A SYNOPTIC CLIMATOLOGY OF NOCTURNAL RAINFALL EVENTS DURING MAY, JUNE AND JULY FOR NORTHEAST KANSAS, 1950-2012

by

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Abstract

Nighttime rainfall has long been thought of as an important component to the central Great Plains hydroclimate during the wettest three-month period known as the "late spring - early summer precipitation maximum." Research has suggested that nocturnal rainfall in the region results from a phenomenon known as the nocturnal Great Plains Low-Level Jet (GPLLJ). The jet, which originates in the Gulf of Mexico, transports moisture into the Great plains during the nighttime hours and often provides fuel for nighttime convection. The climatological characteristics of nighttime rainfall, as well the configuration of the low-level winds and the mechanisms behind its formation during this three-month wet period, however; are not well understood. Using hourly rainfall data from Topeka, KS, the nighttime rainfall characteristics are examined Topeka, KS and other Kansas stations for a 63-year period from 1950-2012 for May-July. Additionally, using the NCEP/NCAR Reanalysis data, the structure and configuration of the southerly wind phenomenon was analyzed based on its horizontal and vertical characteristics for nighttime rainfall events in May, June and July. A subsequent analysis also analyzed the larger synoptic-scale environment in place for six half-month periods from May to July. The results indicate that nighttime rainfall is a major contributor to the overall moisture budget in the Great Plains, contributing close to 50% of the overall rainfall total for the three-month period. The percentage of nighttime rainfall increases from west-east across the state, as well as temporally from May to July. The southerly winds are at their strongest during May events, tends to reach its peak at 850 mb at 6z (0000LST) near south-central Oklahoma, and forms as the result of both synoptic and thermal mechanisms. The synoptic mechanisms in place that generate the a southerly wind component change by month, leading to incredible variation in terms of its characteristics during nighttime rainfall events.

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At orientation during my first day at Kansas State, I was assigned an advisor in the history department. The advisor never showed up, and instead a professor working that day kindly offered to be an advisor for a few students needing to select classes. That advisor, Dr. M.J. Morgan, soon became one of the best professors and most influential and helpful people I have ever met. Not only did you help me find my niche in historical studies, but you also encouraged me to make the switch to geography. Your inspiring words and consistent belief in me have led me to where I am now, and without that fateful first day at orientation who knows where I would be at this point in my life.

And last, but certainly not least, is my family. My wife, Courtney, has been there through it all with me. Marrying you will always be the best decision I've ever made. You push me to be better in every aspect of life. Going forward, I will do everything in my power to make you proud of me, and I'm so excited to see what the future holds for us. I also want to thank my mother, father and yes even my sister (alright and even all the animals that live at the house). Your guidance and patience with me will hopefully not go unrewarded. This is only one accomplishment in what I hope to be many down the road, and all of them will be the result of everything you have done for me. Even though we live apart now, I still think of you every day and do my best to make you proud; that will never change. Love you guys!

Dedication

I dedicate this thesis the two my favorite coaches, Courtney, my wife, and the man they call Coach Dome, Skyler Yeaton. Love you both so much.

CHAPTER ONE
INTRODUCTION TO NIGHTTIME RAINFALL IN THE CENTRAL GREAT PLAINS

"If I were running the world I would have it rain only between 2 and 5 a.m. Anyone who was out then ought to get wet."

~William Lyon Phelps

1.1 Overview:

The frequency and intensity of precipitation in the central Great Plains of the United States varies greatly throughout the year. In the spring and summer months, the region receives the bulk of its precipitation, with numerous nocturnal rainstorms contributing to this seasonal maximum. The warm-season precipitation maximum occurs in May, June and July, and forms as a result of a number of atmospheric properties that combine to promote thunderstorm activity across the Great Plain. Atmospheric circulation is an influential element of the Great Plains hydroclimate since moisture transports greatly contribute to local precipitation. The predominance of warm season nighttime rainfall in the Great Plains remains one of the interesting climatic features in North America (Trewartha, 1981). During the warm season, the combination of intense surface heating (leading to a very high temperature lapse rate), high influxes of moisture, as well as a location protected from oceanic influences would normally create a strong daytime maximum in diurnal rainfall. However, the Great Plains does not fit the mold, as warm season precipitation in the region has a pronounced diurnal variability (Pitchford and London, 1962; Balling, 1985). Precipitation fluctuations are common on daily and sub-daily time scales in the region and comprise showers and thunderstorms. The precipitation over the central Great Plains exhibits a nocturnal diurnal peak, differing from the afternoon maximum over most inland regions (Jiang et al, 2006). In the Great Plains more than 60% of summer (June-July-August) thunderstorms occur during 1800-0600 local standard time (LST), with over 50% of summer thunderstorms occur during the period of 0000-0600LST (Pitchford and London, 1962). Rainfall in the Great Plains also displays a distinct west-east diurnal characteristic, as areas in the west experience the most precipitation during early evening to late evening hours, whereas areas eastward receive the most precipitation in the early morning hours (Figures 1.01 and 1.02).

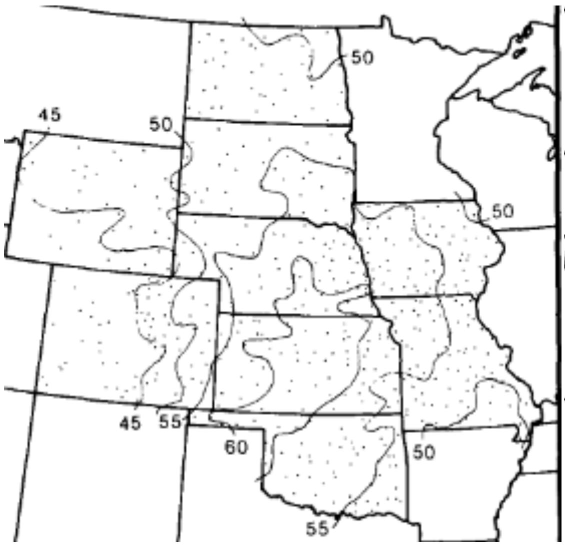


Figure 1.01: Percentage of nocturnal (2000-0800 TST) warm season precipitation frequencies greater than .25mm per hour.

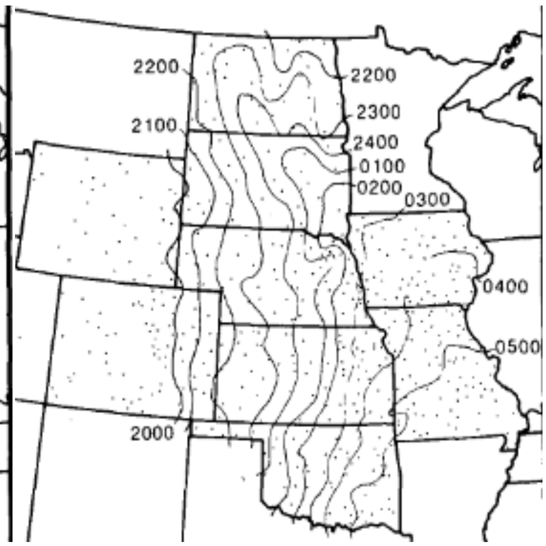


Figure 1.02: time of maximum rainfall frequency (Reproduced from Balling, 1985).

