggesting that this signal is robust (3.6c). This composite difference pattern indicates higher values of VIWVF convergence for more-extreme events in the western portion of the domain.

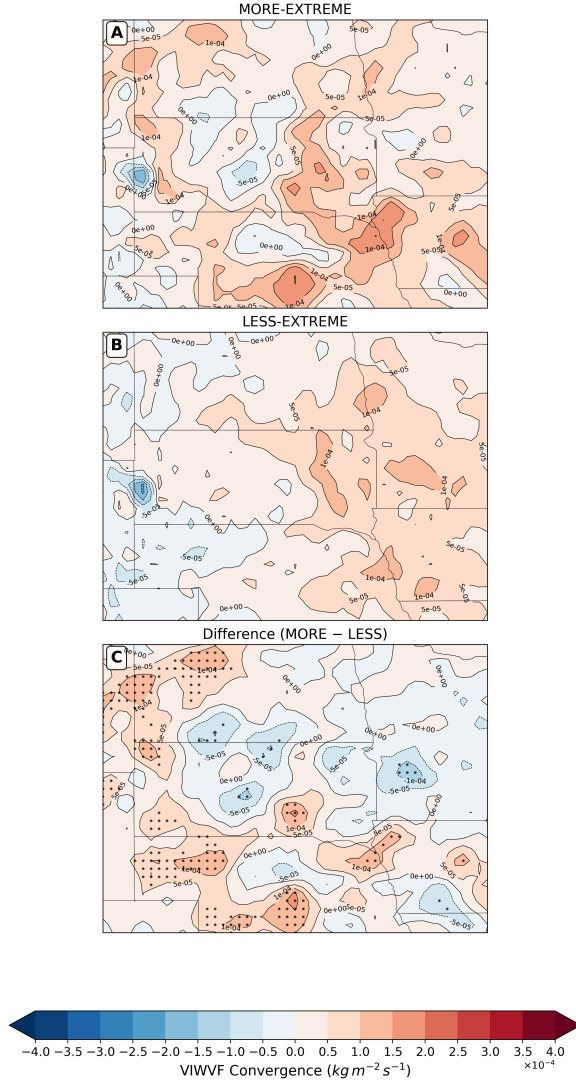


Figure 3.6: Composite vertically integrated water vapor flux convergence (VIWVF convergence; shaded and contoured; $\text{kg m}^{-2} \text{s}^{-1}$) for (a) more-extreme and (b) less-extreme nocturnal extreme rainfall events. (c) Composite difference (more-extreme minus less-extreme) with stippled regions indicating statistically significant differences at the 95% confidence level.

It is important to note that a single composite of all events will only capture one mean representative pattern. While useful for identifying the average large-scale atmospheric features, composites can smooth over important pattern variability associated with the nocturnal summertime EREs. As a result, multiple distinct synoptic pat-

terns leading to the EREs being analyzed may be missed. To address this limitation, the following sections will examine some of the synoptic patterns identified using the trained SOM and these patterns will be compared to these initial bulk composites and composite differences, thereby further addressing the research questions posed in this study.

3.2 General Overview of the Trained 3x3 SOM

Figure 3.7 shows the final 3x3 SOM trained on 500-mb geopotential height anomalies. The SOM groups events with similar 500-mb geopotential height anomaly patterns into nine representative nodes based on spatial structure of the anomaly fields. Therefore, each node represents a distinct representative 500-mb geopotential height anomaly pattern or regime of all events inside that node. This SOM also exhibits meteorologically coherent transitions between neighboring nodes, indicating that the training successfully preserved topological similarity. The nodes that are farthest away from each other on the SOM indicate regimes with very different patterns from each other. For example, node (0,0) indicates a regime with enhanced negative 500-mb geopotential height anomalies on both the west and east coasts of the continental United States while node (2,2) shows broad, weaker positive 500-mb geopotential height anomalies across the entirety of the continental United States (CONUS) (Fig. 3.7a and 3.7i).

500-mb Height Anomaly SOM

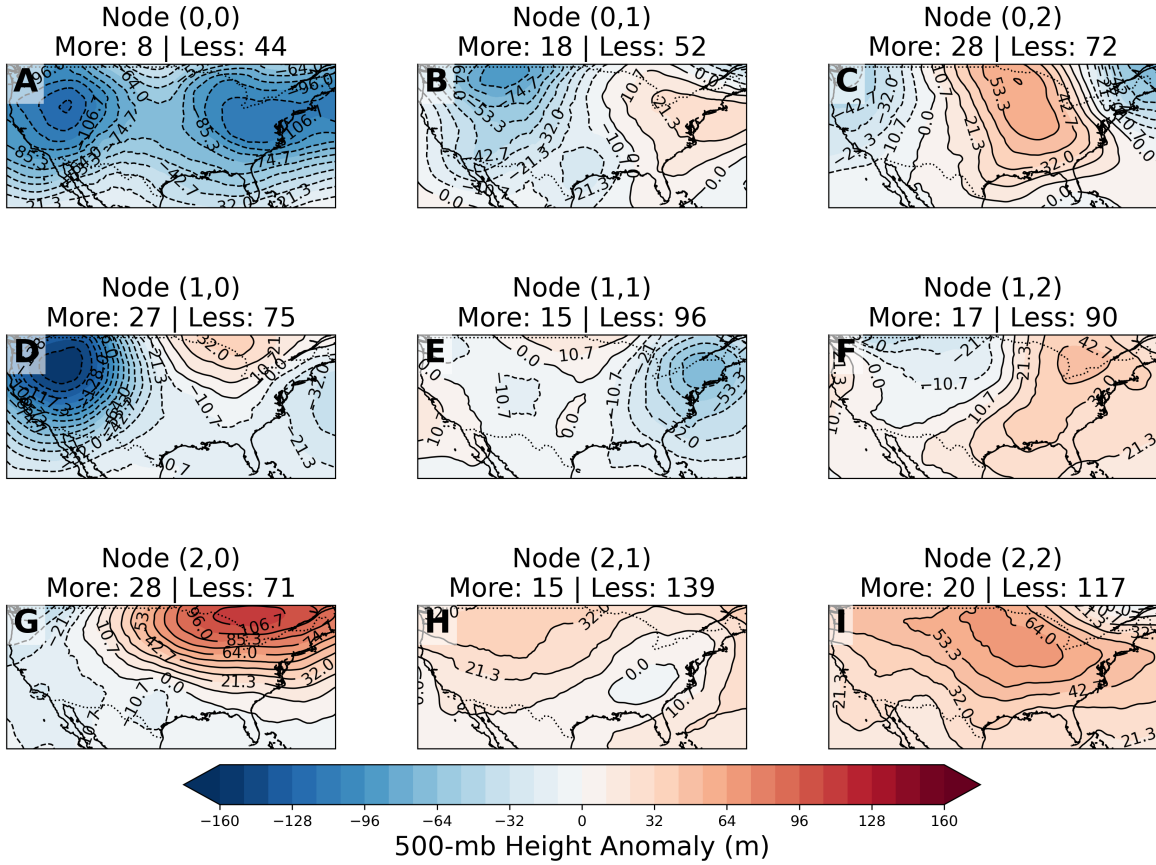


Figure 3.7: SOM of 500-mb geopotential height anomalies (shaded and contours; m). Each node represents a distinct synoptic-scale circulation regime. Each node includes the number of more-extreme (More) and less-extreme (Less) events classified within each SOM pattern,

Figure 3.7 also presents the distribution of more-extreme and less-extreme events across SOM nodes. Node frequencies provide a measure of how often each synoptic pattern occurs within the dataset, allowing the most representative regimes to be distinguished from less frequently occurring patterns. Table 3.1 shows the node frequencies ranked from highest to lowest. It is calculated that node (2,1) has the highest frequency of events and node (0,0) has the lowest frequency of events.

Ranked Overall Node Frequencies

Rank	Node	Overall Frequency
1	(2, 1)	0.161
2	(0, 0)	0.145
3	(1, 1)	0.136
4	(2, 2)	0.115
5	(0, 1)	0.106
6	(1, 2)	0.091
7	(2, 0)	0.085
8	(0, 2)	0.084
9	(1, 0)	0.077

Table 3.1: Ranked overall node frequencies for the SOM, showing the relative occurrence of each node across the full event record.

The event count also is useful in determining which nodes are more-extreme event dominant or less-extreme event dominant. However, due to the total number of less-extreme events being much lower than the total number of more-extreme events, the counts alone cannot properly determine the dominance of each node. To account for the unequal sample sizes, these frequencies were normalized within each event category by dividing the number of events in each SOM node by the total number of events in that category. This produces class conditional relative frequencies that represent the proportion of more-extreme and less-extreme events assigned to each node. Table 3.2 shows node-wise frequencies for more-extreme and less-extreme events. This approach allows for differences in node occupancy to reflect relationships between SOM node and event types rather than differences in sample size. Nodes (0,0), (1,1), (1,2), (2,1),

(2,2) all show higher frequencies for less-extreme events while nodes (0,1), (0,2), (1,0), and (2,0) all show higher frequencies for more-extreme events. This process allows for the identification of which synoptic patterns are associated more strongly with each event type.

More vs Less Extreme Event Dominance by SOM Node

MORE > LESS LESS > MORE				
Node	More Count	More Freq.	Less Count	Less Freq.
(0,0)	8	0.045	44	0.058
(0,1)	18	0.102	52	0.069
(0,2)	28	0.159	72	0.095
(1,0)	27	0.153	74	0.098
(1,1)	15	0.085	96	0.127
(1,2)	17	0.097	90	0.119
(2,0)	28	0.159	71	0.094
(2,1)	15	0.085	139	0.184
(2,2)	20	0.114	117	0.155

Table 3.2: More-extreme versus less-extreme event frequencies by SOM node. Yellow cells indicate nodes where more-extreme events occur proportionally more often than less-extreme events, while blue cells indicate the opposite.

Figure 3.8 shows the seasonality of events by SOM node and separated by extreme severity. A clear pattern of seasonality appears across the SOM lattice. Nodes (0,0) and (1,0) show a clear peak in June for both extreme severity. Node (2,0) shows a peak in July for less-extreme events and a slight peak in July for more-extreme events but with little variability between the three months. Nodes (2,1) and (2,2) show large peaks in both July and August for less-extreme events with July having the higher peak. These large peaks are related to these nodes having the largest total number of less-extreme events. Less-extreme events stay under a total of 10 in June. For more extreme events, the peak is also in July and August with totals showing little variability between the two months. Node (2,1) shows no more-extreme events in June while node (2,2) shows little. Node (0,2) and node (1,2) show less-extreme event totals gradually increasing

from June to August with the peak in August. For node (1,2) there is little variability in more-extreme event totals if any between the three months. For node (0,2) there is a peak in total more-extreme events in June, however there little variability between the three months. Nodes (0,1) and (1,1) show a decrease in less-extreme events totals from June to July but then an increase again going into August. For node (0,1) there is a gradual decrease in more-extreme event totals from June to August with a clear peak in June. For node (1,1) this is not the case as there is little variability if any between the three months.

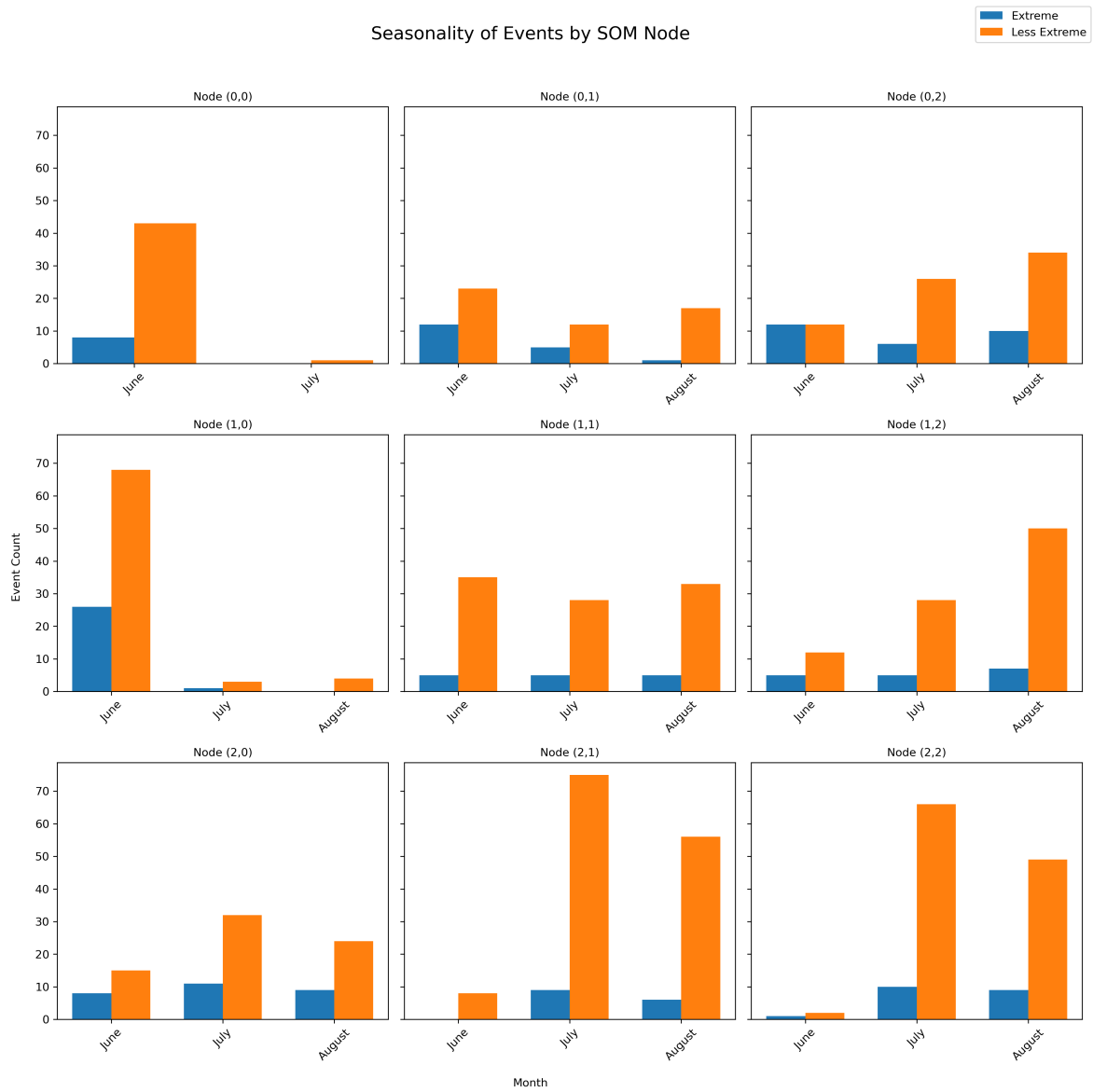


Figure 3.8: Seasonal distribution of more-extreme and less-extreme nocturnal rainfall events for each SOM node. Bars show event counts for June, July, and August, with blue indicating more-extreme events and orange indicating less-extreme events.

3.3 Environmental Characteristics of Individual SOM Nodes

Although the SOM provides a broad classification of synoptic regimes, examining individual nodes allows for a more detailed understanding of the associated atmospheric structures. This next section further examines some of the node regimes identified by the SOM. These node regimes were chosen because the numbers of more-extreme and less-extreme events had the most comparable distributions, making them the most ideal for comparison when identifying differences seen in the later section 3.4. This subsection evaluates node (1,0) and its associated atmospheric patterns.

3.3.1 Node (1,0)

At 500-mb, strong negative height anomalies are centered over the western continental United States, indicating an amplified trough, while weaker positive height anomalies are present over portions of central Canada (Fig. 3.9a). This pattern is consistent with the 500-mb geopotential height field, which shows an amplified trough over the western United States and a less amplified downstream ridge over eastern portions of the country (Fig. 3.9b). Together, these features produce southwesterly flow that extends into the event domain and are consistent with the bulk composites shown in Section 3.1.1. In addition, there is an area of enhanced mid-level winds extending from the western United States into the central United States, which suggests that the strongest mid-level flow is downstream of the western trough (3.9c). This trough-ridge pattern is consistent with the patterns associated with extreme rainfall events described by Chiappa et al. (2024)

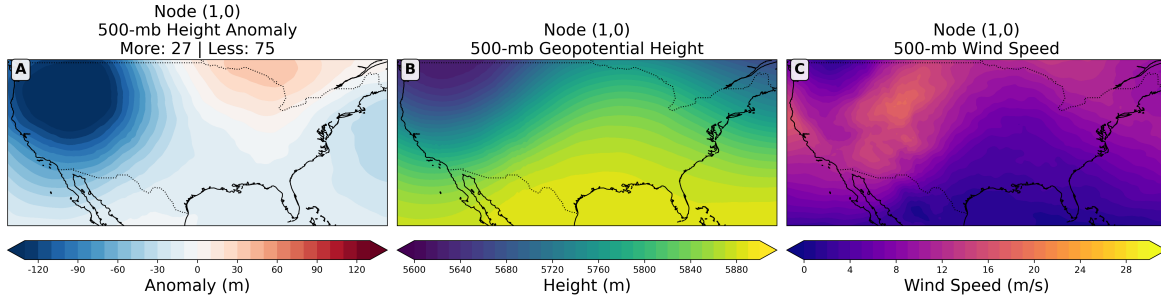


Figure 3.9: Composite mid-tropospheric environment for Node (1,0) showing (a) 500-mb height anomalies (m), (b) 500-mb geopotential height (m), and (c) 500-mb wind speed (m s^{-1}) for all events within node.