Nocturnal precipitation events have long been thought to have been connected with a southerly low-level wind maximum known as the Great Plains low-level jet (GPLLJ). The GPLLJ can span from the Gulf Coast up into the northern Great Plain and range at speeds of 25 to 75 knots (Banta et al*.,* 2002). Vertical placement of the low-level jet in the atmosphere tends to vary based on various mechanisms discussed later in this study, although, the low-level wind maximum generally occurs in the lowest kilometer of the troposphere (Figure 1.03) (Whiteman et al*.,* 1997).

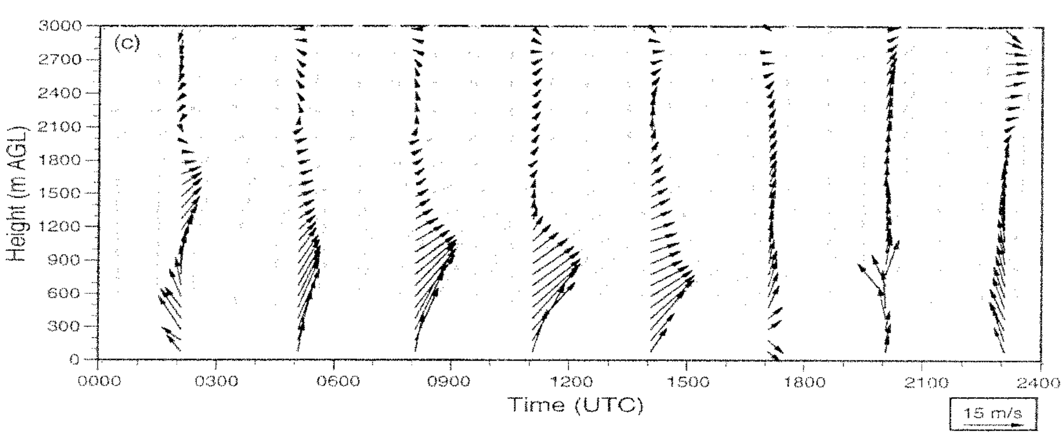


Figure 1.03: Rawinsonde wind profiles at Oklahoma ARM CART site on 31 July 1994 displaying the characteristics of the nocturnal low-level jet (Whiteman et al.., 1997).

1.2 Research Goals, Questions and Objectives:

Although numerous studies exist regarding the diurnal characteristics of the nocturnal low-level jet and its importance to nighttime rainfall, there still lacks a long-term, comprehensive climatology that analyzes nighttime rainfall on a case-by-case basis. More specifically, a long-term synoptic climatology analyzing the circulation patterns associated with nighttime rainfall events for the central Great Plains is not known to exist. Given the relatively frequent occurrence of nighttime thunderstorms during late spring and early summer, developing a climatology spanning a period of multiple decades relating nocturnal rainfall to atmospheric circulation patterns would be pertinent towards better discerning the relationship between nocturnal rainfall and the synoptic environment.

Moreover, given the concerns of precipitation variability in the face of climate change in the Great Plains, analyzing if the characteristics of nighttime rainfall have changed through time could perhaps contribute to regional studies of climate change. Therefore, this thesis attempts to advance the study of the diurnal rainfall variability, using a synoptic-climatological approach to investigate the circulation patterns present during nighttime rainfall events. Utilizing hourly precipitation data for Topeka as well as other stations across the state of Kansas, this thesis develops a 63 year synoptic climatology of the late spring-early summer precipitation maximum in the central Great Plains. Chapter Two summarizes relevant literature and Chapter Three of the thesis provides a more in-depth discussion of the methods used in the research. Naturally, the various components of this research raise an abundant number of research questions. The research questions addressed include:

I. What are the temporal and spatial characteristics of late spring-early summer nighttime rainfall in Kansas?

II. What are the characteristics of the low-level wind maximum during these rainfall events?

III. What other meteorological mechanisms are needed to develop a southerly nocturnal low-level wind and increase precipitation chances for north-central Kansas during the nighttime hours?

IV. Are there any discernible temporal patterns that exist with the number of wind maxima events occurring on an hourly/daily/monthly or longer time period?

To address the above questions, the specific objectives of this research are:

I. Identify, define and statistically analyze nocturnal rainfall events first for Topeka and then at other stations in Kansas from 1950-2012 during the months of May, June and July. Compare nocturnal events based on a daily, monthly, seasonal, annual, and multi-year basis across the stations being analyzed

II. Describe the characteristics of the nocturnal low-level winds for each nocturnal precipitation event

III. Identify the larger-scale synoptic patterns in place during these rainfall events. Describe the identified synoptic patterns in terms of near surface and mid-tropospheric atmospheric variables.

IV. Compare and contrast the characteristics of the various synoptic patterns associated with nocturnal events

Prior to addressing the research questions listed above, however, it is important to provide an in-depth look at the various methods of synoptic climatology assessment and how they can be applied to a research topic such as nighttime rainfall. Additionally, given the broad scope of mid-tropospheric jet stream and low-level jet stream research, understanding what contributions have been made to the topic can offer a basic framework for this particular research as well as reveal areas that need further examination. Chapter two provides a relevant review of existing literature that helps frame the rationale for and the scholarly contribution developed in the thesis.

CHAPTER TWO:
LITERATURE REVIEW OF CENTRAL UNITED STATES NOCTURNAL RAINFALL, SYNOPTIC CLIMATOLOGY AND LOW-LEVEL JET RESEARCH

"Twice and thrice over, as they say, good is it to repeat and review what is good."

-Plato

2.1: Nocturnal Rainfall in the Great Plains:

Although extensive research has been dedicated to the seasonal variation of the hydrologic cycle in the Great Plains, very few studies have examined the mechanisms which trigger a late-spring precipitation maximum, and more specifically, the mechanisms behind an increase in nighttime precipitation (Wang and Chen, 2009). Though, previous research provides a variety of hypotheses regarding the various influences that create an increase in warm season nocturnal rainfall in the region. Convective activity has been the most common association with the warm-season precipitation maximum in the Great Plains. Deep convection is often coupled with the synoptic-scale environments that promote upward motion of the air through a large area of the vertical (Carleton et a*l.,* 2008). For instance, upper-level disturbances have long been associated with thunderstorm development, as these short waves moving through the long wave upper-level westerly pattern provide pulses of energy needed for uplift and thunderstorm development. In a similar manner, the passage of fronts provides a triggering mechanism for rapid upward motions of air surface and resulting convection. Additional theories attribute nighttime rainfall to the advection of evaporation-cooled layers from afternoon showers near the Rockies as a possible cause. Afternoon rain on the Front Range of the Rockies could perhaps destabilize the nighttime atmosphere of the central plains (Hales, 1977). Studies analyzing the life cycle of mesoscale convective complexes (MCCs) and mesoscale convective systems (MCSs) have shown their eastward propagation in the afternoon and nighttime hours largely accounts for the nocturnal character of warm season rainfall (Maddox, 1980). The tracks of MCSs and MCCs can stretch for hundreds and miles, persist for hours, and these storms can bring copious amounts of moisture to the central Great Plains. Quasi-linear convection systems (QLCSs) are also a common feature during the warm season and result from large-scale upper-tropospheric dynamics such as divergence, vorticity advection, and the passing of frontal boundaries, all of which are associated with wind maxima in the vertical, baroclinic waves, surface cyclones as well as warm and cold fronts. (Carleton et al., 2008). In the absence of larger synoptic features, land-surface influences have been attributed as the culprit for convection. While much of the convective precipitation in the central United States results from warm air advection and moisture transport from the Gulf of the Mexico, convection can also result from local destabilization associated with land surface evaporation.