At 850-mb, the strongest wind speeds are located over most of the United States Great Plains with predominantly southerly flow originating from the Gulf of Mexico supporting the transportation of warm, moist air into this region (Fig. 3.10a). The event domain is positioned near the nose of the 850-mb jet streak, where wind maxima are associated with enhanced ascent. This finding is consistent with the bulk 850-mb wind composites described in section 3.1.2 and the findings by Zhong et al. (1996). This suggests that the events in the event domain may be strongly influenced by jet-driven moisture transport and dynamically forced ascent associated with the LLJ, creating a favorable environment for the development and maintenance of extreme rainfall. Consistent with this enhanced southerly flow, positive values of temperature advection dominate the event domain, indicating widespread warm air advection at 850-mb (Fig. 3.10b). This widespread warm air advection may provide dynamical forcing for the EREs within the domain through enhanced upward vertical motion. This pattern is also consistent with the pattern seen in the 850-mb temperature advection bulk composite in Section 3.1.3 and further supports the role of the LLJ in transporting warm, moist air from the Gulf of Mexico into the Great Plains region. The agreement between the SOM node composites and the bulk composites also demonstrates that the SOM successfully captures the key synoptic environments associated with nocturnal

extreme rainfall.

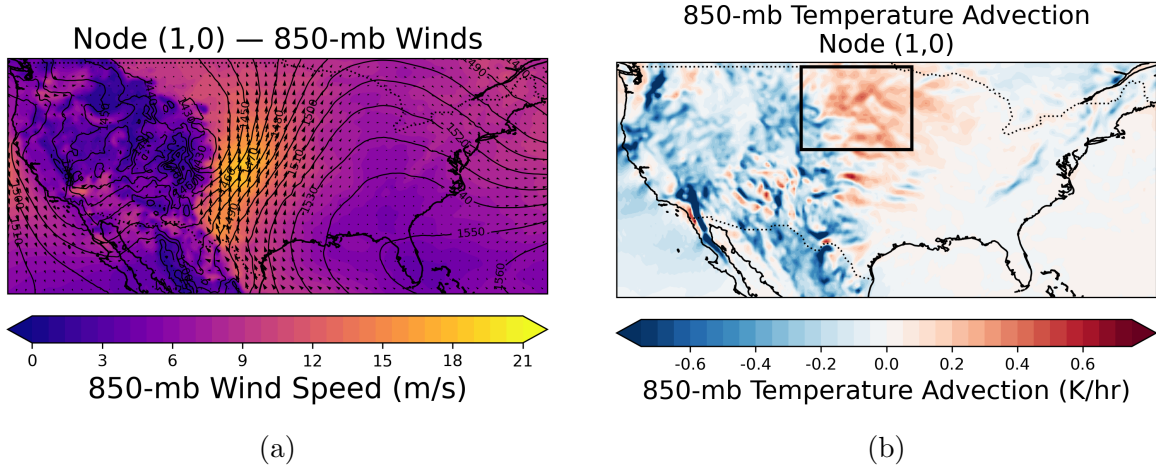


Figure 3.10: (a) Composite 850-mb wind field for Node (1,0), showing geopotential height (m; contours), wind speed (m s^{-1} ; shaded), and wind direction (barbs). Shading indicates the magnitude of the 850-mb wind speed, while contours depict the 850-mb geopotential height field. (b) Composite 850-mb temperature advection for Node (1,0). Shading shows temperature advection (K h^{-1}), with positive values indicating warm-air advection and negative values indicating cold-air advection. The boxed region denotes the event domain.

Consistent with the enhanced low-level flow, a broad region of enhanced TCWV values extends from the Gulf of Mexico northward across much of the United States Great Plains (Fig. 3.11). This TCWV pattern resembles the patterns shown in the bulk composites in section 3.1.4. The strongest values of TCWV in the event domain occur in the eastern portion of the event domain. This moisture distribution is further supported by the VIWVF vapor transport field (Fig. 3.12a), which further illustrates the importance of moisture transport in this regime. A corridor of enhanced VIWVF extends northward/northeastward from the west coast of Gulf of Mexico, closely aligning with the regions of strong 850-mb winds and elevated TCWV values discussed previously. This pattern also aligns with what was seen in the bulk VIWVF composites in section 3.1.5. Also like the bulk composites in section 3.1.5, an enhanced area of VIWVF weakens in the event domain, suggesting a region of moisture convergence

where the northward transport decelerates. This interpretation is confirmed by the positive values of VIWVF convergence dominating the event domain (Fig. 3.10b), suggesting enhanced upward vertical motion which may act as an additional dynamical forcing mechanism for ascent alongside the widespread 850-mb warm air advection.

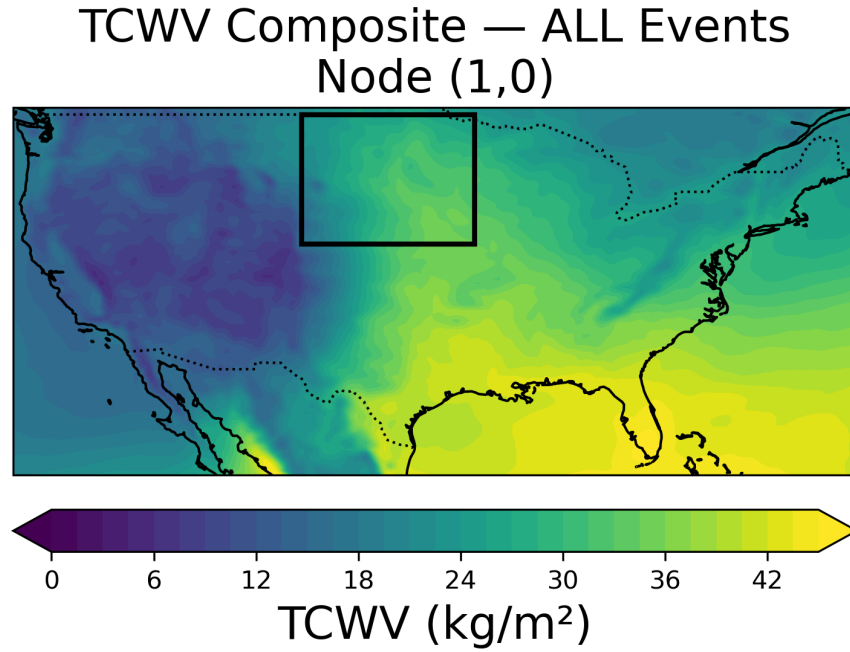


Figure 3.11: Composite total column water vapor (TCWV) for Node (1,0). Shading indicates TCWV (kg m^{-2}). The boxed region denotes the event domain.

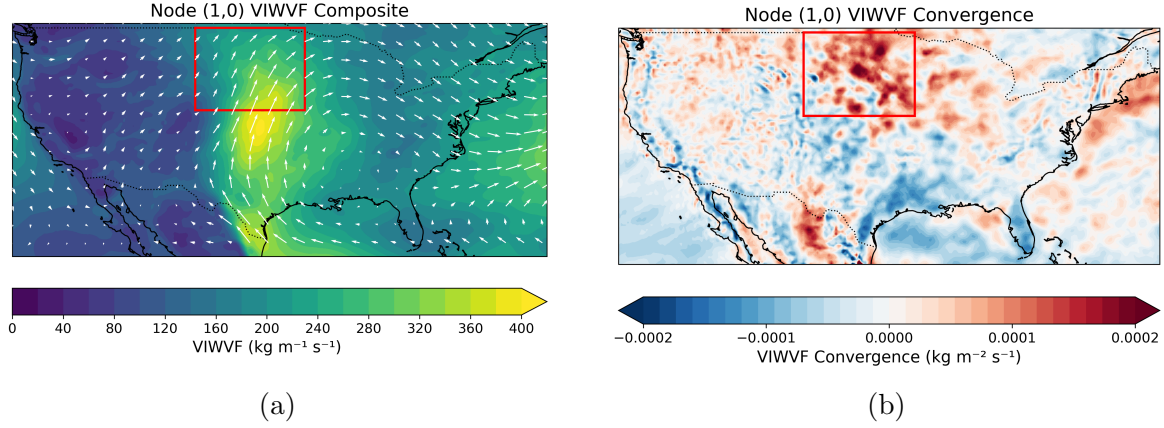


Figure 3.12: (a) Composite vertically integrated water-vapor flux (VIWVF) for Node (1,0). Shading shows VIWVF magnitude ($\text{kg m}^{-1} \text{s}^{-1}$), and arrows indicate the direction and relative magnitude of the moisture-flux vectors. (b) Composite vertically integrated water-vapor flux (VIWVF) convergence for Node (1,0). Shading shows VIWVF convergence in $\text{kg m}^{-2} \text{s}^{-1}$, with positive values indicating moisture-flux convergence and negative values indicating divergence. The boxed region denotes the event domain.

3.3.2 Node (2,0)

Having characterized the atmospheric environment associated with node (1,0), the analysis next examines node (2,0) to similarly identify distinct patterns within this synoptic regime. The patterns in node (2,0) are also compared to the patterns in node (1,0) to identify key differences and similarities between the synoptic regimes.

At 500-mb, there is a large area of strong positive 500-mb geopotential height anomalies centered over parts of Midwest, Northeast United States, and southeastern Canada (Fig. 3.13a). There are also weaker negative height anomalies centered over northwestern United States and southwestern Canada. In response, a broad 500-mb ridge extends over most of the continental United States and a trough off the West Coast (Fig. 3.13b). Due to this setup, there is southwesterly flow over the event

domain. The strongest winds at 500-mb occur downstream of the mid-level trough like what was seen in the previous node regime (Fig. 3.13c).

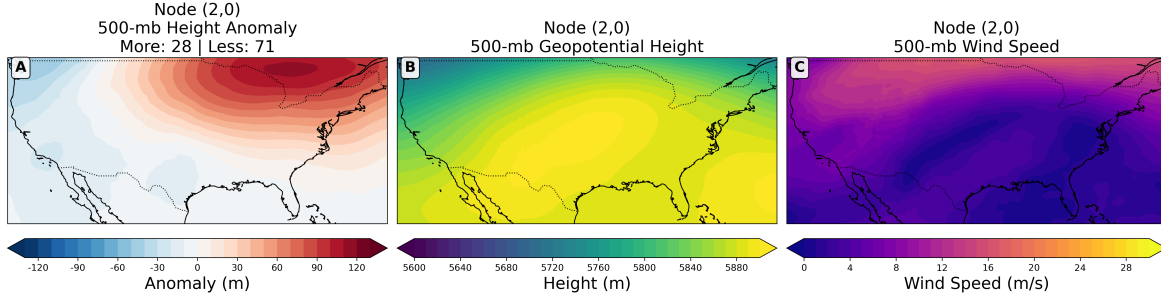


Figure 3.13: Composite mid-tropospheric environment for Node (2,0) showing (a) 500-mb height anomalies (m), (b) 500-mb geopotential height (m), and (c) 500-mb wind speed (m s^{-1}) for all events within node.

Investigating the LLJ at 850-mb, the strongest winds are southerly and extend across the U.S. Great Plains, transporting warm, moist air up from the Gulf of Mexico (Fig. 3.14a). Additionally, the event domain is positioned near the nose of the LLJ, where the wind maxima occur, and associated enhanced 850-mb ascent is favored. The LLJ also decelerates and terminates in the event domain. Although the overall spatial setup in node (2,0) resembles that of node (1,0), the LLJ associated with node (1,0) exhibits stronger winds. This pattern indicates that, while the large-scale environments of these nodes share similar characteristics, differences among node regimes arise from variations in the magnitude and placement of smaller-scale atmospheric features within those broader patterns. Like node (1,0), 850-mb warm air advection extends over most of the event domain in node (2,0), which is consistent with the southerly flow of the LLJ transporting warm air into the domain (Fig. 3.14b). However, when compared with the setup in node (1,0), the magnitude of warm-air advection is weaker, consistent with the weaker 850-mb wind speeds shown in Figure 3.14b. This supports the idea that the LLJ plays a significant role in the presence of warm air advection within the event domain. The presence of warm air advection suggests synoptic-scale ascent, which may

support the development and maintenance of EREs within in event domain.

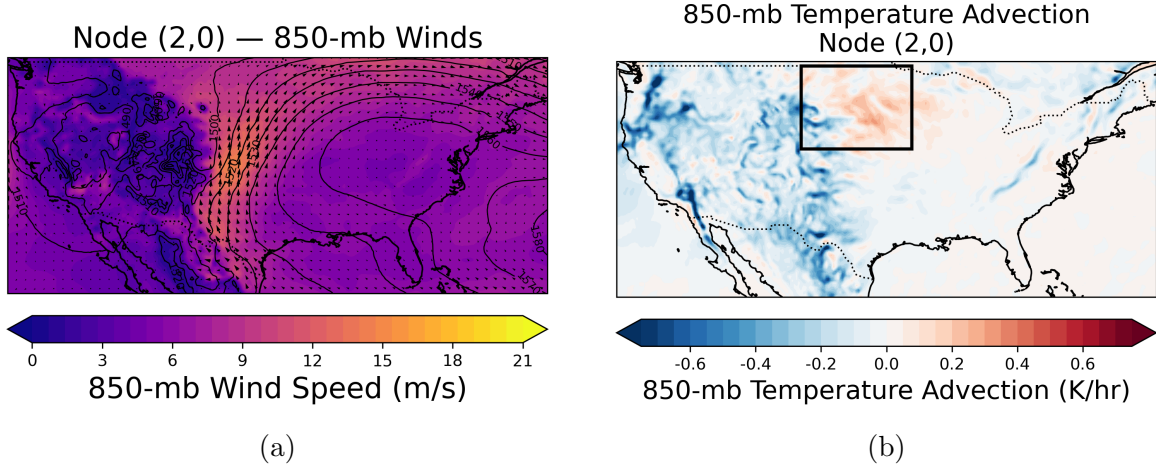


Figure 3.14: (a) Composite 850-mb wind field for Node (2,0), showing geopotential height (m; contours), wind speed (m s^{-1} ; shaded), and wind direction (barbs). Shading indicates the magnitude of the 850-mb wind speed, while contours depict the 850-mb geopotential height field. (b) Composite 850-mb temperature advection for Node (2,0). Shading shows temperature advection (K h^{-1}), with positive values indicating warm-air advection and negative values indicating cold-air advection. The boxed region denotes the event domain.

Like node (1,0), a broad region of enhanced TCWV values extends the Gulf of Mexico northward across much of the United States Great Plains (Fig. 3.15). Again, the strongest values of TCWV inside the event domain occur in the eastern portion of the domain. The enhanced TCWV values are also consistent with the role of the NLLJ in transporting moisture from the Gulf of Mexico into the Great Plains. The distribution of enhanced values of TCWV across the Great Plains is further illustrated by the pattern of VIWVF. Like the previous node, enhanced values of VIWVF extend up off the west coast of the Gulf northward/northeastward into the Great Plains region (Fig. 3.16a). There is a VIWVF maximum over the central Great Plains with the event domain located over the northern half of this maximum. The VIWVF convergence pattern for node (2,0) shows areas of localized VIWVF convergence within the event domain again like what was seen in the initial bulk VIWVF convergence composites

in section 3.1.5 (Fig. 3.16b). This pattern suggests localized areas of upward vertical motion and ascent which may aid in the forcing for ascent for EREs in the northern Great Plains.

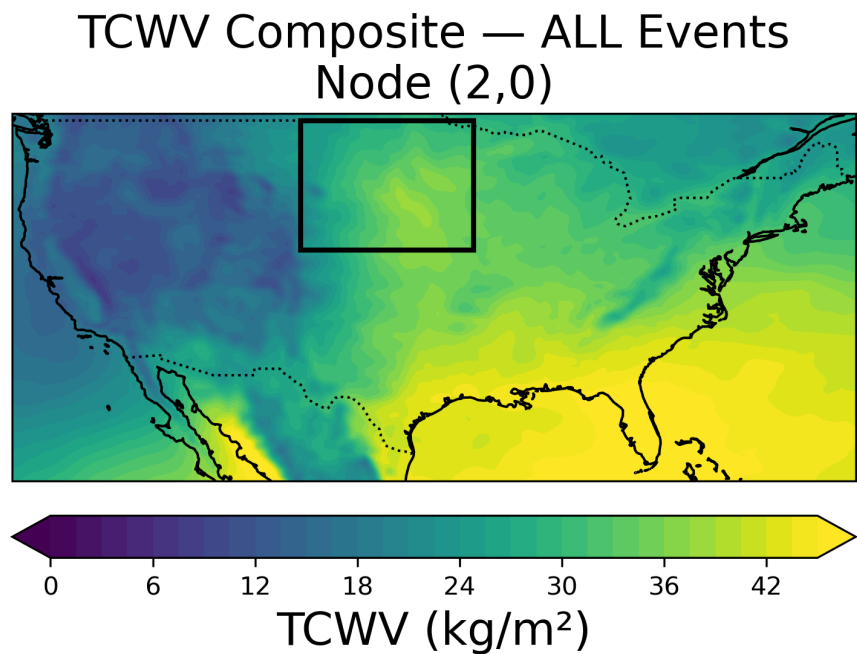


Figure 3.15: Composite total column water vapor (TCWV) for Node (2,0). Shading indicates TCWV (kg m^{-2}). The boxed region denotes the event domain.

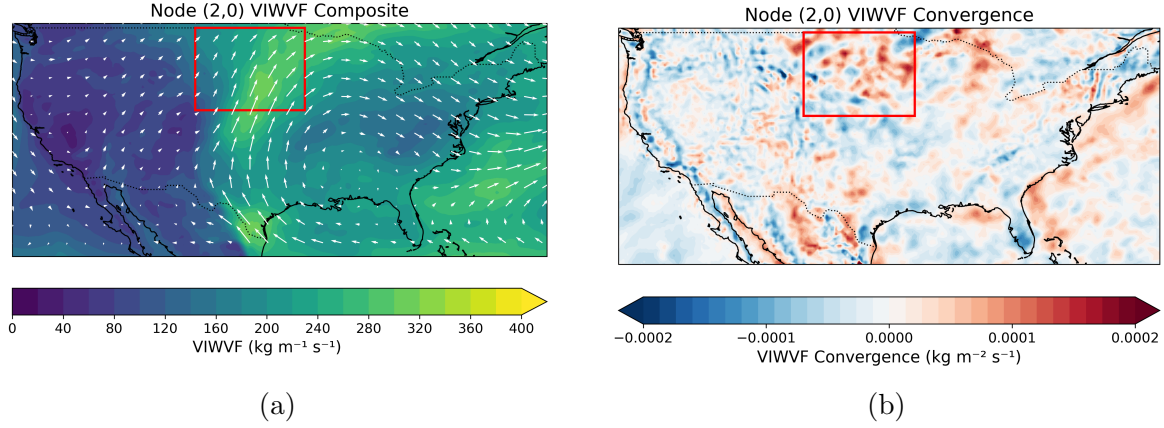


Figure 3.16: (a) Composite vertically integrated water-vapor flux (VIWVF) for Node (2,0). Shading shows VIWVF magnitude ($\text{kg m}^{-1} \text{s}^{-1}$), and arrows indicate the direction and relative magnitude of the moisture-flux vectors. (b) Composite vertically integrated water-vapor flux (VIWVF) convergence for Node (2,0). Shading shows VIWVF convergence in $\text{kg m}^{-2} \text{s}^{-1}$, with positive values indicating moisture-flux convergence and negative values indicating divergence. The boxed region denotes the event domain.

3.3.3 Node (0,2)

Having characterized the atmospheric environment associated with node (1,0) and (2,0), the analysis next examines node (0,2) to similarly identify distinct patterns within this synoptic regime. This node is also compared the patterns to the previous nodes.

At 500-mb, an area of positive geopotential height anomalies extends over most of the central and eastern portions of the CONUS in node (0,2) (Fig. 3.17a). An area of negative 500-mb geopotential height anomalies is located over the West Coast and the Northeast CONUS. Consistent with this setup, the 500-mb geopotential height plot shows a ridge centered over the eastern edge of the event domain and troughs over the west coast and the northeastern United States (Fig. 3.17b). The event domain

is located just upstream of the mid-level ridge axis and downstream of the mid-level trough axis over the west coast. Like nodes (1,0) and (2,0), this positions the event domain in an area of southwesterly mid-level flow. Enhanced areas of 500-mb wind speeds are located downstream of the western trough and upstream of the eastern trough (Fig. 3.17c).

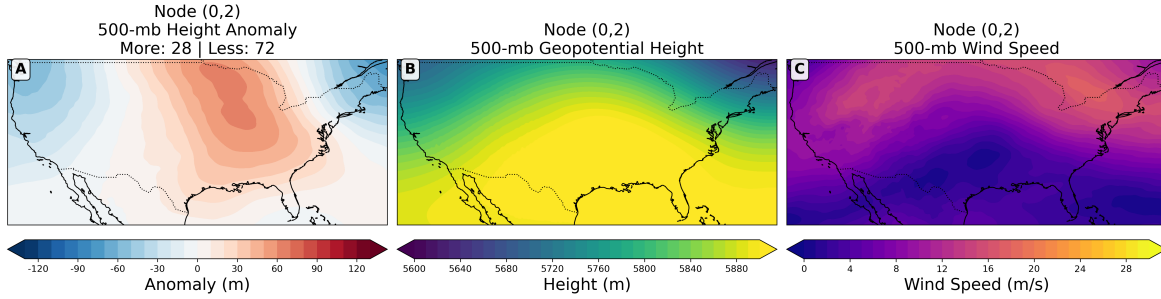


Figure 3.17: Composite mid-tropospheric environment for Node (0,2) showing (a) 500-mb height anomalies (m), (b) 500-mb geopotential height (m), and (c) 500-mb wind speed (m s^{-1}) for all events within node.

The 850-mb wind pattern for node (0,2) stays consistent with the patterns found in the previous nodes (Fig. 3.18a). A broad area of enhanced 850-mb winds extends up from the west coast of the Gulf of Mexico and moves northward through most of the Great Plains suggesting the transport of warm, moist air up from the Gulf. The event domain is located just to the north of the strongest winds, and the NLLJ deaccelerates inside the event domain possibly implying the accumulation of moisture. The 850-mb temperature advection pattern for node (0,2) is also like the patterns described for nodes (1,0) and (2,0) (Fig. 3.18b). Positive values of 850-mb temperature advection extend over most of the event domain. This pattern indicates widespread warm air advection inside the event domain suggesting possible widespread synoptic ascent aiding in storm initiation and maintenance.

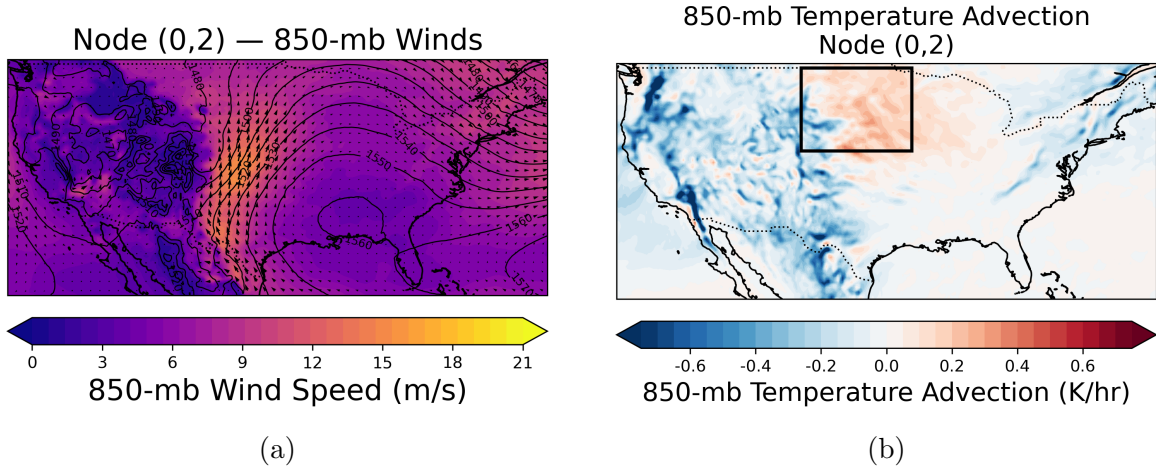


Figure 3.18: (a) Composite 850-mb wind field for Node (0,2), showing geopotential height (m; contours), wind speed (m s^{-1} ; shaded), and wind direction (barbs). Shading indicates the magnitude of the 850-mb wind speed, while contours depict the 850-mb geopotential height field. (b) Composite 850-mb temperature advection for Node (0,2). Shading shows temperature advection (K h^{-1}), with positive values indicating warm-air advection and negative values indicating cold-air advection. The boxed region denotes the event domain.

Resembling the previous nodes, node (0,2) shows an area of enhanced TCWV extending northward from the Gulf of Mexico into most of the Great Plains Region (Fig. 3.19). The eastern portion of the event domain also has the highest values of TCWV in comparison to the rest of the domain. This pattern is consistent with the transportation of moisture by the LLJ. This is further supported by the VIWVF pattern, as an area of enhanced VIWVF extends up from the west coast of the Gulf of Mexico northward into the Great Plains region (Fig. 3.20a). The area of the strongest VIWVF values over the Great Plains region closely aligns with the nose of the NLLJ shown in the 850-mb wind composite suggesting that enhanced VIWVF is closely connected to the NLLJ. The enhanced area of VIWVF decelerates within the event domain, possibly leading to moisture accumulation and possible VIWVF convergence. This interpretation is further supported by the VIWVF convergence pattern of localized enhanced areas of VIWVF convergence across the event domain, especially over

the eastern portion of the domain (Fig. 3.20b). This VIWVF and VIWVF convergence pattern is consistent spatially with patterns seen in nodes (1,0) and (2,0). This setup indicates event wide localized areas of ascent which implies another form of forcing for the events within node regime.

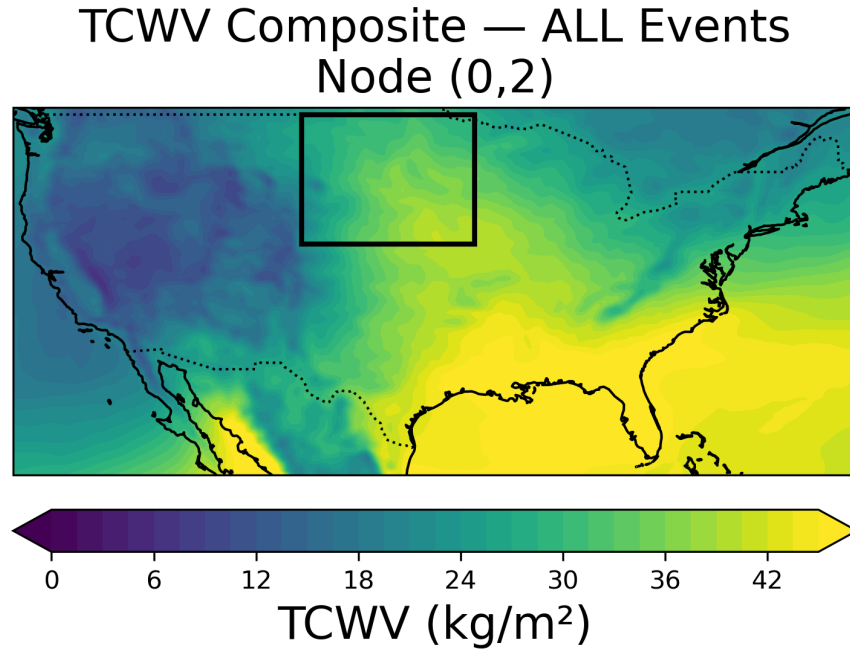


Figure 3.19: Composite total column water vapor (TCWV) for Node (0,2). Shading indicates TCWV (kg m^{-2}). The boxed region denotes the event domain.

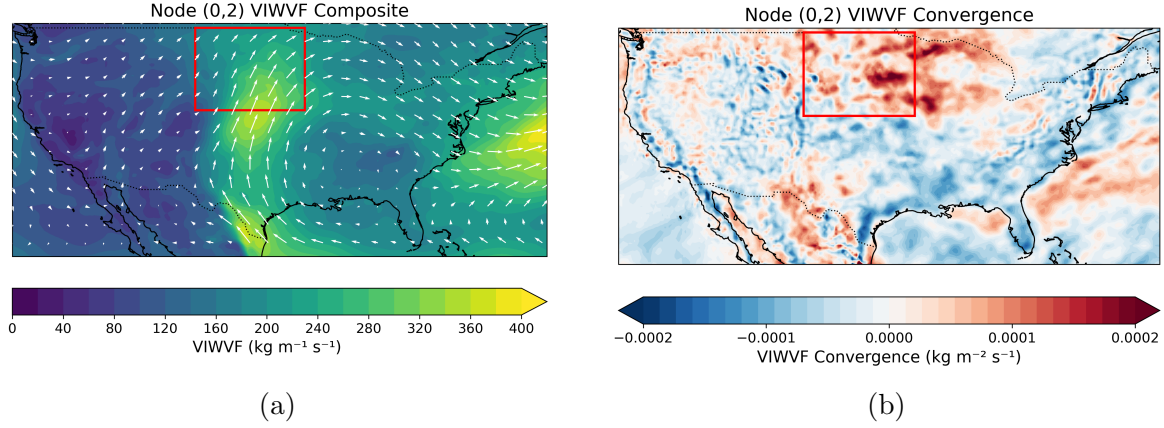


Figure 3.20: (a) Composite vertically integrated water-vapor flux (VIWVF) for Node (2,0). Shading shows VIWVF magnitude ($\text{kg m}^{-1} \text{s}^{-1}$), and arrows indicate the direction and relative magnitude of the moisture-flux vectors. (b) Composite vertically integrated water-vapor flux (VIWVF) convergence for Node (2,0). Shading shows VIWVF convergence in $\text{kg m}^{-2} \text{s}^{-1}$, with positive values indicating moisture-flux convergence and negative values indicating divergence. The boxed region denotes the event domain.

3.4 Differences Between More-Extreme and Less-Extreme Events

3.4.1 Node (1,0)

At lower levels of the atmosphere, a region of statistically significant positive 850-mb wind speed differences is evident over the southwestern portion of the event domain (Fig. 3.21). This indicates that more-extreme events are associated with statistically significant stronger 850-mb winds in this region compared to less-extreme events. There is also an area of negative composite differences that extends from the Gulf of Mexico to the eastern portion of the event domain and just east of the event domain. Even

though few areas east of the event domain exhibit statistical significance most likely due to uneven sample sizes, there is a pattern of stronger winds for less-extreme events across the Great Plains within this node regime.

Both more-extreme and less-extreme events exhibit similar large-scale 850-mb temperature advection patterns, with warm-air advection extending over most of the event domain and cold air advection located in portions of the southeast corner of the domain (Fig. 3.22a and 3.22b). However, more-extreme events experience localized areas of both enhanced warm air and cold air advection compared to less-extreme events. This is confirmed by the localized differences present within the event domain (Fig. 3.22c). Portions of these localized differences exhibit statistically significant differences between the two types of extremes. This may indicate stronger thermal gradients during more extreme events which in return may be associated with an enhanced frontal structure. Compared to the bulk composite difference pattern in section 3.1.3, the SOM-based composites more clearly isolate these statistically significant localized structures, which are otherwise muted when all events are bulked together. There is also a noticeable area of positive differences in the northwestern portion of the event domain. This is consistent with the stronger 850-mb wind speeds for more-extreme events within the western portion of the domain which suggests the LLJ playing a key role in the transportation of warmer air into the event domain (Fig. 3.22).