Though the mechanisms are well-known, it continues to be a well-recognized problem that most of the current global circulation models (GCMs) have trouble simulating the diurnal cycle of warm season precipitation. The GCMs tend to develop rainfall too early in the day and produce too little precipitation at night, due in large part to the models over-reliance and sensitivity to diurnal heating in the boundary layer (Lee et al., 2010).

2.2 Synoptic Climatology:

Synoptic climatology has been described as an explanation of local climates in terms of large-scale circulation. Founded during World War II by the Army Air Forces' Weather Service, synoptic climatology describes the totality of weather conditions over a small region (Court 1957). Synoptic climatology has been given varying definitions over the years. According to Barry and Perry (1973), all synoptic climatology studies share two characteristics, 1.) atmospheric circulation classifications, and 2.) the assessment of the relationship between atmospheric circulation categories and the weather elements at a given location or region. Court further believed that synoptic climatology should utilize synoptic scale variables such as air pressure patterns, air masses and map patterns when analyzing and categorizing atmospheric characteristics. Yarnal (1993) added two characteristics; stating synoptic climatology examines climate variability of the surface environment and focuses on the spatial unit of a region. Additionally, Yarnal considered the surface environment to a particular meteorological phenomenon, the overall weather conditions in the planetary boundary layer or the non-direct meteorological phenomenon such as air pollution in a region. More recently, Barry and Carelton (2001) defined the field as the study of the relationship between local and regional climate conditions to the atmospheric circulation. Synoptic climatology can also be broken down into four different aspects that help define the approach: airflow, pressure-fields, air masses, and map-patterns (Court, 1957). As upper-air data became available after World War II, researchers began to relate patterns in the upper troposphere with surface features. With increased data across the globe and its accessibility, synoptic studies on all temporal and spatial scales have been conducted (Yarnal et al*.,* 2002). For instance, on the hemispheric scale, studies of ENSO and other teleconnections have been examined. On the local level, circulation features responsible for micro- and meso-scale phenomena such as severe weather have become important areas of interest (Harman, 1991).

2.3 Background:

Following World War II, synoptic climatology focused on synoptic patterns corresponding to surface weather. Many synoptic climatological studies concentrated on developing classification methods and themes, which led to two highly regarded manually derived synoptic classifications in the 1970s: the Lamb Weather Types (Lamb, 1972), as well as the Muller Classification in 1977 (Yarnal, 1993). With increasing concern for environmental issues during the late 1970s and 1980s, the field of synoptic climatology expanded from surface weather and began using pattern recognition to address a variety of environmental problems such as pollution, crop yields and stress, water quality and quantity as well as pollen density (Kalkstein and Corrigan, 1986). Shifts toward conceptual model building also emerged during this time, as the synoptic climatology approach was applied not only to environmental issues, but to weather phenomenon such as severe weather, tornado outbreaks, snowfall and precipitation (Leathers and Ellis, 1996; Ryan, 2009; Davis and Rogers, 1992; Leathers, 1993; Bentley et al., 2000; Cohen, 2010).

In recent years, however, synoptic climatology has rapidly turned towards regional or dynamic climate modeling, as the field links the cascades from global to local. Regional climate modeling combines the two methods as well as the two definitions, making it an attractive area of study for researchers in the field. It allows researchers to discover the physical mechanisms that couple the dynamics between the large-scale atmospheric circulation and the surface environment (Yarnal et al*.,* 2002). Additionally, researchers have also developed methods of analyzing historic synoptic-scale features through the use of tree-rings. The field, known as synoptic-dendroclimatology, has conducted a number of studies such as examining flooding and storm tracks in India. Analyzing tree ring growth and the circulation patterns during flood years revealed the trajectory of storms over the region. (Hirschboeck et al*.,* 1996).

Going forward, the field of synoptic climatology should continue to focus on process-based understanding and providing the climatic basis to improve our ability to predict future event variability. Synoptic climatology's unique ability to simplify an array of complex meteorological variables through categorization into discrete groups makes it an attractive field for research. In its simplest form, science attempts to impose order and consistency in the world by classifying complex information and revealing patterns often hidden in the raw data (Abler et al., 1971). Classification is a necessity of scientific research, as Abler, Adams and Gould state (1971), "If every object and event in the world were taken as distinct and unique – a thing in itself unrelated to anything else – our perception of the world would disintegrate into complete meaninglessness." The statement holds especially true for synoptic climatology, as the core of synoptic climatological research attempts to find patterns and relations between surface phenomenon and the constantly changing atmospheric circulation.

At the very core of synoptic climatology lie many geographic concepts. Synoptic climatology, like geography, shares an interest for the study of space, time and the surface environment. The breadth of geography allows researchers to analyze phenomena on various spatial scales and cross-scale dynamics (Harman and Winkler, 1991). With its local, regional, and global components, synoptic climatology is linked with spatial analysis. The hierarchical and temporal complexity of climates appeals to the integrative and synthetic nature of geography. As a result, geography remains an ideal perspective/discipline for climatic studies with a synoptic theme.

2.4 Methods of Synoptic Climatology:

According to Yarnal (1993), two general methodological approaches exist within a synoptic climatology study: (a) circulation-to-environment or (b) environment-to-circulation. The former approach first classifies the atmospheric circulation, and then aspects of the surface environment are related to the identified circulation patterns after the classification process. Classifying atmospheric circulation therefore does not depend on a specific characteristic of the surface environment. Using this study as an example, the classification would identify common synoptic patterns and then the researcher would see which patterns occurred in coincidence with nocturnal rainfall development. The events would then be classified into the corresponding synoptic patterns to which they apply.

Conversely, in the latter approach the character of the surface environment influences the classification process. Atmospheric circulation patterns are classified for those times when a particular surface phenomenon occurs. Using this research once again as an example, the first step would be to identify all nocturnal events. Following their identification, atmospheric circulation patterns related to each event would be classified.

Yarnal (1993) also addressed the many classification methods used in synoptic climatological research. Manual synoptic patterns, Eigenvector-based synoptic patterns and map patterns, correlation-based map patterns, and compositing are all used in synoptic climatology. Manual classification involves classifying synoptic maps into categories or patterns and was one of the first methods used methods in the field. The technique remains popular among researchers, as it allows the researcher to become aware of and familiar with the links between the synoptic and surface environments. The researcher can tailor the classification to their data, and can control the entire classification process (Yarnal, 1993). The pitfall of manual classification, however, lies in the incredible amount of time and labor required. Additionally, given the subjective nature of the technique, it is difficult to replicate since the classification process often differs among researchers (Barry and Carleton, 2001).

In the past two decades, however, synoptic climatology has seen a major implementation of computers and large data sets for correlation-based, eigenvector-based and compositing methods. Correlation and eigenvector-based methods rely on algorithms that can determine the synoptic map patterns and sort the data into the patterns. Computer-assisted methods are grounded in mathematical and statistical reasoning. Naturally, the objective nature of eigenvector and correlation analysis made it appealing to researchers. Furthermore, computer-assisted methods are much easier to perform and much more time efficient compared to manual classification and compositing. Compositing involves averaging the synoptic map patterns that correspond to a particular surface phenomenon. Yarnal (1993), however; cautions users of the method, stating that while it provides a decent initial look at the relations between the surface and the associated atmospheric circulation, compositing often produces a false impression of the normal synoptic pattern. As a result, critical synoptic patterns that document the variability and extremes within a dataset may be omitted due to smoothing of the data. Ultimately, the user must decide which synoptic analysis/classification tool should be used in the research. Though some methods tend to be more objective than others, the nature of synoptic climatology requires subjective decisions to be made throughout the research (Johnston 1968).