SOM Node (1,0) MORE – LESS 850-mb Wind Speed Difference

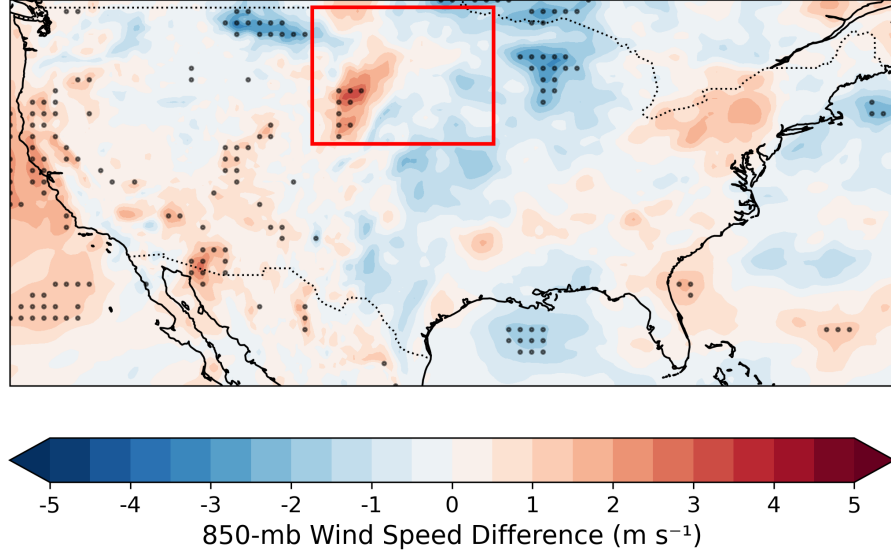


Figure 3.21: Composite difference (more – less) in 850-mb wind speed (m s^{-1}) for Node (1,0). Shading indicates the magnitude of the wind-speed difference. Stippling marks grid points where the composite difference is significant at the 95% significance level. The boxed region denotes the event domain.

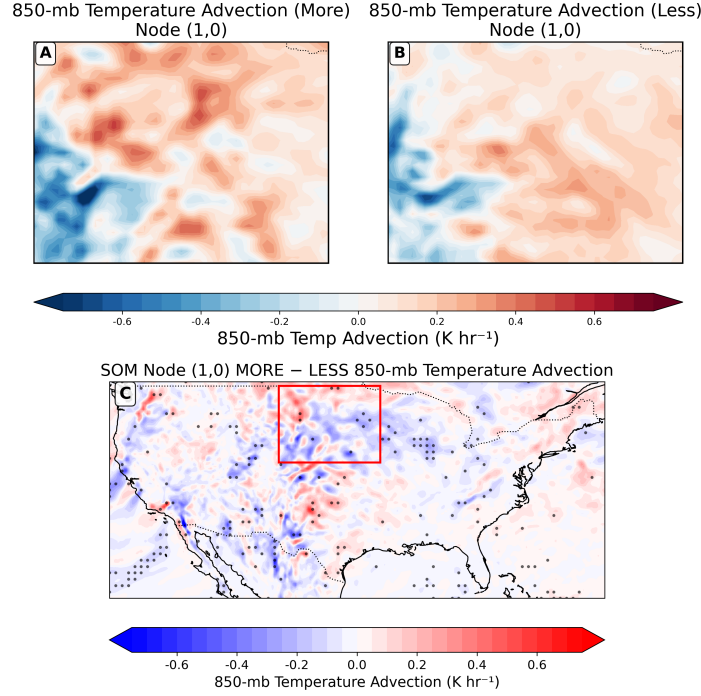


Figure 3.22: Composite 850-mb temperature-advection fields (K h^{-1}) for Node (1,0) for (a) more-extreme events, (b) less-extreme events and (c) the composite difference (more - less) in 850-mb temperature advection. Stippling marks grid points where the composite difference is statistically significant at the 95% significance level. The boxed region denotes the event domain.

There is a pronounced corridor of positive TCWV differences extending from the Gulf of Mexico into the western and northeast portions of the event domain (Fig. 3.23). These positive differences indicate higher values of TCWV for more extreme events. This pattern may suggest more enhanced moisture transport up from the Gulf of Mexico during more-extreme events. Portions of these positive differences are also statistically significant, indicating that this difference between more-extreme and less-extreme events is unlikely to be the result of random variability. The concentration of positive TCWV differences in the western and northeastern portions of the event domain is consistent with the spatial distribution of more-extreme rainfall events within this node, whose centroid is shifted farther northwest relative to the less-extreme events centroid (Fig. 3.24a and 3.24b). This distribution of more-extreme events also is consis-

tent with the enhanced 850-mb winds discussed previously (Fig. 3.21). The centroid for the less-extreme events stays closer to the center to the middle of the event domain, suggesting that the less-extreme events are more evenly distributed across the domain. This pattern differs from the initial composite difference plot in section 3.1.4, which shows a weak corridor of positive TCWV values across the a large portion of the Great Plains and statistically significant positive values of TCWV confined to the northern portion of the event domain. This finding continues to emphasize the importance of synoptic regime separation in isolating patterns that may be lost in a bulk-event composite.

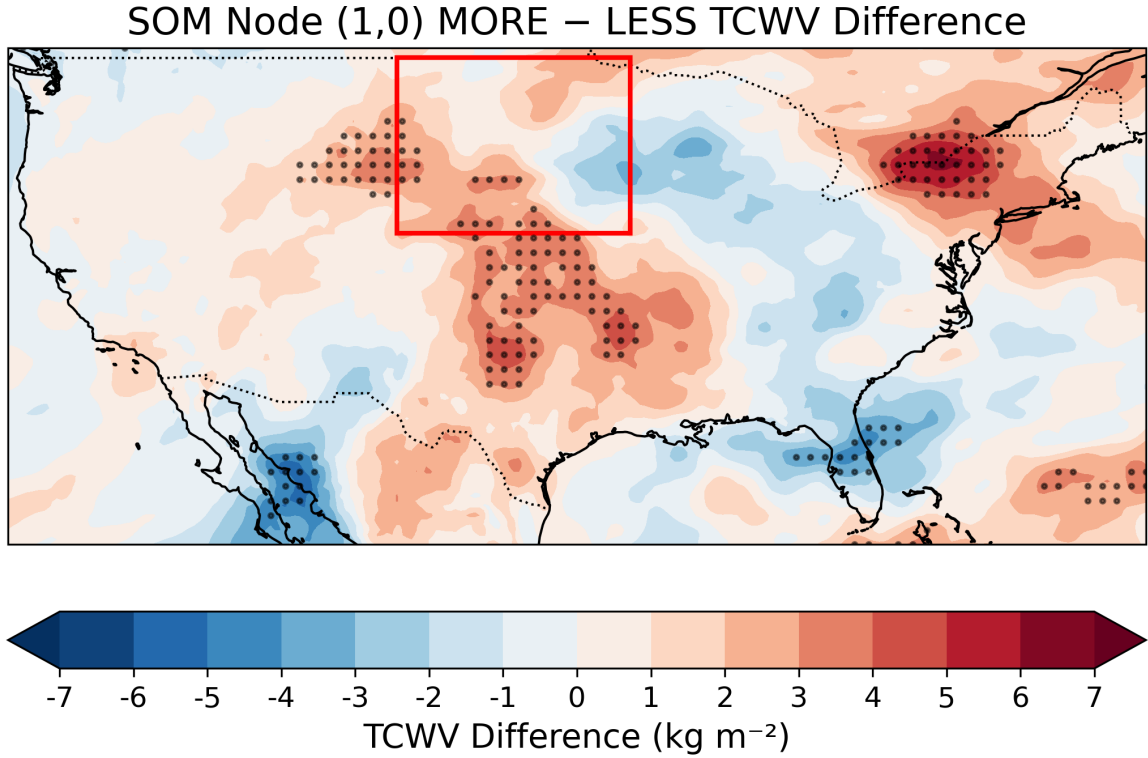


Figure 3.23: Composite difference (more - less) in total column water vapor (TCWV) for Node (1,0). Shading shows TCWV differences (kg m^{-2}). Stippling marks grid points where the composite difference is statistically significant at the 95% significance level. The boxed region denotes the event domain.

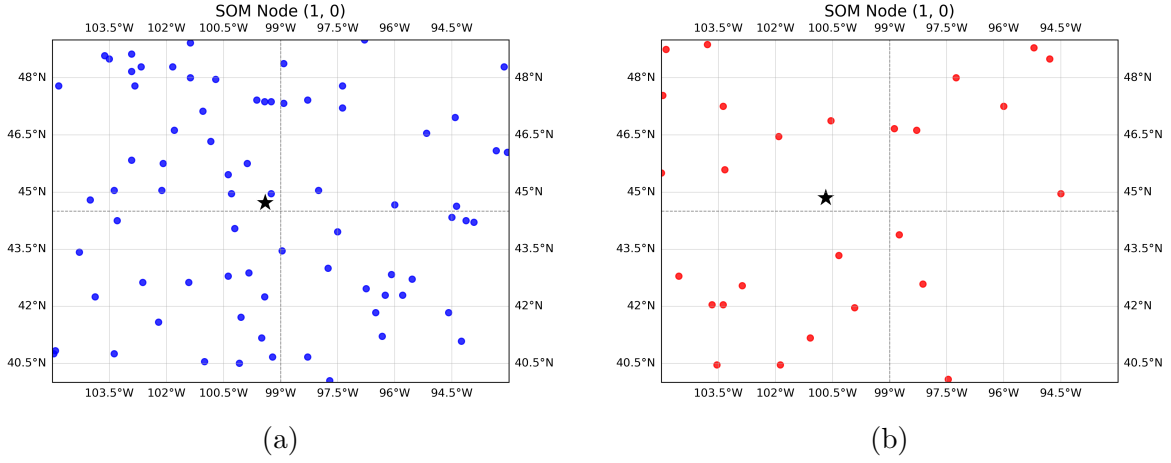


Figure 3.24: Event locations for SOM Node (1,0) shown for (a) less-extreme events and (b) more-extreme events. Each panel displays the spatial distribution of events within the analysis domain, with the black star marking the centroid of all events associated with the node. Dashed lines indicate the centroid latitude and longitude of the event domain.

A broad area of positive VIWVF differences extends across the Great Plains and into the western portion of the event domain (Fig. 3.25). Within this region, statistically significant differences are confined to the southwest portion of the event domain, where more-extreme events have significantly stronger VIWVF compared to less-extreme events. This aligns with the distribution of more-extreme events within the event domain as the centroid of more-extreme events is shifted west compared to the center of the domain (Fig. 3.24b). Although this is the only area of positive statistically significant differences near or within the event domain, there is a pattern of stronger VIWVF for more-extreme events extending over most of the Great Plains region. This pattern differs from the pattern shown in the initial bulk VIWVF composite difference plot in section 3.1.5, which showed a larger area of statistically significant positive differences extending over most of the western portion of the event domain. There is also an area of enhanced statistically significant negative composite differences just east of the domain, suggesting that less-extreme events have stronger

VIWVF. This area aligns with the region of statistically significant enhanced 850-mb winds for less-extreme events seen in Figure 3.21.

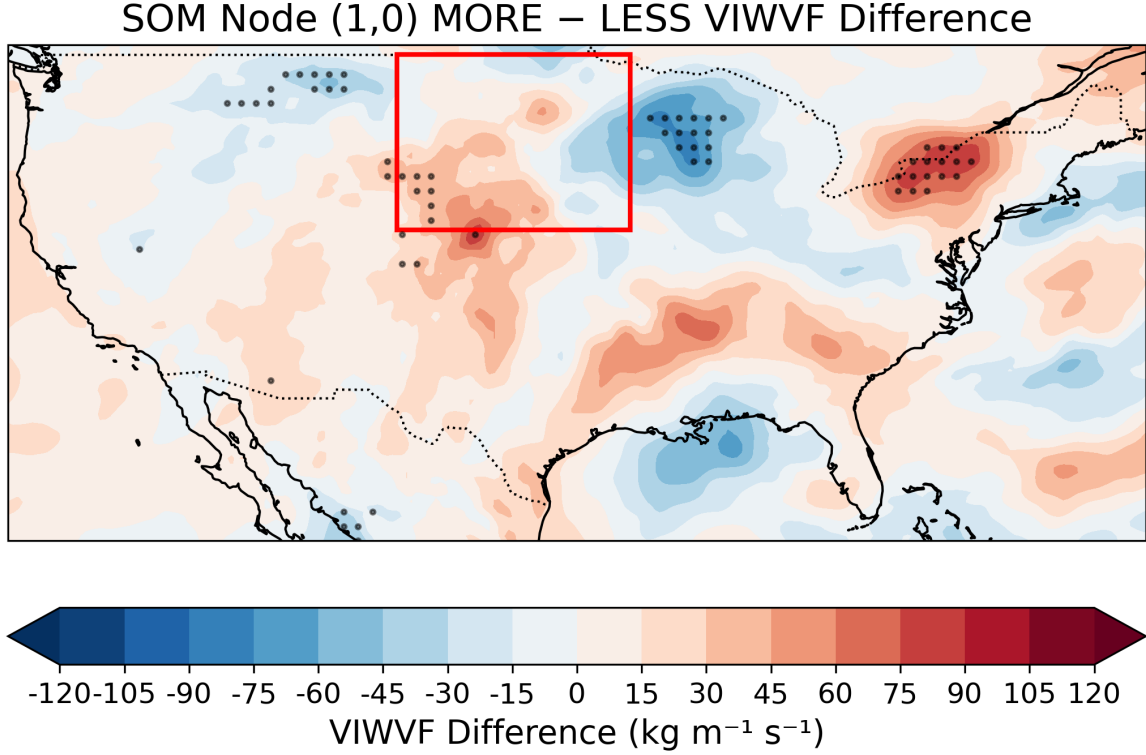


Figure 3.25: Composite difference (more - less) in vertically integrated water-vapor flux (VIWVF) for Node (1,0). Shading shows VIWVF differences ($\text{kg m}^{-1} \text{s}^{-1}$). Stippling marks grid points where the composite difference is statistically significant at the 95% significance level. The boxed region denotes the event domain.

More-extreme events have clear localized areas of stronger VIWVF convergence compared to less-extreme events (Fig. 3.26a and 3.26b). Consistent with this, localized positive VIWVF convergence differences within the event domain suggest that more extreme events are associated with stronger localized moisture convergence (Fig. 3.26c). Although the magnitude of these differences is modest, they are concentrated over the event domain and consistent with the stronger LLJ and enhanced identified for more-extreme events within this node regime. VIWVF A portion of these areas is also found to be statistically significant. These statistically significant stronger lo-

calized areas of convergence are similar to what was seen in the initial bulk VIWVF convergence composite difference plot in section 3.1.5.

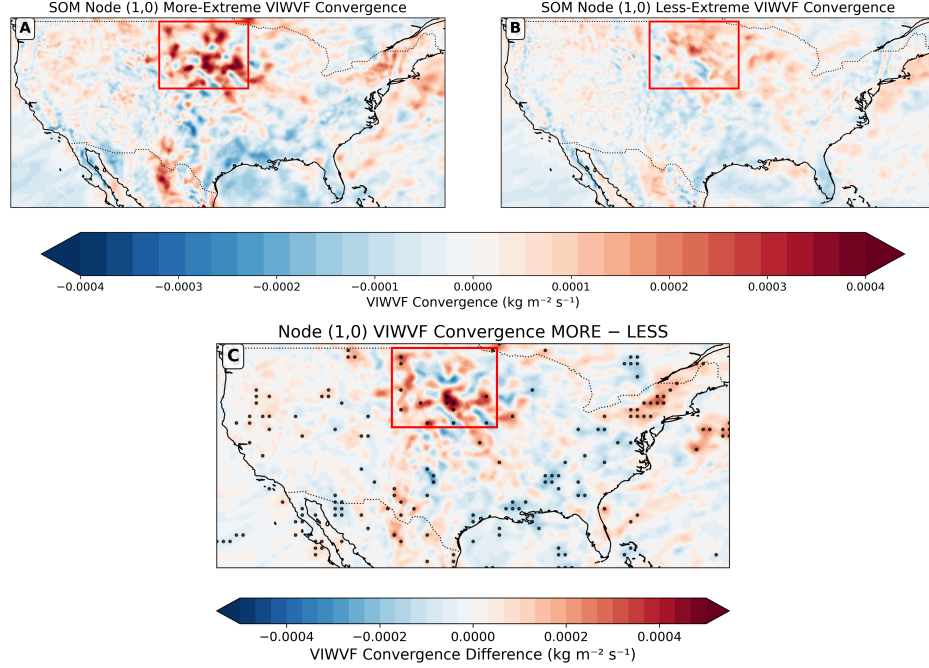


Figure 3.26: Composite vertically integrated water-vapor flux (VIWVF) convergence fields ($\text{kg m}^{-2} \text{s}^{-1}$) for Node (1,0) (a) for more-extreme events, (b) less-extreme events and (c) composite difference (more - less) in VIWVF convergence. Stippling marks grid points where the composite difference is statistically significant at the 95% significance level. The boxed region denotes the event domain.