2.5 Low-Level Jet Research:

The term low-level jet (LLJ) refers to any low-level speed maximum in the vertical wind profile. Low-level jets form in number of environments, vary in temperature and moisture properties, and are known to contribute to extremes in meteorological phenomena. Low-level jets have been found to occur in many areas of the United States and on every continent (Stensrud, 1996; Banta et al*.,* 2002) (Figure 2.01).

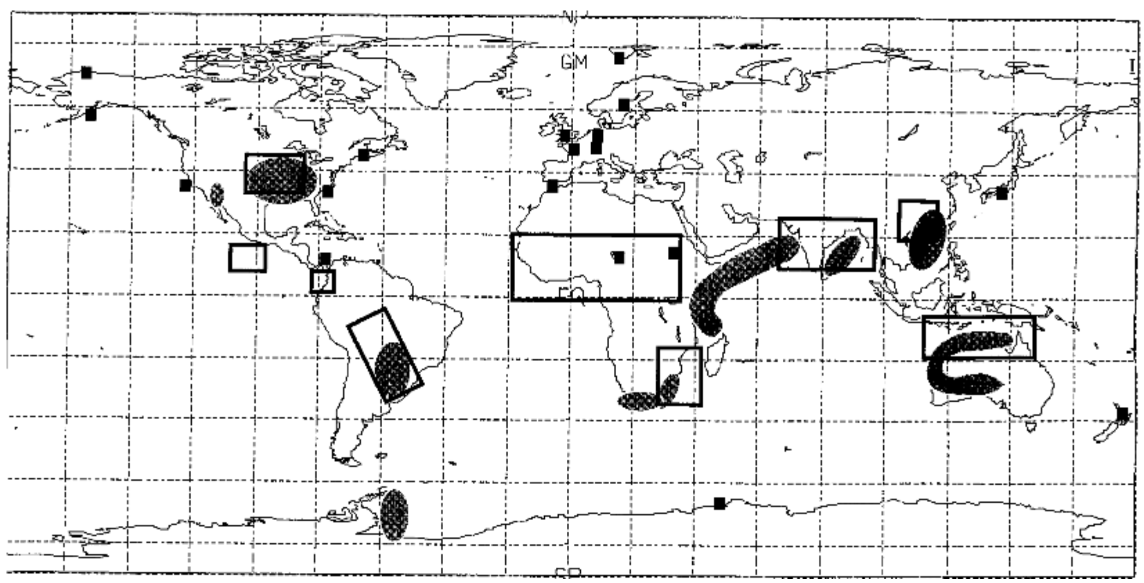


Figure 2.01: Global occurrence of low-level jets. Shaded indicates where LLJs are known or suspected to occur regularly, Whereas the open boxes indicate where MCCs are known to occur frequently during the summer months (Reproduced from Strensrud, 1996).

Two types of low-level jets (LLJ) have been identified: a synoptic form where the jet forms out ahead of a wave cyclone moving eastward across the Great Plains, and the nocturnal form where the jet develops on a diurnal cycle based on changing thermal properties of the atmosphere. The nocturnal low-level jet has been characterized by a strong near surface southerly wind maximum that transports warm and moist air from the Gulf of Mexico northward. The advection of moisture and warmer temperatures from the Gulf promotes instability and near-surface convergence needed for vertical motion and a release of the instability (Means, 1952).

Atmospheric circulation, and subsequently precipitation during the late spring-early summer precipitation maximum profoundly influences agriculture and water resources in the central United States. Circulation patterns, both on a regional and larger-synoptic scale, can greatly influence the regional hydroclimate and precipitation distribution by modulating the position and strength of storm tracks. In addition, the diurnal cycle of precipitation frequency and intensity has large effects on surface hydrology (runoff, evaporation). The diurnal variations modulate surface temperatures, and closely correlate with diurnal cycles of convection and cloudiness that can affect both solar and long-wave radiation (Aiguo et al., 1999).

Research analyzing the origins and development mechanisms of regional-to-continental variability of warm season rainfall exists throughout the literature. Particularly after the 1988 drought that affected much of the continental U.S, studies analyzing the diurnal variability of precipitation in the contiguous United States greatly increased (Ruiz-Barradas and Nigam, 2004). Similarly, many studies have examined the diurnal variability of precipitation regimes throughout North America, though the focus in recent years has been to analyze how well global and regional climate models can capture the diurnal variability of rainfall in the region (Pitchford and London, 1962; Wallace, 1975; Schwartz and Bosart, 1979; Balling, 1985; Tucker, 1993; Higgins et al., 1996). Much of the research displays a distinct geographic pattern of precipitation variations during the summer, characterized by a strong midnight to early morning maximum throughout the Great Plains.

Not surprisingly, the many types of jets and their various locations both vertically and horizontally lead to a bit of ambiguity. Regardless, low-level jets typically form within the planetary boundary layer either within or in the absence of continental-scale synoptic disturbances. In some cases, synoptically driven LLJs can form which extend beyond the boundary layer and help drive convection (Winkler and Walters, 2002). In terms of scale, LLJs tend to be considered meso-alpha (or subsynoptic) circulation features that can be as wide as several hundred kilometers and stretch for over 1,000 kilometers (Orlanski 1975). The spatial extent of a specific low-level jet will largely depend on the mechanism forcing it.

Aside from the LLJs of the Great Plains, of particular interest in recent years has been the study of atmospheric rivers (AR) associated with both low-level and mid-level jet streams. Atmospheric rivers tend to be a narrow corridor of enhanced water vapor transport often found in the warm sector of extratropical cyclones that form over oceans or gather moisture from oceanic regions (White et al*.,* 2009). ARs are important for accomplishing the needed north-south moisture and energy transport within the Earth system (they are responsible for 95% of the global meridional moisture transport). ARs have resulted in some of the largest historical storms on the west coast of the United States. For many California rivers, essentially all historical floods have been associated with atmospheric river storms (White et al., 2009; Dettinger, 2011).

2.6 Background on LLJ Conceptual Development:

Research on low-level jets first began in the 1930s, when Wagner (1939) explained the diurnal oscillation of the wind in the Great Plains as a result of diurnal variations in pressure and temperature fields. Wagner believed the oscillations to be a result of circulation patterns between the continents and ocean, the mountains and the plains, as well as between dry regions and surrounding areas. Though early research on low-level wind maxima began in Africa; significant research on the topic increased in the 1950s when researchers discovered a low-level wind maximum in the central Great Plains. Pilots in Africa in the 1930s and 1940s reported moderate turbulence of the aircraft from strong winds in the low-levels of the atmosphere in eastern Africa. It was eventually discovered in the 1960s from balloon observations and aircraft flights over eastern Kenya that strong low-level winds was the component of large-scale cross equatorial flow during the Northern Hemisphere summer known as the "East African low-level jet"(Stensrud 1996; Rao et al., 1979).

The Great Plains LLJ has been a frequently researched phenomenon since its discovery in the 1950s. Means (1952) first charted a cross section between Albuquerque, NM and Little Rock, AR of a southerly LLJ that occurred during the period of 10-12 July 1951. His study first coined the term "low-level jet." Theories on the jet's origin emerged during the same time when Bleeker and Andre (1951) suggested LLJs form as a result of large-scale drainage winds caused by radiational cooling at night along the slopes of the Rocky and Appalachian Mountains. Blackadar (1957) refuted these claims by arguing jet profiles arise from an inertial oscillation of the ageostrophic wind vector as winds at the top of the planetary boundary later become detached from the air below by the forming of a nocturnal inversion. Additionally, Blackadar (1957) suggested that the LLJ was not confined the 850 mb level and could be found during any season throughout the entire United States. His research suggested that the night-time hours provided the most conducive environment for LLJ development.

Early low-level jet theories only focused on boundary layer friction and topographical thermal gradient forcing, however, Wexler (1961) took a different approach and looked at the Great PLLJ (GPLLJ) from a larger perspective. Wexler further contributed to the origin theories of the LLJ, stating the easterly winds originating out of the Gulf of Mexico turn northward as a result of the Rocky Mountains. Strong anti-cyclonic shear, according to Wexler, offsets the Coriolis parameter. The westward boundary of the jet, with a more upslope terrain, retards the airflow by friction. As a result, a narrow corridor of wind just above the planetary boundary layer develops in the southern and central Great Plains. Wexler was the first to assert that the Atlantic subtropical high greatly influenced the GPLLJ, as the northward turning of the trade winds interacts with the Rocky Mountains. His explanations of jet formation differed from Blackadar's, as Blackadar believed the GPLJJ to only be purely a nighttime phenomenon.

In his discussion of LLJ formation in the absence of an extratropical cyclone, Holton (1967) explained the formation of the LLJ resulted from a sloping terrain from the eastern Great Plains towards the Rocky Mountains. The orographic effect results in a reversal of temperature gradients throughout the entire day. For instance, in the daytime scenario, the surface and air above warm faster in the eastern Great Plains creating an east-west temperature gradient and creating a negative pressure gradient force (PGF) between relatively lower pressures in the east compared to relatively higher pressure in the west. The daytime scenario creates a northerly component of winds over the Great Plains. Coupled with turbulent mixing, a LLJ is frequently absent during the daytime. However, when nighttime comes, because of the different sun setting times, temperatures in the east become relatively cooler than the west. Enhanced radiational cooling in the western Great Plains also plays a role. The relatively higher levels of humidity in the eastern Great Plains traps much more outgoing long wave radiation compared to areas out west. As a result, relatively higher pressure dominates to the east Whereas warmer temperatures and lower pressure values prevail near the Rockies, resulting in an east-to-west pressure gradient force. Air flow reverses during the nighttime hours and more southerly flow persists in the central Great Plains. Since a stable boundary layer forms during the evening, friction is minimal above the layer, therefore allowing an acceleration of the geostrophic winds and creating a southerly nocturnal jet.

Bonner (1968) tested the various hypotheses on LLJ formation by developing the first climatology using radiosonde observations covering the period of January 1959 to December

1960 for 47 stations through the continental U.S. With the observations provided from his study, Bonner (1968) developed the first classification scheme according to four increasingly restrictive criteria. The LLJs were classified into three overlapping groups based on magnitude of the maximum wind, while also requiring the wind speeds to decrease by a certain amount above the height of the maximum to give it "jet-like" characteristics. For the first category, the maximum wind speeds exceed 12 m/s; with the second and third categories exceeding 16 m/s and 20 m/s respectively. The winds also had to have a distinct low-level maximum, having winds decrease to a minimum speed above the jet maximum but below three km above the surface (Bonner, 1968). Additionally, Bonner (1968) displayed that in the central U.S. LLJs typically occur during the late summer and early autumn months and exhibit a strong diurnal oscillation in height, speed, and direction. Synoptic conditions most favorable for LLJ development tended to involve a strong pressure gradient across the Great Plains with northward flow from the Gulf of Mexico as well (Bonner 1968).

Whereas early research on the formation of the GPLLJ linked the phenomenon with boundary layer processes, Uccellini and Johnson (1979) and Uccellini (1980) discovered LLJs can also form under the influence of upper-level synoptic waves and generate stronger convection. The proper positioning of the LLJ along with a strong upper-level jet stream can enhance upward motion throughout the troposphere and result in deep convection In general, the intersection of the two jet streams tends to be the most favorable region for rising motions in the atmosphere. Uccellini and Johnson (1979) additionally discovered that the warm, moist lower tropospheric air associated with the LLJ, along with the cool, dry air at the higher levels of the atmosphere produce patterns of low and upper tropospheric advection that act to destabilize the atmosphere and create the environment for severe weather outbreaks. LLJs are often embedded within the poleward lower tropospheric arm of an upper-level disturbance in the exit region of the jet streak. The short-wave disturbance forms in response to upper-level divergence in the exit region of the jet streak, which causes pressure falls in the lower levels of the atmosphere. As a result, low-level airflow accelerates towards the lower tropospheric pressure center, causing a LLJ to form. The warm conveyor belt of a mid-latitude cyclone has also been found to contain LLJs. These wind maxima align with the comma of a low-pressure system and are located close to the surface frontal boundary (Winkler and Walters, 2008). Strong horizontal pressure gradients have been shown to be an important forcing mechanism for the development of LLJs in the synoptic environment.

Mitchell et al. (1995) used hourly observations from wind profiles during the warm season months of 1991 and 1992 to develop a new climatology of the LLJ. His findings showed the timing of LLJ maximum frequency tends to occur during September, in part because of the favorable synoptic patterns in place as well as an enhanced boundary layer forcing. Generally, maximum overall frequency of LLJ occurs in the southern part of the Great Plains, Whereas the maximum frequency of the stronger LLJs extended north and east into Kansas and Nebraska. The strongest jets are six times more likely to occur within a few hours of local midnight than during the day. Similar to Uccellini (1979), Mitchell et al. (1995) also found that the warm sector of extratropical cyclones promotes LLJ formation. On the other hand, subtropical ridges or southeastward moving polar highs tend to hinder its development.

Winkler and Walters (2001a and 2001b) perhaps provide the most comprehensive climatology of LLJ characteristics. The authors analyzed the airflow characteristics of low-level wind maxima in the Great Plains, the thermodynamic environments in which LLJs form (boundary layer-forced or synoptically-driven southerly LLJs), convergence fields as they relates to different airflow patterns, and as well as the cloud-to-ground lightning activity. Their analysis used upper air data from the National Climatic Data Center (NCDC) for eight rawinsonde stations in the central United States during the warm season months of April-September of 1991 and 1992. Rawinsonde stations collect observations twice daily, therefore the authors gathered wind speed and direction data for each station at 0z (1800LST) and 12z (0600LST). Naturally, this creates limitations in LLJ studies, as the two observations a day does not provide a comprehensive depiction of the diurnal cycle of the LLJ. Additionally, twice-a-day observations relegated to 0z (1800LST) and 12z (0600LST) more than likely misses a number of key times when peak LLJ winds occur. The distance between stations raises another issue in Walters and Winkler (2001) study. Some stations were separated by as much as 400km, causing a horizontal generalization of LLJ events. Regardless, the authors believed rawinsonde observations were the most reliable source of data for LLJ studies.