3.4.2 Node (2,0)

Moving down to the lower levels of the atmosphere, at 850-mb, positive wind speed composite difference values extend over parts of the Great Plains region and most of the event domain, with few areas showing statistical significance (Fig. 3.27). Compared to node (1,0), there are fewer areas of statistical significance. Even with less statistical significance compared to the previous node, there is a pattern of more-extreme events associated with stronger 850-mb winds than less-extreme events within this area. East of this broad area of positive differences, there is also a pattern of negative composite

differences extending up from the Gulf of Mexico into southeast portions of the event domain with a very few, localized area of statistical significance. This pattern indicates less-extreme events having stronger 850-mb winds than more-extreme events within this area.

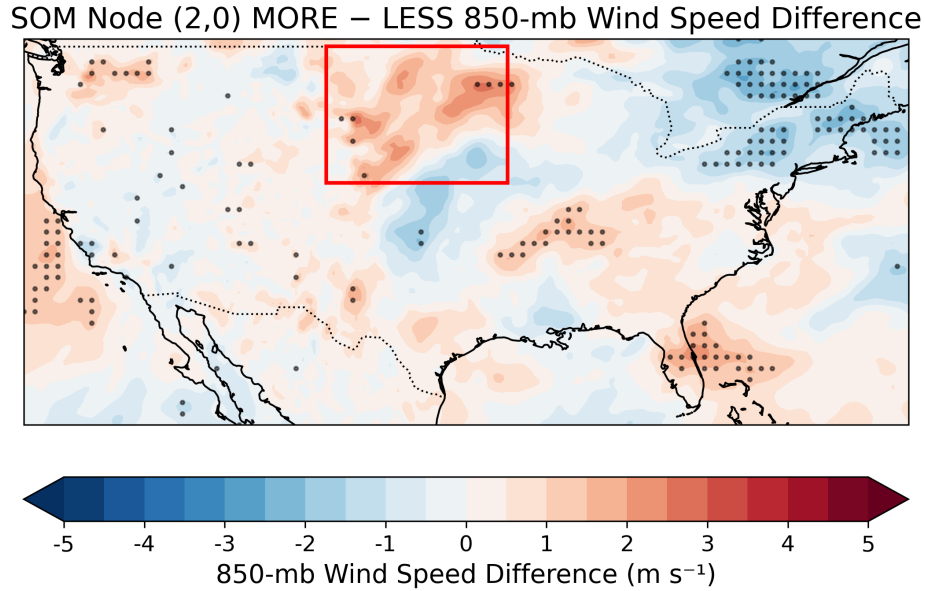


Figure 3.27: Composite difference (more – less) in 850-mb wind speed (m s^{-1}) for Node (2,0). Shading indicates the magnitude of the wind-speed difference. Stippling marks grid points where the composite difference is significant at the 95% significance level. The boxed region denotes the event domain.

Also, at 850-mb, both more-extreme and less-extreme events exhibit very similar large-scale temperature advection spatial patterns (Fig. 3.28a and 3.28b), consistent with the behavior seen in the previous node. However, localized positive differences are present within the event domain, where more-extreme events exhibit more enhanced warm air advection adjacent to regions of more enhanced cold air advection relative to the less-extreme events (Fig. 3.28c). Similar to node (1,0), the northern half of the event domain is dominated by positive composite difference values, with portions of this region exhibiting statistical significance. This pattern suggests that more enhanced warm air advection associated with more-extreme events may extend farther north

compared to less-extreme events, which aligns with the stronger 850-mb wind speeds shown in Figure 3.29. These results suggest that intensified low-level flow contributes to more efficient heat transport during more-extreme events.

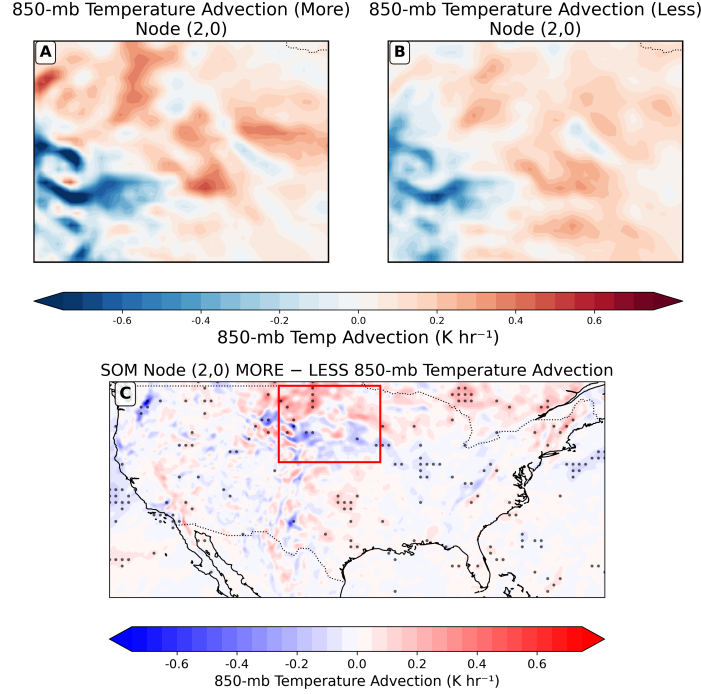


Figure 3.28: Composite 850-mb temperature-advection fields (K h^{-1}) for Node (2,0) for (a) more-extreme events, (b) less-extreme events and (c) the composite difference (more - less) in 850-mb temperature advection. Stippling marks grid points where the composite difference is statistically significant at the 95% significance level. The boxed region denotes the event domain.

Analysis of the TCWV patterns reveals that, unlike node (1,0), only the northern portion of the domain has positive values of TCWV differences, indicating that more-extreme events will have higher values of TCWV within this area than less-extreme events (Fig. 3.29). This pattern is like the pattern of the bulk TCWV composite differences in section 3.1.4. However, unlike the bulk composite, this area does not have any statistical significance. Similar to node (1,0) the concentration positive TCWV differences is consistent to the spatial distribution of more-extreme events. The more-extreme events centroid is shifted farther north relative to the less-extreme events

centroid (Fig. 3.30a and 3.30b). The position of the more-extreme events centroid is also consistent with the location of enhanced 850-mb wind speeds and enhanced warm air advection (Fig. 3.27 and 3.28c). The centroid for the less-extreme events is closer to the center of the event domain, similar to node (1,0) suggesting that less-extreme events are more evenly distributed within the event domain. The southern region of the domain is dominated by negative TCWV values, indicating that for this region of the domain, less-extreme events have higher values of TCWV. Additionally, most of the southern region has statistical significance. This pattern suggests that more-extreme events in this node rely less on moisture pooled in the domain as a whole and more on the efficient northward transport and concentration of moisture in the convection region. This is also consistent with the pattern of more enhanced warm air advection in the northern region of the domain for more-extreme events, as seen in Figure 3.27c, suggesting that the LLJ is playing a key role in transporting both warmer air and moisture into this area.

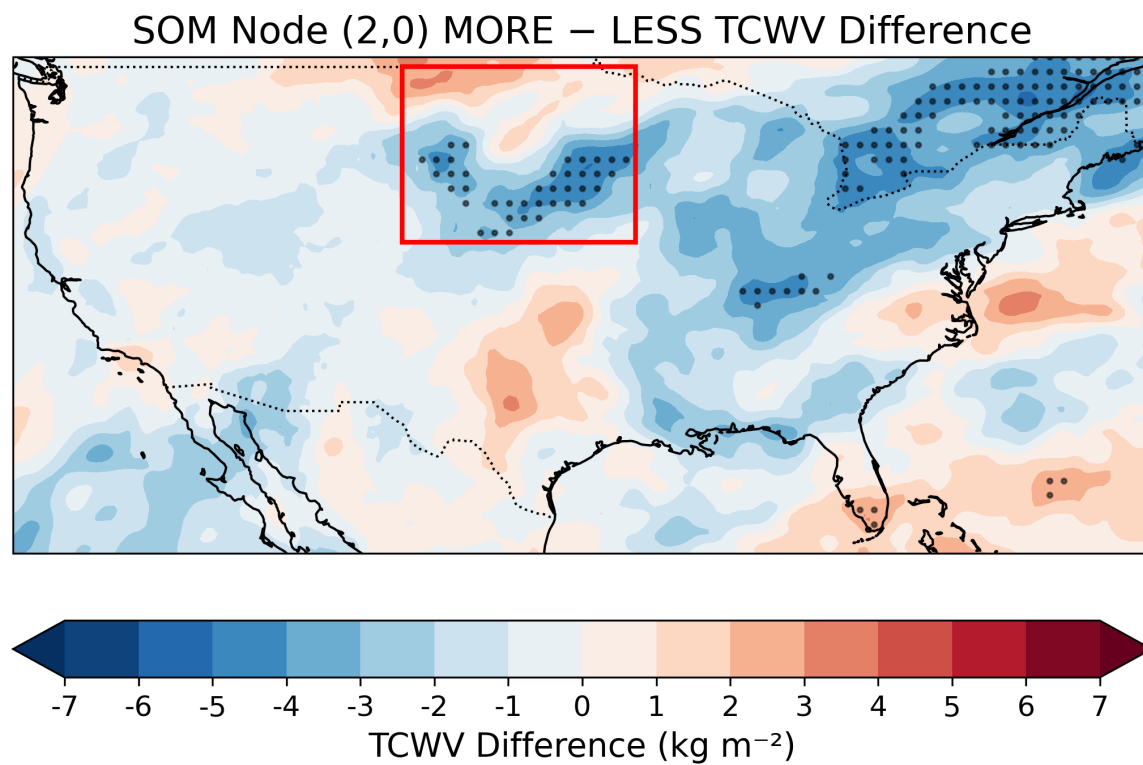


Figure 3.29: Composite difference (more - less) in total column water vapor (TCWV) for Node (2,0). Shading shows TCWV differences (kg m^{-2}). Stippling marks grid points where the composite difference is statistically significant at the 95% significance level. The boxed region denotes the event domain.

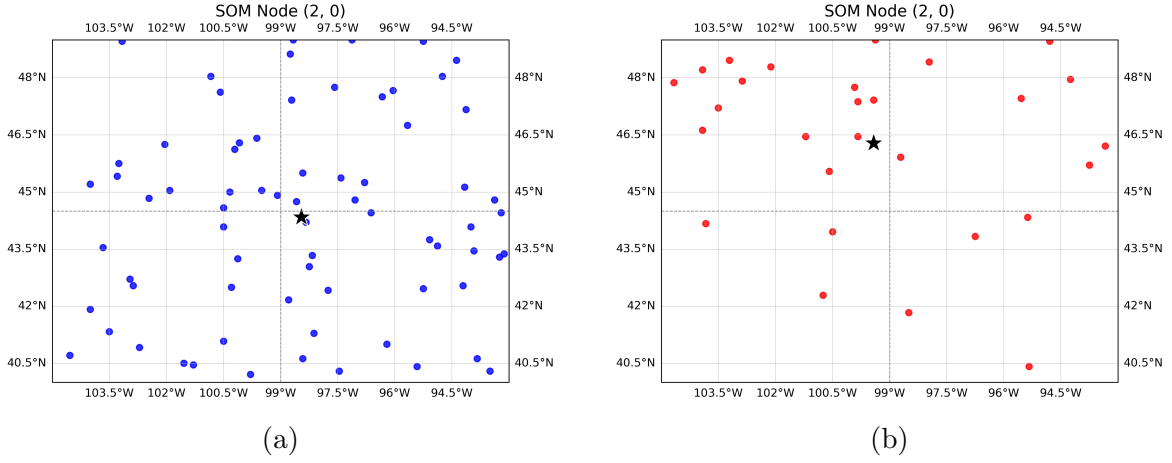


Figure 3.30: Event locations for SOM Node (2,0) shown for (a) less-extreme events and (b) more-extreme events. Each panel displays the spatial distribution of events within the analysis domain, with the black star marking the centroid of all events associated with the node. Dashed lines indicate the centroid latitude and longitude of the event domain.

Moisture transport patterns reveal a broad area of positive composite difference values that extended over most of the Great Plains and into the event domain (Fig. 3.31). These positive values indicate higher values of VIWVF for more extreme rainfall events. Only a couple of small, localized areas exhibit statistical significance. However, this pattern aligns with the pattern of stronger 850-mb winds for more-extreme events seen in Figure 3.27 which suggests that the LLJ plays an important role behind the VIWVF pattern within this node regime. There is also an area east of this pattern, across portions of the eastern Great Plains, that has higher values of VIWVF for less-extreme events. This also aligns with the negative difference pattern seen in the 850-mb wind composite difference plot again emphasizing the importance of the LLJ for moisture transport.

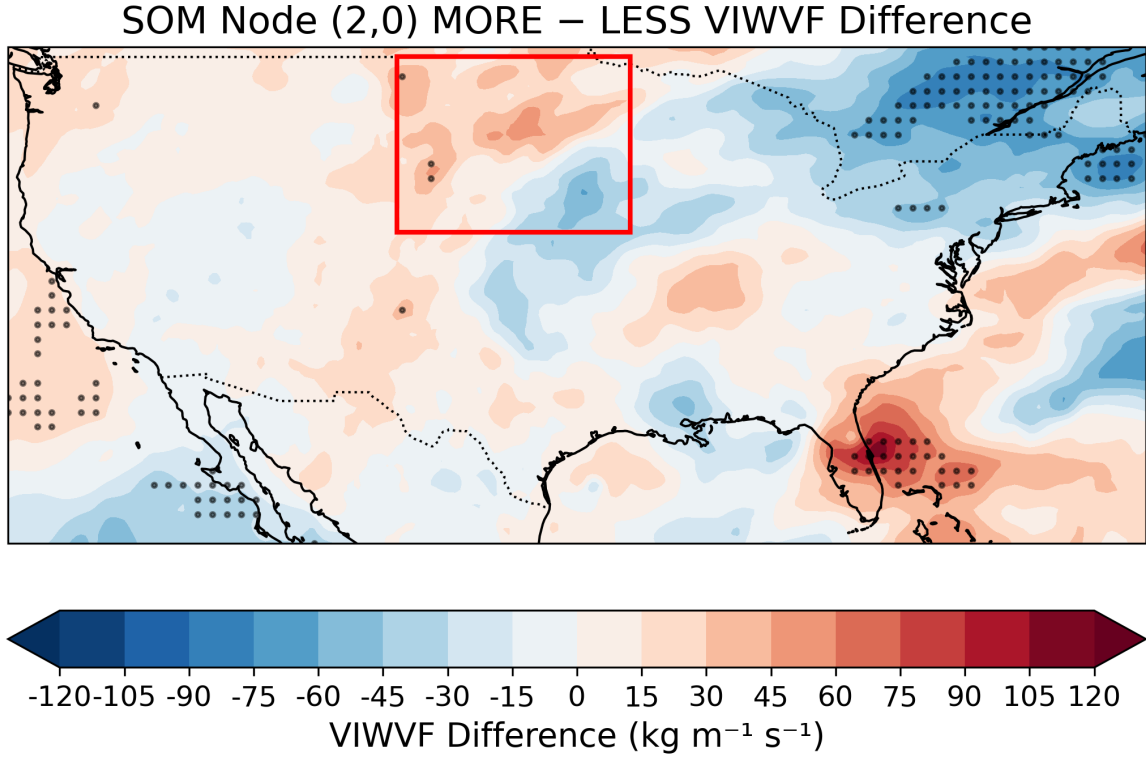


Figure 3.31: Composite difference (more - less) in vertically integrated water-vapor flux (VIWVF) for Node (2,0). Shading shows VIWVF differences ($\text{kg m}^{-1} \text{s}^{-1}$). Stippling marks grid points where the composite difference is statistically significant at the 95% significance level. The boxed region denotes the event domain.

The pattern of VIWVF convergence is less distinct in node (2,0) compared to node (1,0), with fewer regions exhibiting clear enhanced VIWVF convergence for more extreme events (Fig. 3.32a and 3.32b). This suggests that differences in VIWVF convergence between more-extreme and less-extreme events are less spatially coherent, implying that stronger moisture convergence is a less robust discriminator between extreme types within this node. Still, localized areas of positive VIWVF convergence composite differences within the event domain indicate locations where more-extreme events have stronger VIWVF convergence (Fig. 3.32c). Some of these areas also show statistical significance indicating that these areas consistently have stronger VIWVF convergence during more-extreme rainfall events.