Like Bonner and other studies, Winkler and Walters (2001a) defined their own criteria for jet events, which included a wind speed greater than 8 m/s (17.9 mph) at or below 700 mb (millibars), or roughly 3000m up, a vertical wind shear between the level of stronger winds and the earth's surface which exceeded 4 m/s (8.9 mph), and finally a vertical shear between the level of strongest winds and either the next highest wind minimum or 550mb, whichever was the lowest elevation, that equaled or exceeded 4 m/s (8.9 mph). The authors used a compositing approach to analyze the streamlines, wind speeds, temperature, and moisture fields for each LLJ type using both pressure fields and isentropic surfaces.

Walters and Winkler (2001a) only retained southerly LLJ profiles with wind azimuths between 90˚ and 270˚, and then plotted the streamlines and isotachs of each LLJ on the pressure surface that corresponded to the vertical location of the strongest wind speed. Of the 260 LLJ events documented during the 1991-1992 warm seasons, their research found that the spatial configuration of southerly low-level jets varied greatly; leading to a classification of twelve distinct jet types based on latitudinal extent, streamline curvature, as well as orientation of deformation and confluence zones (Figure 2.02).

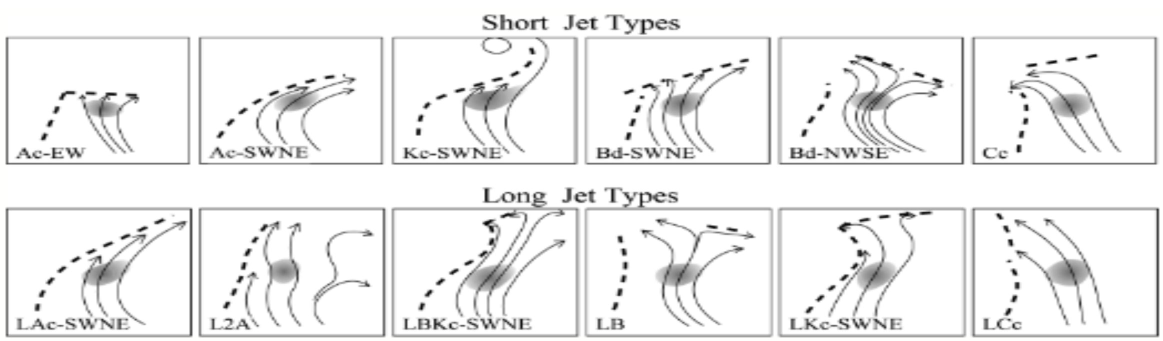


Figure 2.02: Schematic of the characteristic airflow of the twelve jet types. The solid lines represent streamlines, whereas location of significant confluence and/or deformation is indicated with heavy dashed lines. The shaded areas show the location of the jet cores. (Reproduced from Walters and Winkler (2001).

From their study, Walters and Winkler (2001a) found that over 95% of all jet events were associated with cloud-to-ground lightning and tended to be strongly related to the areas of strong convergence. Areas of the strongest convergence tended to occur in the elongated region of the confluence zone along the west flank of the jet axis in the lower levels of the atmosphere. Vertically higher LLJs, on the other hand, demonstrate convergence in areas downstream of a jet axis. Surprisingly, the temperature and moisture profiles did not vary greatly despite the vastly different LLJ airflow configurations, which the authors hypothesize may be the result of the distinct Great Plains geography. A very noticeable feature of the thermodynamic composites was the tongue of warm air co-located with a distinct moisture gradient to the southwest of the jet axis. The pattern the authors noticed correlates well with previous studies which found that the preferred location for development of southerly LLJs tends to occur on the east side of warm air advection from the southwestern U.S. Though the authors observed a wide spectrum of temporal, geographical, and vertical characteristics in the southerly LLJs, ultimately; whether the jets were synoptically or baroclinically driven remains a difficult question to answer. Many times LLJs form as the result of both mechanisms, making it difficult to pinpoint the more important culprit. Regardless, the Walters and Winkler (2001a and 2001b) study of southerly LLJs greatly increased the climatological knowledge of LLJs.

Winkler and Walters (2008) once again greatly expanded on the climatological research of the LLJ by developing a 40-year (1961-2000) climatology of southerly and northerly LLJs using daily rawinsonde observations from 36 stations in the central U.S. Using 7 regions within in the central U.S, the authors found low-level jet characteristics vary in shape and extent by season indicative of varying forcing mechanisms. The climatology developed by Winkler and Walters greatly enhanced the temporal and spatial understanding of the low-level jet by providing a more thorough description of the frequency, direction, wind speeds, and the elevation on annual, seasonal, and monthly time scales (Winkler and Walters, 2008).

The importance of the LLJ to convection in the Great Plains has been the subject of a number of studies. In fact, certain studies have discovered that moisture influx and convergence resulting from the LLJ are more important than upper-level winds to the development and maintenance of severe storm storms and heavy rainfall events in the summer (Wilson and Schreiber 1986). After analyzing 653 convective storms over a 5000 km² area in the Great Plains for the summer of season of 1984, Wilson and Schreiber (1986) concluded that nearly 80% of all storms initiated within a region of boundary-layer or low-level convergence lines. Additionally, early research showed LLJs formed out ahead of an upper-level trough, resulting in frequent nocturnal thunderstorm activity in the central United States (Means 1952, Bonner 1968). The low-level jet has also been a strong component of major flood events such as the 1951 Kansas City floods and the more recent 1993 Midwest floods (Means, 1952; Means 1954). During the floods, the LLJ provided additional fuel for excessive rainfall during convective outbreaks (Means, 1952; Bell and Janowiak, 1995; Aritt et al., 1997). The seasonal distribution of rainfall in the central Great Plains also shares a relationship with the formation of low-level jets. During the summer, in excess of 25% more precipitation falls during the nighttime hours than during the daytime over a large portion of the Great Plains (Higgins et al., 1996). More recently, Wang and Chen (2009) studied the late-spring maximum of rainfall over the U.S Central Plains and the role of the low-level jet. They found that approaching baroclinic waves lead to stronger moisture convergence than what occurs under the influence of an upper-level ridge. Moreover, wave-induced jets occur most frequently during the late spring and contribute to more than 60% of the monthly rainfall total. On a more global scale, the 2010 study by Monaghan et al. *(*2010) found that nocturnal low-level jets also help fuel nocturnal precipitation in Tibet, northwest China, India, Southeast Asia, southeast China, Argentina, Namibia, Botswana, and Ethiopia.

Given its location in the lower kilometer of the atmosphere, research suggests the GPLLJ also has an important contribution to the overall atmospheric moisture budget for the region. The LLJ plays a key role in the budget by transporting almost 1/3rd of all the moisture that enters the continental U.S, with much of it being transported during the nighttime hours (Rasumusson 1967; Higgins et al., 1996; Helfand and Schubert 1995). In fact, nocturnal low-level inflow from the Gulf of Mexico increases by more than 45% of average during the summer compared to the nocturnal mean values (Higgins et al., 1996).

Yet, up until a couple decades ago, analyzing the diurnal and synoptic variability of the GPLLJ was limited. The main issue lay with the coarse temporal resolution of upper-air observations, since the radiosondes were only launched twice a day. This made capturing detailed features of the low-level jet incredibly difficult. Eventually, the study of precipitation and the GPLLJ's influence steered towards the modeling community. Over the past 20 to 25 years, the low-level jet has been modeled both on the mesoscale and with general circulation models (GCM). Modeling by Zhong et al. (1996) captured the features of a jet event, leading to other studies trying to determine important mechanisms that produce the low-level jet. General circulation models have also been able to capture the frequency of the Great Plains Low-level Jet, though only for a two-year period (Weaver, 2007). Additionally, Ting and Wang (2006) used the GCM from the Geophysical Fluid Dynamics Laboratory along with linear and non-linear wave models to examine the role of the North American topography in maintaining the Great Plains summer mean and LLJ precipitation. The GCM model indicated that the LLJ over the Great Plains is due in large part to the presence of the North American Cordillera including the Rockies and Sierras in Mexico. This nonlinear stationary wave model suggests that the trade wind associated with the southern flank of the Bermuda subtropical high is a dominant factor in determining the strength of the jet. In fact, Ting and Wang's modeling showed that the trade winds over the Caribbean may have caused an unusually strong low-level jet which led to the devastating 1993 floods in the Great Plains, Whereas a weaker than usual jet during 1988 was influential in the severity of the drought.

Weaver et al. (2008) revealed that the GPLLJ variability is also linked with large-scale upper-level height patterns over the Pacific and with surface pressure as expressed by the North Atlantic Oscillation (NAO) in the Atlantic. The Pacific height pattern was linked to tropical diabiatic heating anomalies in both the west-central basin and in the eastern Pacific sector. Though the NAO is a robust predictor of winter variability, the NAO structure of sea level pressure variability with a N-S or meridional-dipole sin the North Atlantic basin is present in other seasons. Using regressions of July NAO, Weaver et al. (2008) found that the NAO is a key component in low-level jet variability.

Although research on the Great Plains Low-level Jet has been extensive over the past 60 years, there is more to be learned from a long-term climatology of the synoptic-scale environment associated with the LLJ and nocturnal precipitation patterns over the United States during the late-spring/early-summer precipitation maximum. Moreover, studies analyzing the characteristics of the long-term relationship between nighttime rainfall in the central Great Plains and variability in the low-level jet appears to be absent from the research. Filling in these research gaps will be essential in gaining a better grasp of the various mechanisms and characteristics associated with nighttime rainfall in the central Great Plains. An improved understanding of the GPLLJ will further our ability to anticipate how climatic change will impact both LLJ processes and nocturnal rainfall patterns in the central United States.

CHAPTER THREE:
DATA AND METHODS FOR DEVELOPING A SYNOPTIC CLIMATOLOGY OF NOCTURNAL RAINFALL

"Research is formalized curiosity. It is poking and prying with a purpose."

-Zora Neal Hurston

3.1: Study Area and Time Period:

The primary objective of this research is to analyze the long-term characteristics of nighttime rainfall in Kansas and how they relates to the sub-synoptic and synoptic environments during the months of May, June, and July from 1950-2012. Topeka's Municipal Airport was selected as the primary study site because it offers consistent and reliable hourly precipitation data for the study period. Available data from 1950-2012 provides a lengthy period that enables analysis of trends in the data. Additionally, the 1950s decade offers important variations in the climatic record with two climatic anomalies worth investigating: the devastating flood in 1951 known as "The Great Flood" decimated areas of north-east Kansas, as well as the 1950s drought that ravaged much of the southern Great Plains

Accessed from the National Climatic Data Center (NCDC), hourly weather data came in 2 forms: scanned PDFs of monthly weather summaries produced for a particular station, or through recorded data from the Global Historical Climate Network (GHCN). The former provides weather data of interest for a number of variables for every station in the state, including hourly precipitation data when available. The GHCN integrates databases of daily and monthly climate summaries from surface observations around the globe. The daily observations comprise daily climate records from sources integrated and subjected to numerous quality of assurance reviews. The GHCN can be accessed through the NCDC's interactive GIS map, where over 75000 stations containing data for multiple weather variables dating as far back as 1697 can be accessed. The year of 1892 represents the earliest monthly temperature and precipitation records for Kansas, whereas the earliest hourly precipitation data begins in 1948. While accessing the GHCN data proved to be more time efficient, utilization of both methods increased the amount of quality data available for this research.

3.2 Diurnal Time Period:

Though the 1950-2012 period represents the longer time period, an additional time period for the diurnal cycle had to be included as well in order to analyze nocturnal rainfall on a day-by-day basis. Previous literature has varied quite a bit on the beginning and concluding time of a nocturnal event, therefore deciding what constitutes a nocturnal event becomes an arbitrary or subjective process. Timescales ranging anywhere from 1700 Local Standard Time (LST) the previous day, to 1000LST in the morning and anywhere in between have all been times used to analyze diurnal rainfall timing. Selecting a time window for nocturnal events becomes even more complex when analyzing these events from a mechanistic standpoint, as rainfall mechanisms rarely follow a set clock. Often times it is difficult to distinguish between daytime thunderstorms and nighttime thunderstorms, as a complex of storms that form during the late afternoon hours often persist into the nighttime hours. Regardless, distinguishing a time slot allows for comparison among rainfall events and analysis of differences within the diurnal cycle.

Based on previous research, to be considered a nocturnal event, the rainfall had to begin between the hours of 0100 and 0900LST. However, hourly precipitation records the precipitation occurring the hour before. For example, if a weather station records 0.50 inches (12.7 mm) of rain at 0100LST, it means 0.50 inches (12.7 mm) of rain fell between the hours of 0000LST and 0100LST. The rainfall data used in this analysis were recorded in English units and, as such, this thesis will report findings in English units with SI units provided in parentheses. Though storms may begin precipitating during the time period, often times the rains persist past 0900LST. Conversely, numerous events form prior to 0100LST but the majority of the rainfall occurred between the hours of 0100LST and 0900LST. Special consideration based on their intensity had to be given to outlier events not fitting the designated time period. For instance, rainfall events beginning prior to 0100LST had to have at least .25 (6.35 mm) inches of rain occur during the designated time period for them to be considered nocturnal. In addition, total rainfall for an event had to exceed 0.25 inches (6.35 mm) to be considered a nocturnal event for this analysis. Including rainfall events below 0.25 (6.35 mm) inches adds a large number of additional events where analysis of those events would create a tedious and time consuming process, perhaps with little important information gains. Moreover, rainfall events less than 0.25 inches (6.35 mm) do not contribute much to the overall rainfall budget.

3.3 Additional Stations:

The spatial and temporal aspects of rainfall vary greatly, particularly during the warm season when convection forms as a result of lower atmospheric convergence which triggers the building of a more isolated and vertically oriented rain storm as opposed to linear-structured storms associated with the passing of frontal boundaries. Therefore, using additional observation stations allows the analysis to develop results that enable a more robust understanding of the spatial and temporal characteristics of nocturnal rainfall. Given the eastward propagation of thunderstorms in the central Great Plains during the late spring and early summer months, stations to the west and southwest of Topeka offer the best opportunity to better understand the spatial characteristics of nocturnal thunderstorms. Though many stations throughout Kansas do not contain hourly recorded precipitation data dating back to the early 1950s, enough stations exist west of Topeka with hourly precipitation data spanning the entire 63 year period to make useful comparisons. The stations selected from Kansas include locations at Hays, Kanopolis Reservoir, Oakley, Concordia Blosser Municipal Airport, Arlington and Plevna (which will be referred to as Hutchinson), and Wichita Mid-Continent Airport (Figure 3.01).