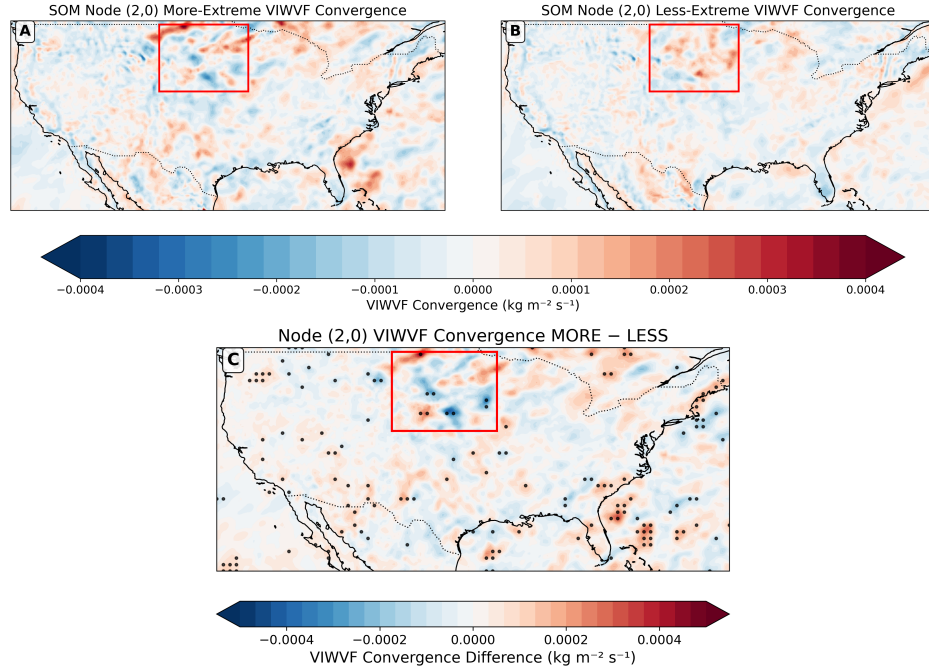


Figure 3.32: Composite vertically integrated water-vapor flux (VIWVF) convergence fields ($\text{kg m}^{-2} \text{s}^{-1}$) for Node (2,0) (a) for more-extreme events, (b) less-extreme events and (c) composite difference (more - less) in VIWVF convergence. Stippling marks grid points where the composite difference is statistically significant at the 95% significance level. The boxed region denotes the event domain.

3.4.3 Node (0,2)

At lower levels, the environmental differences between more-extreme and less-extreme become clearer. There is an area of broad statistically significant positive differences over the western portion of the event domain and just outside of the western end of the event domain (Fig. 3.33). This pattern indicates that this area of positive differences, more-extreme events consistently have stronger 850-mb winds. There is also a broad area of enhanced negative composite differences extending up from the Gulf of Mexico into the eastern portion of the event domain and areas just east of the domain, with a few areas of statistical significance. This suggests that there may be a clear pattern of stronger 850-mb winds for less-extreme events within these areas. The few areas of

statistical significance may be the result of more-extreme events having a much smaller sample size than the less-extreme events.

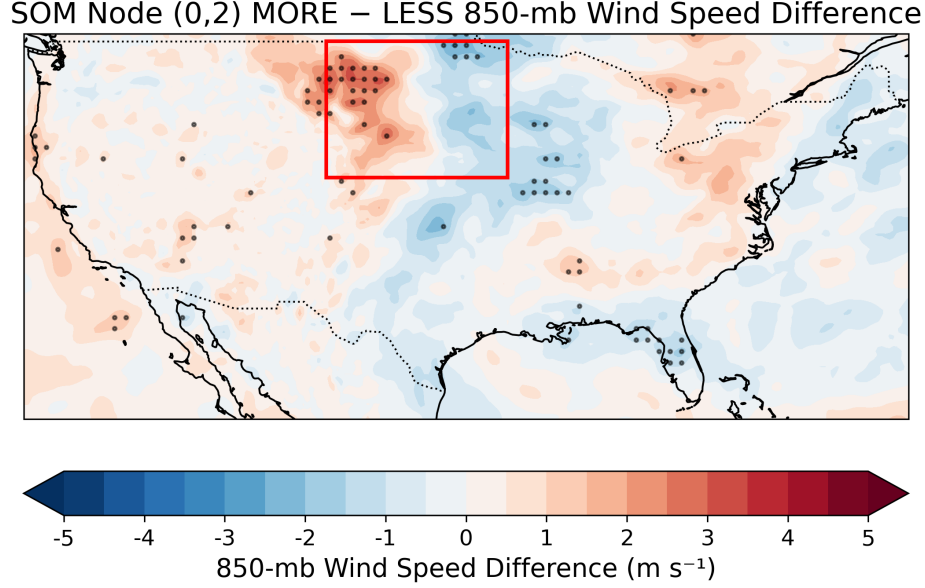


Figure 3.33: Composite difference (more – less) in 850-mb wind speed (m s^{-1}) for Node (0,2). Shading indicates the magnitude of the wind-speed difference. Stippling marks grid points where the composite difference is significant at the 95% significance level. The boxed region denotes the event domain.

Continuing in the lower levels of the atmosphere, more-extreme and less-extreme events show a consistent pattern of a broad area of warm air advection extending over most of the event domain similar to the previous nodes (Fig. 3.34a and 3.34b). Additionally, like the previous nodes, more enhanced areas of localized warm air advection are evident for more-extreme events. This pattern is confirmed through the composite difference plot that shows localized areas of positive differences (Fig. 3.34c). However, only some of the localized areas exhibit statistical significance. This finding suggests that the difference between event types in this regime may depend more on localized areas of enhanced warm air advection rather than large differences in broad areas of warm air advection.

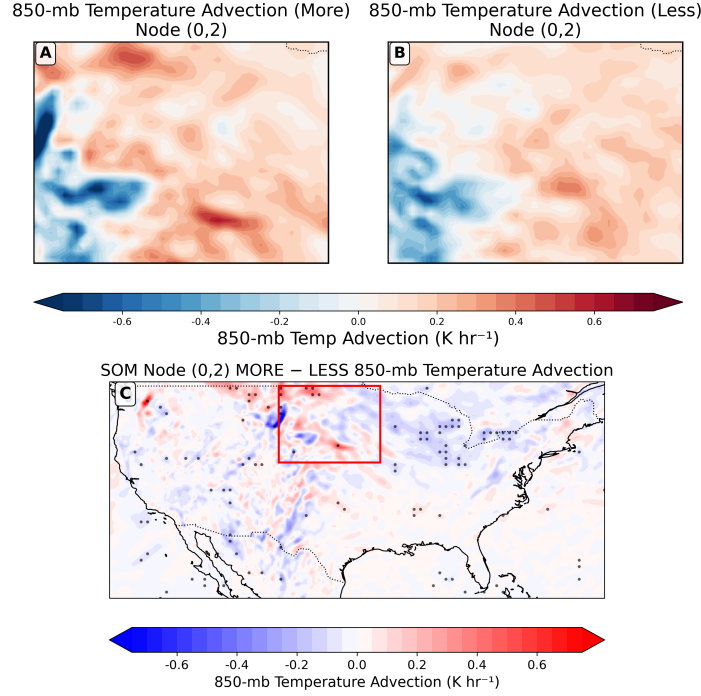


Figure 3.34: Composite 850-mb temperature-advection fields (K h^{-1}) for Node (0,2) for (a) more-extreme events, (b) less-extreme events and (c) the composite difference (more - less) in 850-mb temperature advection. Stippling marks grid points where the composite difference is statistically significant at the 95% significance level. The boxed region denotes the event domain.

Compared to the previous nodes, there is less variability in TCWV between more-extreme and less-extreme events within the event domain (Fig. 3.35). This is the case for most of the CONUS as well. There are also no areas of statistical significance within the event domain and across most of the Great Plains region. Throughout the Great Plains region, there are multiple areas of weak positive and negative TCWV difference values but there is no clear TCWV pattern across the region. The centroid of more-extreme events is located west/northwest of the domain center (Fig. 3.36b). This may be due to an area of enhanced TCWV for more-extreme events in the north-western corner of the event domain and just outside the western border of the domain. This suggests that the spatial distribution of higher TCWV values may influence the preferred location of more-extreme events within the event domain. However, the lack

of a clear TCWV pattern within this node regime limits the ability to draw a consistent relationship between TCWV structure and event distribution. The less-extreme events centroid is again located closer to the event domain center, like the previous nodes, suggesting that less-extreme events are more evenly distributed and appear less dependent on the magnitude and distribution of TCWV (Fig. 3.36a).

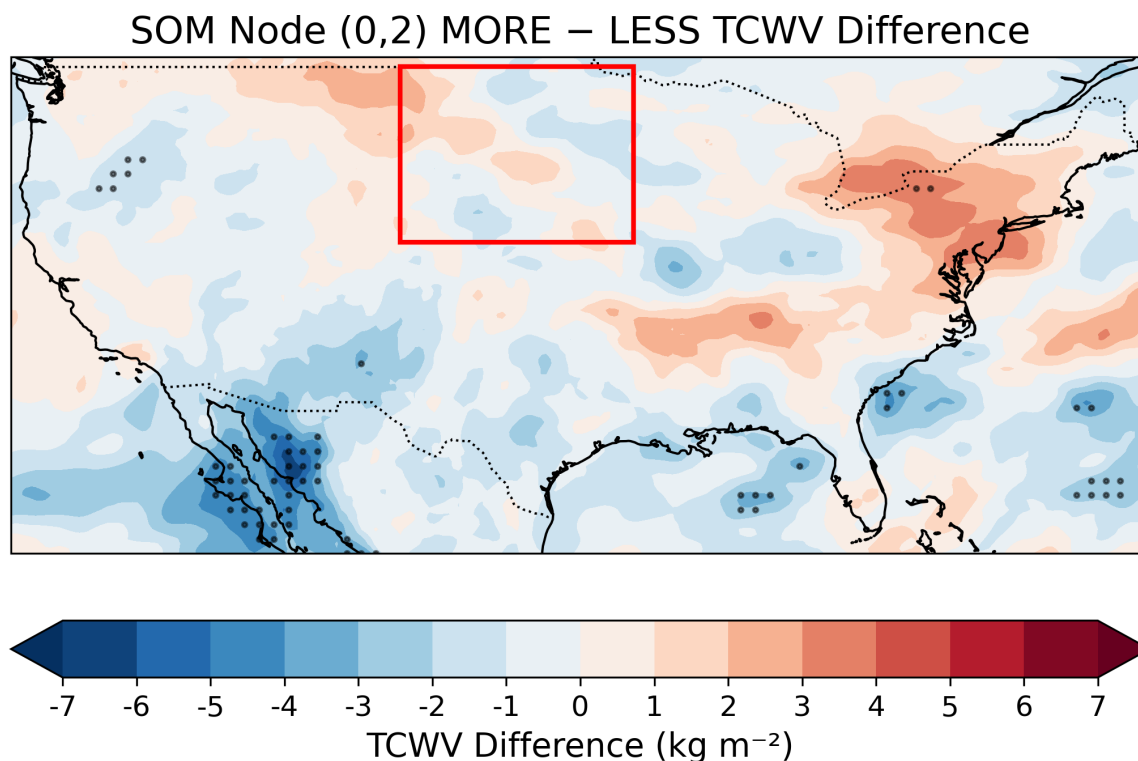


Figure 3.35: Composite difference (more - less) in total column water vapor (TCWV) for Node (0,2). Shading shows TCWV differences (kg m^{-2}). Stippling marks grid points where the composite difference is statistically significant at the 95% significance level. The boxed region denotes the event domain.

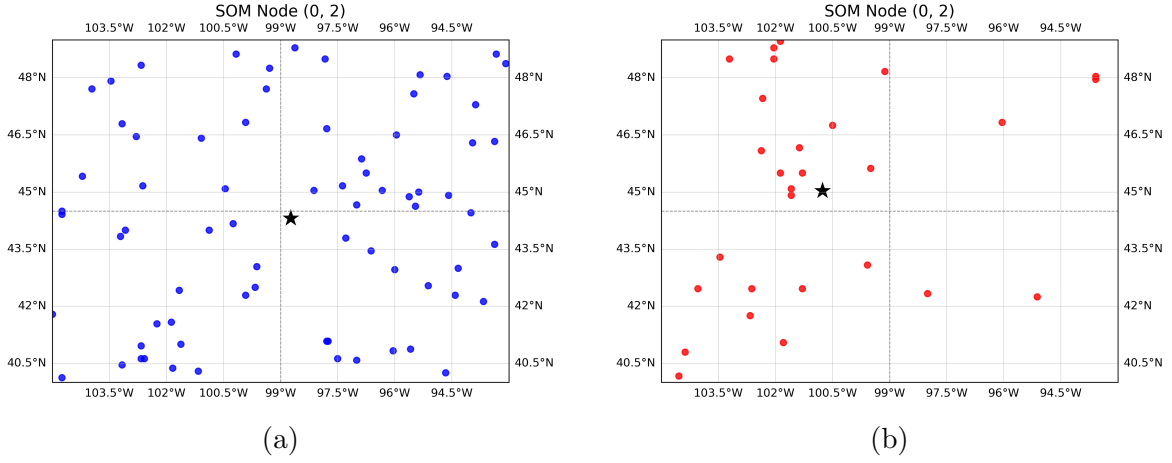


Figure 3.36: Event locations for SOM Node (0,2) shown for (a) less-extreme events and (b) more-extreme events. Each panel displays the spatial distribution of events within the analysis domain, with the black star marking the centroid of all events associated with the node. Dashed lines indicate the centroid latitude and longitude of the event domain.

There is a broad area of enhanced negative difference values extending up from the Gulf of Mexico into the eastern portion of the event domain, with a few small areas of statistical significance (Fig. 3.37). This suggests a clear pattern of stronger VIWVF for less-extreme events, and this pattern aligns with the enhanced 850-mb wind pattern in Figure 3.36. There is also an area of enhanced positive composite difference values in the western portion of the event domain that aligns with an area of enhanced positive 850-mb wind composite differences (Fig. 3.33). These findings continue to emphasize the key role of LLJ in transporting moisture up from the Gulf of Mexico. This area of positive differences has a broader area of statistical significance, indicating consistently higher values of VIWVF in these areas for more-extreme events. This corridor of enhanced VIWVF and 850-mb wind speeds for more-extreme events is consistent with the westward positioning of the more-extreme events centroid (Fig. 3.36b). This suggests a possible connection between the structure of VIWVF and the LLJ and more-extreme event distribution.

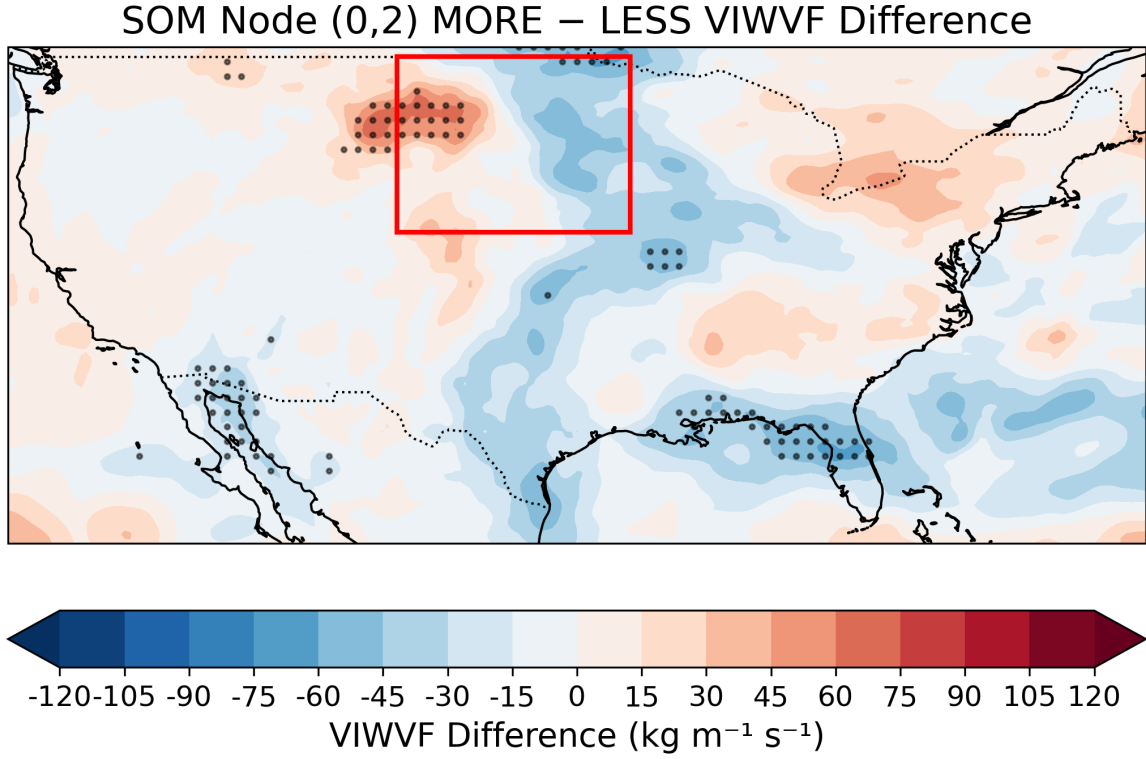


Figure 3.37: Composite difference (more - less) in vertically integrated water-vapor flux (VIWVF) for Node (0,2). Shading shows VIWVF differences ($\text{kg m}^{-1} \text{s}^{-1}$). Stippling marks grid points where the composite difference is statistically significant at the 95% significance level. The boxed region denotes the event domain.

When looking at the more-extreme and less-extreme event VIWVF convergence plots, similar to node (1,0), the more-extreme events show a clear pattern of stronger localized VIWVF convergence throughout the event domain (Fig. 3.38a and 3.38b). This is confirmed through the VIWVF convergence composite difference plot where there are localized areas of positive composite differences in the same locations where the enhanced convergence was for more-extreme events (Fig. 3.38c). Some of these areas also show statistical significance suggesting that localized areas of enhanced VIWVF convergence may play a key role in distinguishing between extreme severity.

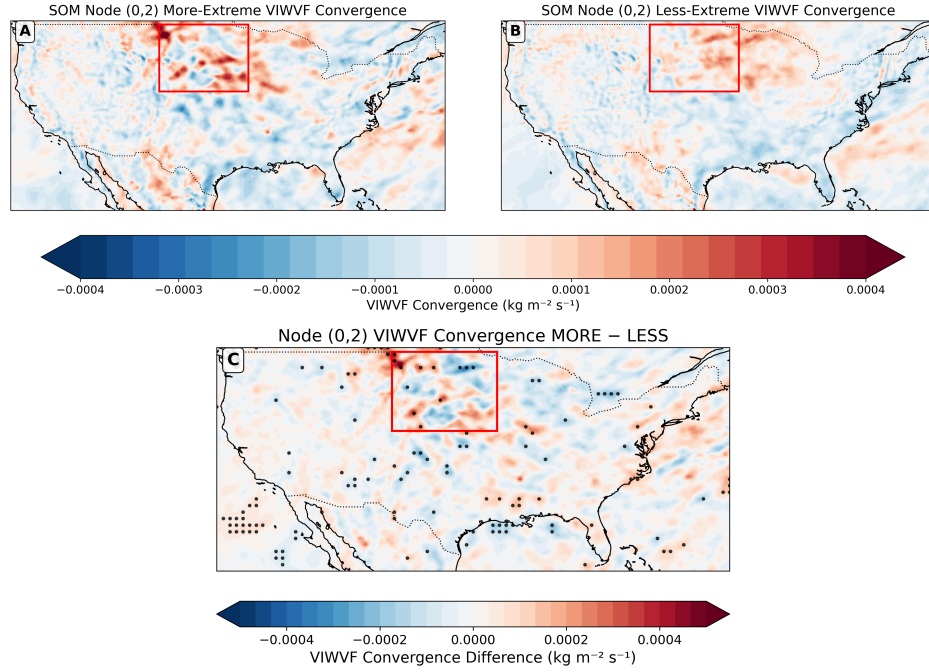


Figure 3.38: Composite vertically integrated water-vapor flux (VIWVF) convergence fields ($\text{kg m}^{-2} \text{s}^{-1}$) for Node (0,2) (a) for more-extreme events, (b) less-extreme events and (c) composite difference (more - less) in VIWVF convergence. Stippling marks grid points where the composite difference is statistically significant at the 95% significance level. The boxed region denotes the event domain.

3.5 500-mb Precursor Synoptic Patterns

To investigate the large-scale atmospheric conditions leading up to summertime nocturnal extreme rainfall events in the Northern Great Plains, 500-mb geopotential height anomalies are examined the day before and three days before the event onset. These anomalies are also compared to the anomalies at the time of the events to understand the evolution of the setup. The anomaly differences between more-extreme and less-extreme events are also analyzed to determine if precursor 500-mb environments can be used to distinguish between extreme events.

3.5.1 Node (1,0)

For node (1,0), three days before the time event onset, negative anomalies are present over the western CONUS (Fig. 3.39a). From three days before to one day before event onset, these anomalies strengthen substantially, indicating a pronounced amplification of the 500-mb synoptic-scale pattern (Fig. 3.39b). These anomalies also migrate farther east while an area of positive anomalies begins to develop over the Great Lakes region. Although the magnitude of anomalies continues to increase between one day before and the event day itself, the magnitude of this increase is much smaller, suggesting a deceleration in the rate of intensification leading up to the event (Fig. 3.39c). This 500-mb geopotential height anomaly evolution suggests that there is clear amplification of the trough-ridge pattern during the days leading up to the events within this node regime. The slower rate of intensification between the day before and the event day indicates that the large-scale pattern may become established prior to the event onset.

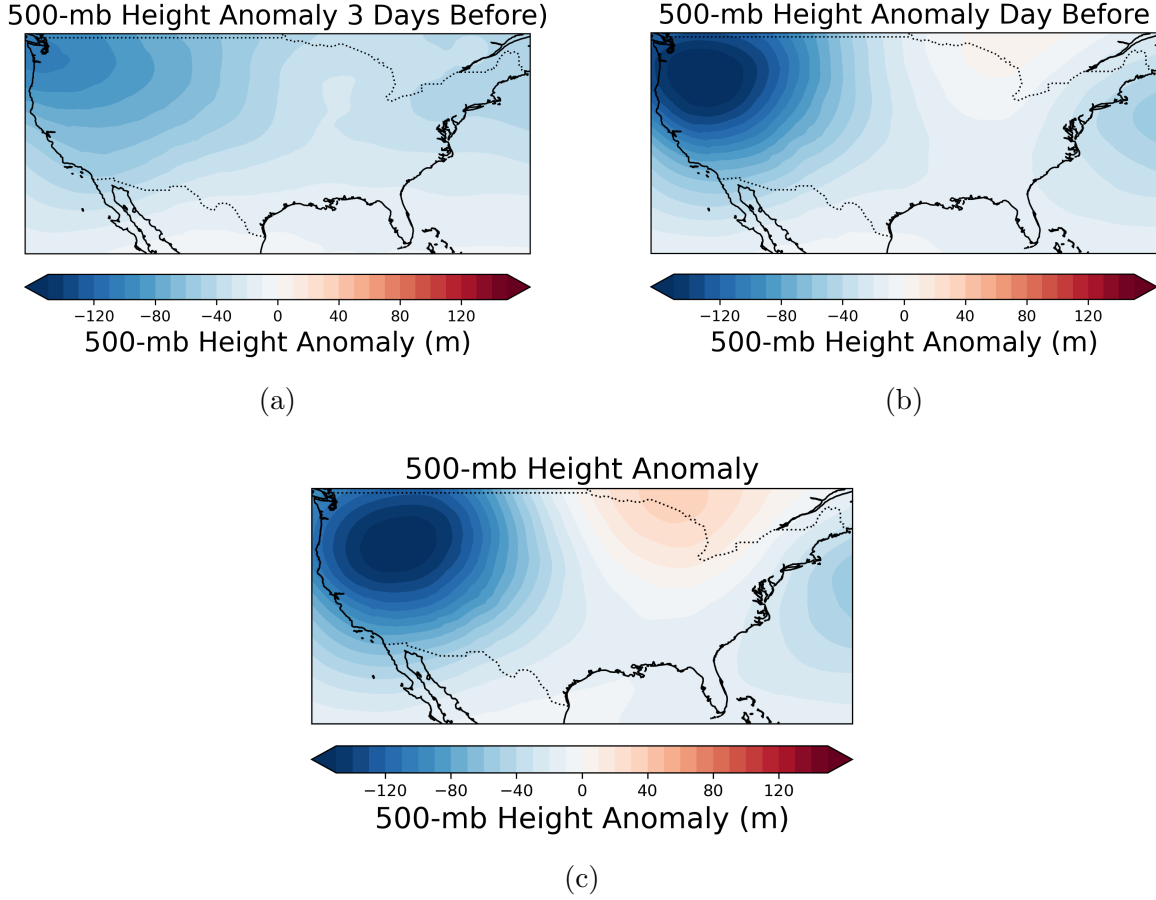


Figure 3.39: 500-mb geopotential height anomalies (m) for (a) 3 days before, (b) 1 day before, and (c) the day of the events for node (1,0).

Across both time periods, more-extreme events exhibit slightly stronger negative anomalies over the western CONUS (Fig. 3.40 and 3.41). Similar differences are present in positive anomalies over the Great Lakes. Although these differences are apparent three days before the event, an organized pattern becomes evident closer to event onset. However, the magnitude of these differences remains weak with no statistical significance between more-extreme and less-extreme events in any of the examined time periods (Fig. 3.40c and 3.41c). These results indicate that the evolution of the 500-mb synoptic-scale pattern is broadly similar between more-extreme and less-extreme events. Because the SOM was trained on 500-mb geopotential height

anomalies, the similar evolution observed suggests that differences in the 500-mb precursor synoptic environment are not a primary factor in distinguishing between event intensity. This pattern also suggests that other precursor processes such as those in the lower-troposphere may be needed to identify precursor patterns that distinguish between event intensity.

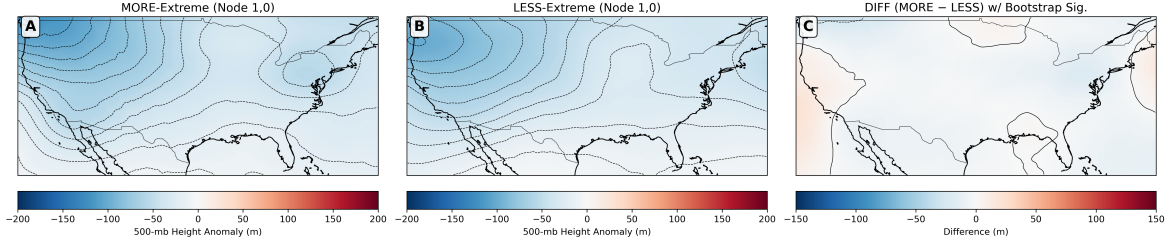


Figure 3.40: 500-mb geopotential height anomalies three days before events for Node (1,0) under more-extreme (panel a) and less-extreme (panel b) event conditions, along with their difference (panel c).

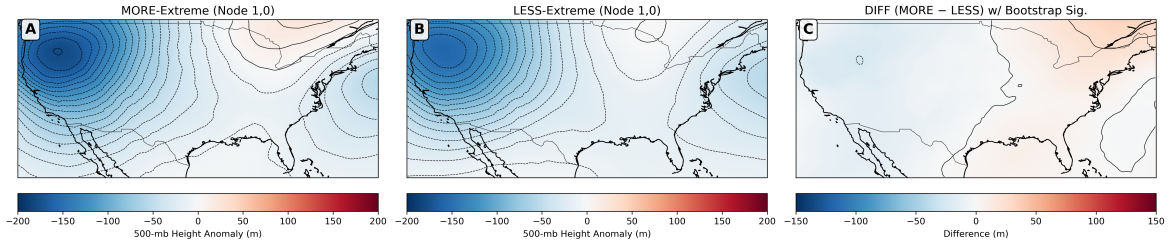


Figure 3.41: 500-mb geopotential height anomalies on the day before events for Node (1,0) under more-extreme (panel a) and less-extreme (panel b) event conditions, along with their difference (panel c).

3.5.2 Node (2,0)

For Node (2,0), three days before event onset, negative anomalies are present over the western CONUS, and positive anomalies are present over the northeastern CONUS (Fig. 3.42a). Similar to the previous node, from three days before the event to one day before event onset, these anomalies strengthen substantially, indicating a pronounced amplification of the synoptic-scale pattern closer to event onset (Fig. 3.42b). Also

similar to the previous node, the anomalies between the day before the event and the day of the event change very little indicating that the large-scale pattern may already be established prior to event onset (Fig. 3.42c).

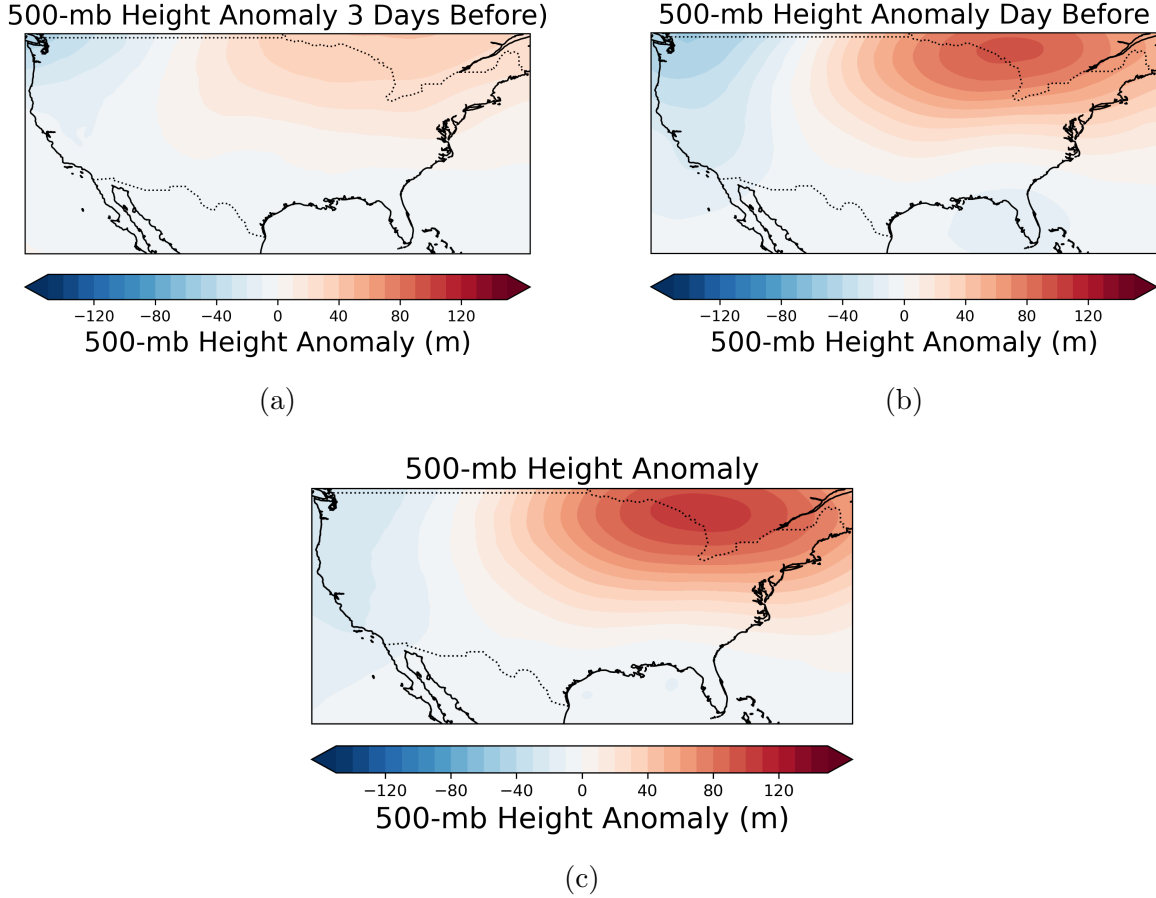


Figure 3.42: 500-mb geopotential height anomalies (m) for (a) 3 days before, (b) 1 day before, and (c) the day of the events for node (2,0).

Differences between geopotential height anomalies are present three days before event onset. An anomaly difference pattern becomes more evident the day before event onset, with negative anomaly differences occurring over portions of the West and East Coast and positive anomaly differences occurring over a large portion of the central United States. Like node (1,0), despite these differences, no statistically significant differences are found between more-extreme and less-extreme events at any

of the examined time periods (Fig. 3.43 and 3.44). Because the SOM was trained on 500-mb geopotential height anomalies, this pattern suggests that the evolution in the 500-mb synoptic-scale pattern continues to not be a primary factor distinguishing between more-extreme and less-extreme events within this node regime and further analysis of other precursor processes at other levels is needed.

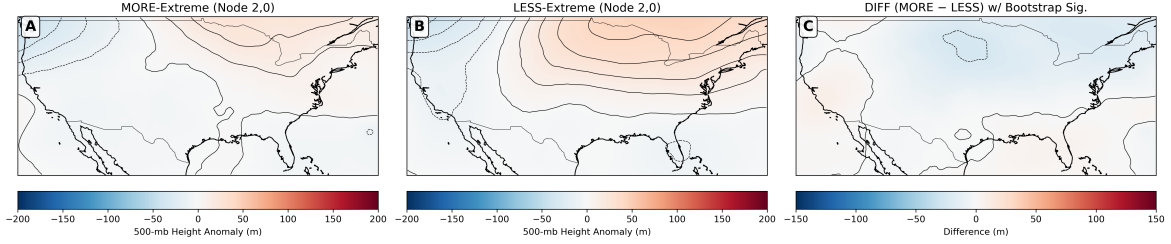


Figure 3.43: 500-mb geopotential height anomalies three days before events for Node (2,0) under more-extreme (panel a) and less-extreme (panel b) event conditions, along with their difference (panel c).

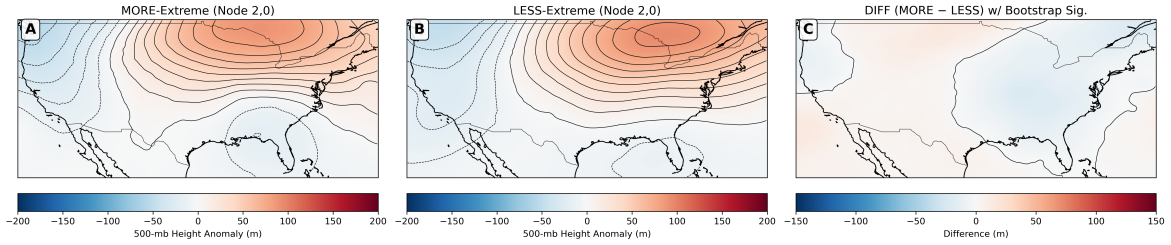


Figure 3.44: 500-mb geopotential height anomalies on the day before events for Node (2,0) under more-extreme (panel a) and less-extreme (panel b) event conditions, along with their difference (panel c).

3.5.3 Node (0,2)

Three days before the event onset, negative anomalies are present over the western United States, and positive anomalies are already present over the central and eastern United States (Fig. 3.45a). From three days before the event to one day before, these anomalies strengthen in magnitude but not as substantially as seen in the prior two node regimes (Fig. 3.45b). The magnitude of anomalies between the day before

the event and the day of the event change very little (Fig. 3.45c). Again, like both nodes, the evolution of 500-mb geopotential height anomalies suggests that there is clear amplification of the trough-ridge pattern leading up to the event onset, with the ridge anomaly strengthening substantially throughout the pre-event period.

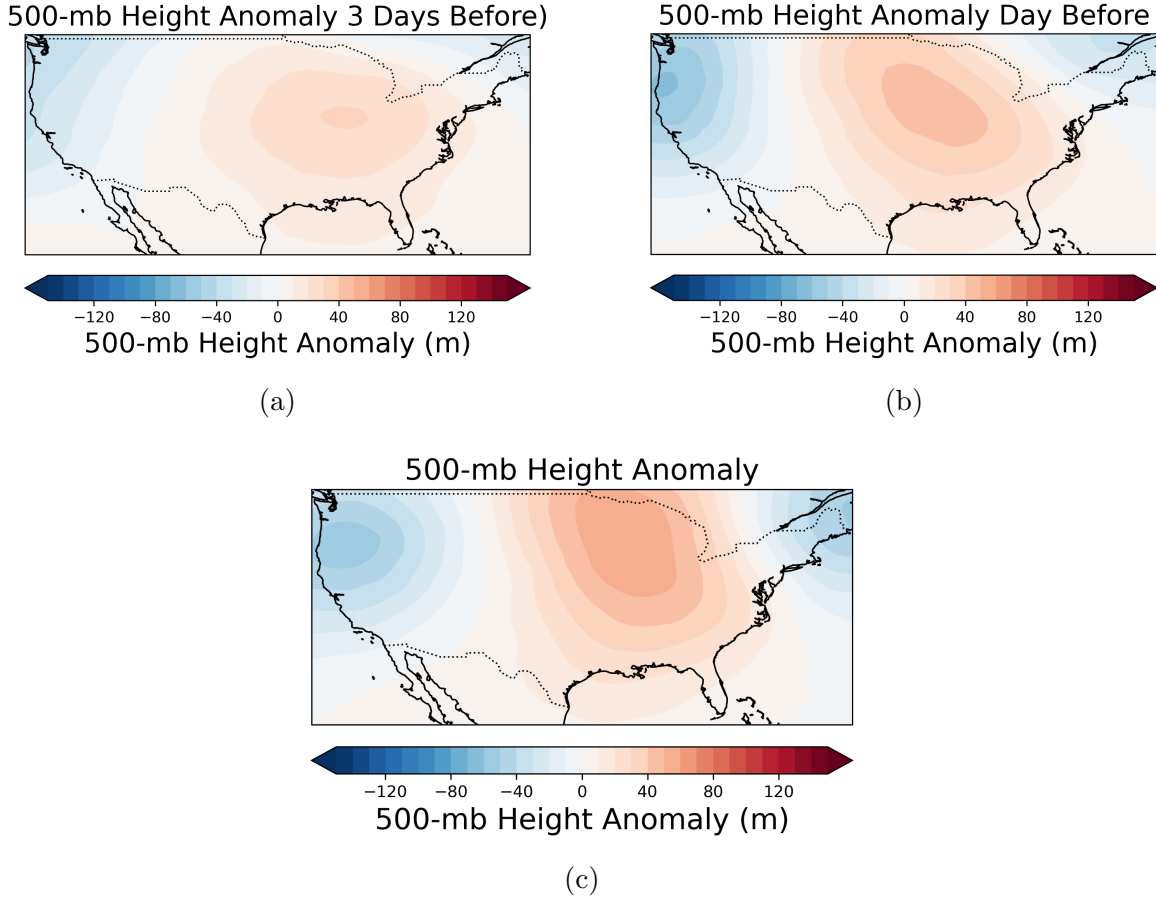


Figure 3.45: 500-mb geopotential height anomalies (m) for (a) 3 days before, (b) 1 day before, and (c) the day of the events for node (0,2).

Like node (1,0) and (2,0), there are no statistically significant 500-mb geopotential height anomaly differences between more-extreme and less-extreme events at any of the examined time periods (Fig. 3.46c and 3.47c). However, unlike the previous nodes, a more organized pattern of differences is evident three days before but becomes less apparent the day before event onset. Despite this, the lack of statistically significant

differences suggests that the evolution in the 500-mb synoptic-scale pattern may not be a primary factor distinguishing between more-extreme and less-extreme events.

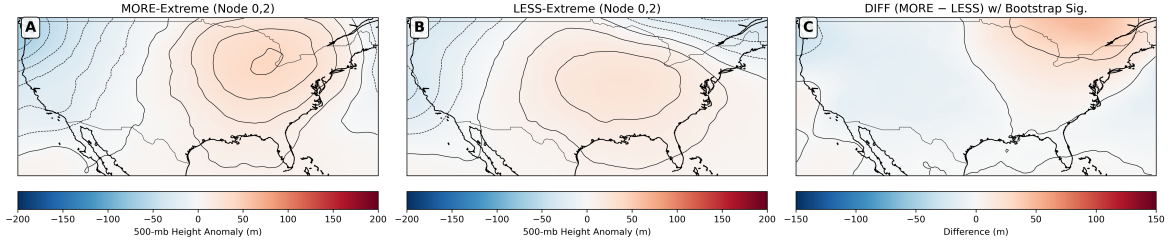


Figure 3.46: 500-mb geopotential height anomalies (m) three days before events for Node (0,2) under more-extreme (panel a) and less-extreme (panel b) event conditions, along with their difference (panel c).

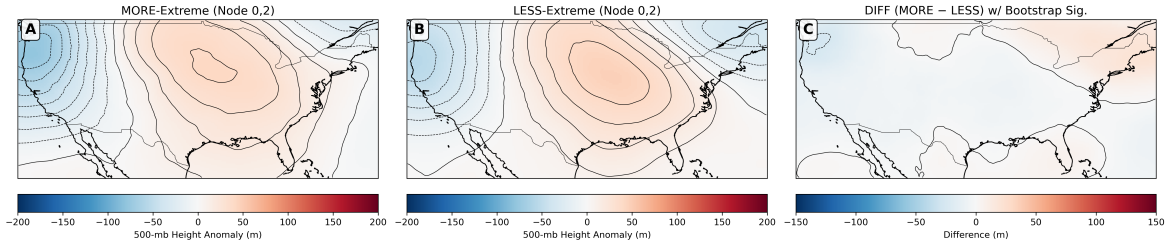


Figure 3.47: 500-mb geopotential height anomalies (m) on the day before events for Node (0,2) under more-extreme (panel a) and less-extreme (panel b) event conditions, along with their difference (panel c).

Chapter 4

Summary and Conclusions

Collectively, the environmental characteristics observed across all node regimes examined in this study support the conceptual models of summertime nocturnal rainfall in the Great Plains and heavy rainfall. Each node in this study has the event domain located downstream of a 500-mb trough and near a 500-mb ridge. This finding is consistent with the results from Maddox et al. (1979) who showed that heavy rainfall events often occur downstream of a mid-level trough and near the western half of a large-scale ridge. This synoptic setup promotes ascent and enhanced moisture transport which creates a favorable mid-level environment for these events. Across all nodes, enhanced TCWV extends across much of the central and eastern United States which is similar to TCWV patterns found by Chiappa et al. (2024) during active ERE years. This enhanced TCWV also extends up until just before the northern end of the event domain. The nose of the NLLJ is consistently positioned just to the south of the event domain, resulting in jet deceleration within the domain. As discussed earlier in this paper, this setup of the NLLJ is consistent with the findings from Higgins et al. (1997) which showed that there is a nocturnal maximum of extreme rainfall located near the maximum NLLJ wind speeds. This configuration leads to widespread 850-mb warm air advection across the event domain. The VIWVF pattern across all nodes closely mirrors the NLLJ pattern, with the nose of the VIWVF corridor also located just to the south of the domain. This VIWVF pattern also leads to the deceleration of VIWVF within the event domain, favoring accumulation of moisture and VIWVF convergence.

This is confirmed by VIWVF convergence composites, which shows localized areas of enhanced VIWVF throughout the event domain within each node regime. These features are consistent with those identified in the bulk composites, suggesting that the bulk composites capture fundamental environmental characteristics common to nocturnal summertime extreme rainfall event across the Northern Great Plains. While the SOM exhibits distinct large-scale synoptic regimes, the moisture transport and atmospheric forcing patterns associated with these events remain consistent across all three node regimes. This finding suggests that similar atmospheric processes support extreme rainfall development and/or maintenance within the Northern Great Plains region, regardless of the broader synoptic setup. However, all nodes should be evaluated to determine if these atmospheric processes are consistent within all nine node regimes. Additionally, the relative importance of these processes may vary upon regions due to different regional environmental features such as moisture sources, terrain, and dominant synoptic patterns.

While the bulk and node composites reveal a common set of environmental features associated with summertime nocturnal extreme rainfall events in the Northern Great Plains, important differences emerge when events are separated into more-extreme and less-extreme categories. The bulk composites tended to obscure important distinctions between more-extreme and less-extreme events. By stratifying events by node, differences in the environments became more apparent. This was the case for 850-mb temperature advection, with the SOM node regimes analyzed showing a clear pattern of enhanced localized warm air advection within the event domain for more-extreme events, some of which were statistically significant. A previous study showed that widespread events were associated with stronger low level warm air advection compared to more concentrated events (Stevenson and Schumacher, 2014). Although event classifications differ, both suggest that enhanced low-level warm air advection plays an

important role differentiating between ERE types. In contrast, the bulk composites showed no differences in temperature advection that were statistically significant between the two extreme types. Additionally, for VIWVF, the bulk composite difference showed a clear pattern of a statistically significant corridor of stronger VIWVF values for more-extreme events across most of the Great Plains. In contrast, the node regime composites differences, even without large areas of statistical significance, showed a clear corridor of higher VIWVF for more-extreme events over the Western Great Plains with a clear corridor of higher VIWVF for less extreme events over the Eastern Great Plains. This suggests that the spatial distribution of the moisture transport from the Gulf of Mexico may be as important as its overall magnitude in distinguishing more-extreme from less-extreme events, providing a potential forecast signal for identifying environments for favorable for extreme rainfall. This aligns with a previous study done by ? who found a strong connection between the location of integrated vapor transport and EREs.

While certain features, such as the 850-mb warm air advection, consistently distinguish between more-extreme and less-extreme events, other features depend on the underlying node regime. Nodes (1,0) and (0,2) had a clear pattern of localized statistically significant enhanced VIWVF convergence for more-extreme events. This pattern also occurred in the bulk composite differences. In contrast, node (2,0) showed very few, and less organized localized areas of statistically significant enhanced VIWVF convergence for more extreme events. This finding suggests that enhanced moisture convergence is not a universally distinguishing characteristic of more-extreme events across all node regimes. This finding is consistent with Banacos and Schultz (2005) who found that moisture flux convergence may be helpful in forecasting convection initiation but can not be used as a primary predictor due to its significance relying on the surrounding environmental conditions. A similar pattern is observed for TCWV, as the

spatial distribution of TCWV differed for all three nodes evaluated in this study. Node (1,0) showed enhanced statistically significant TCWV for more-extreme events across most of the Great Plains region and into the event domain. However, node (2,0) had a pattern of enhanced statistically significant TCWV for less-extreme events throughout the southern portion of the event domain and a pattern of enhanced TCWV for more-extreme events within northern portions of the event domain. For both of these nodes, areas of enhanced TCWV associated with more-extreme events appears to be consistent with the spatial distribution of more-extreme events within the event domain. The centroid of more-extreme events is located closer to the regions of enhanced TCWV. This finding indicates a potential link between event extremity and moisture distribution. Node (0,2) showed no clear pattern of TCWV across the Great Plains region. These differences likely indicate both variability in the synoptic environments and the distribution of events within each node, indicating that TCWV is not a consistent variable that can be used to distinguish between extreme types within 500-mb height anomaly defined node regimes. However, if the node regimes were defined by other variables, such as TCWV, different relationships between moisture variability and event severity may appear.

For all node regimes evaluated in this thesis, there is a clear pattern of amplification of the trough-ridge pattern leading up to event onset. This pattern is consistent across more-extreme and less-extreme event environments. This evolution in the mid-level pattern is consistent with previous studies that showed that EREs are commonly associated with the amplification of a trough-ridge pattern (e.g., Moore et al., 2015; Milrad et al., 2019). In return, the amplification in the trough-ridge pattern is also expected to strengthen southwesterly flow and further support moisture transportation into the region of events. However, differences in magnitude of the 500-mb setup between more-extreme events and less-extreme events are weak and not statistically

significant. Given that the SOM was trained on 500-mb height anomalies, this finding suggests that 500-mb geopotential height anomaly precursor environments may not be a primary factor in distinguishing event severity within these node regimes. The lack of strong differences in the 500-mb pattern is consistent with the expectation that moisture variables and mesoscale processes may provide more direct controls on extreme rainfall intensity in regimes organized by mid-level synoptic patterns.

Overall, the bulk composites were found to effectively identify the common environment characteristics associated with summertime nocturnal extreme rainfall events across the Northern Great Plains. Comparison with the SOM node regimes showed that many of these key features were consistently present, regardless of the large-scale synoptic pattern. These results suggest that bulk composites provide a useful representation of the fundamental environmental conditions favorable for nocturnal extreme rainfall events in this region. However, the bulk composites were unable to fully capture all the different patterns that distinguish more-extreme events from less-extreme events. The SOM analysis demonstrated an ability to not only identify common 500-mb geopotential height anomaly patterns at event onset, but also to reveal both similar and differing patterns between more-extreme and less-extreme events within these node regimes.

From an operational forecasting perspective, these results demonstrate how SOMs can potentially be useful for identifying large-scale synoptic regimes and anticipating how moisture transport and lower atmospheric forcing is expected to evolve. Rather than relying on traditional composites, which provide only a single mean representative pattern, forecasters could use SOMs as a pattern recognition framework. Forecasters could first identify which synoptic regime best represents the current environment, and then evaluate the moisture transport and lower atmospheric forcing patterns to differentiate if the environment favors more-extreme or less-extreme events. Although

additional work is needed before this approach can be used in an operational setting, these results show that there is potential for SOMs to be used to improve the forecasting of nocturnal summertime EREs in the United States Northern Great Plains.

Future work should expand on the analysis to investigate and compare all nine SOM nodes to determine whether the patterns identified in this study are consistent across the full range of node regimes. Additional atmospheric variables, such as convective available potential energy (CAPE), vertical wind shear, and other measures of convective organization, should also be examined to determine if any additional patterns can be identified with the SOM. Extending this methodology to other regions across the United States would further assess the applicability of this SOM approach for identifying environmental patterns associated with extreme rainfall events. Future work should also incorporate field significance testing in addition to the grid-point bootstrap significance testing used in this study. This method would help to further determine whether observed patterns are meaningful rather than the result of random chance.

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