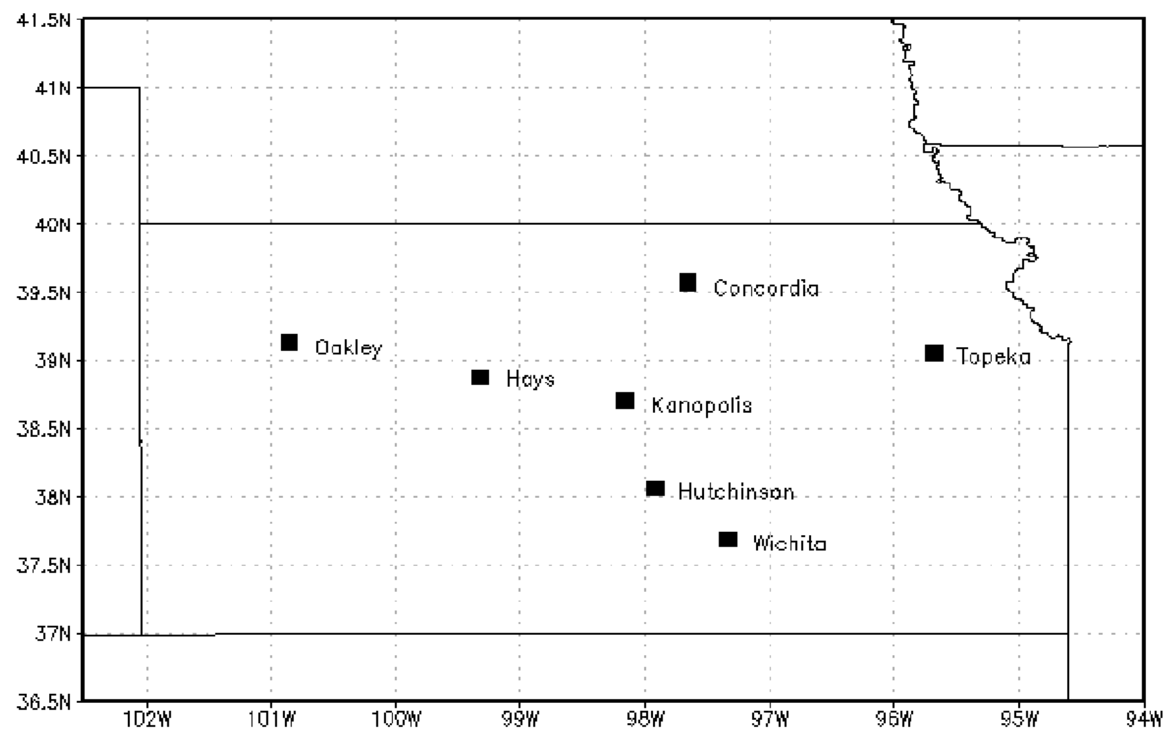


Figure 3.01: Locations of stations used to analyze hourly precipitation data from 1950-2012.

All nocturnal rainfall events recorded at Topeka were then compared with the hourly rainfall data for the same date from the 6 other stations selected. Rainfall had to exceed a total of 0.10 inches (2.54 mm) at the additional stations in order to be considered significant and related to the same event recorded at Topeka. Since Topeka represents the furthest eastward station and Kanopolis and Wichita represent the closest stations to Topeka at approximately 140 miles away, precipitation data at other stations often did not temporally match rainfall data at Topeka. Therefore, rainfall had to be recorded at or within a four-hour time window prior to the event beginning at Topeka. Moreover, stations also had a two-hour time window after the first recorded rainfall at Topeka. For instance, if Topeka first began recording rainfall at 0200LST, additional stations would need to report rain within the 2200-2300LST-0400LST time range in order to be considered the same event. Though areas to the west of Topeka generally recorded rainfall first, some events associated with passing frontal boundaries result in stations recording rainfall at the same time. Squall lines formed out ahead or behind cold fronts can stretch for hundreds of miles generally oriented southwest to northeast and often feature sections that jut out further eastward than the rest of the line. Therefore, in some instances Topeka actually reported rainfall first despite being the furthest station eastward.

With the four-hour time window, many times precipitation recorded at additional stations did not have a nighttime signature. Nevertheless, it provided a glimpse into the spatial and diurnal propagation of rainfall in Kansas. Despite the inclusion of other stations, Topeka remained the primary weather station used in selecting events for the development of the synoptic climatology. Though rainfall has considerable spatial variability, the precipitation data time series from one station for a 63 year period likely records more than enough nighttime rainfall events needed to understand both nocturnal rainfall characteristics and synoptic-scale patterns.

Precipitation data for additional stations were compiled from the GHCN along with climate summaries to increase reliability of the data. However, in some instances the data contained errors or no entries. In fact, in the past three years (2009-2012), the Hays station failed to record any hourly data. Sometimes in the absence of the data, the GHCN utilized a spatial modeling technique to fill in the missing values rather than use the recording direct observations. Stations with issues in the data included Plevna and Arlington (Hutchinson), Kanopolis Reservoir, Hays, and Oakley. In years with no recorded hourly precipitation, data from nearby weather stations (generally within a 50 mile radius) were used. In all cases, data from multiple stations had to be acquired, due in large part to the sporadic coverage of hourly weather data for many stations. Table 3.01 lists the stations and years with missing data, as well as the other stations used in the research. While rainfall data for additional stations did not include a synoptic analysis, it nevertheless showed the characteristics of nocturnal rainfall based on latitude and longitude of a station across the state of Kansas.

Stations	Years With Missing Data	Stations Used
Oakley	'55, '56, '84, '90, '93, '98, 99, '01, '03, '06	Colby, Morland, Healy
Hays	'54, '99, '04, '09-'12	Ellis, Galatia
Kanopolis	'53, '61, '70, '92, '93	Beaver, Salina
Plevna, Arlington (Hutchinson)	'67,'75-'76, '81,'83, '93, '97, '00, '12	Halstead, Bentley

Table 3.01: Stations with missing data and supplemental stations used to fill in the record.

3.4 Low-Level Wind Data:

Following the compilation of rainfall data and analysis of nocturnal events, the next portion of this research involved obtaining the needed synoptic weather data pertaining to the low-level jet for the nocturnal events. Data from 3 pressure levels, including 925 mb (~750m), 850 mb (~1450m) and 700 mb (~3000m) were used to capture the low-level wind characteristics. Low-level jets, however; contain both vertical and horizontal characteristics. In general, convective-related Great Plains low-level jets occur anywhere from west to east Texas, and up to the northern Great Plains. A 22.5˚ x 17.5˚ latitude-longitude box located over the central United States was created to analyze low-level jet characteristics (Figure 3.02). The box spans from 27.5°N-50°N, and 107.5°W-90°W, or from Texas-Mexico border up to the Canadian border and from the Four Corners region to the Mississippi River (Figure 3.02). The low-level wind characteristics were analyzed for all events exceeding 1.5 inches (37.6mm), or 86 out of 463 events. Given their anomalous nature, a rainstorm of 1.5 inches or heavier (37.6mm) raises questions about the strength of the mechanisms associated with such a heavy rainfall event, hence the reason for their selection. Moreover, 86 rainfall events exceeded 1.5 inches (37.6mm); more than enough to develop a comprehensive synoptic climatology describing the nature of the LLJ.

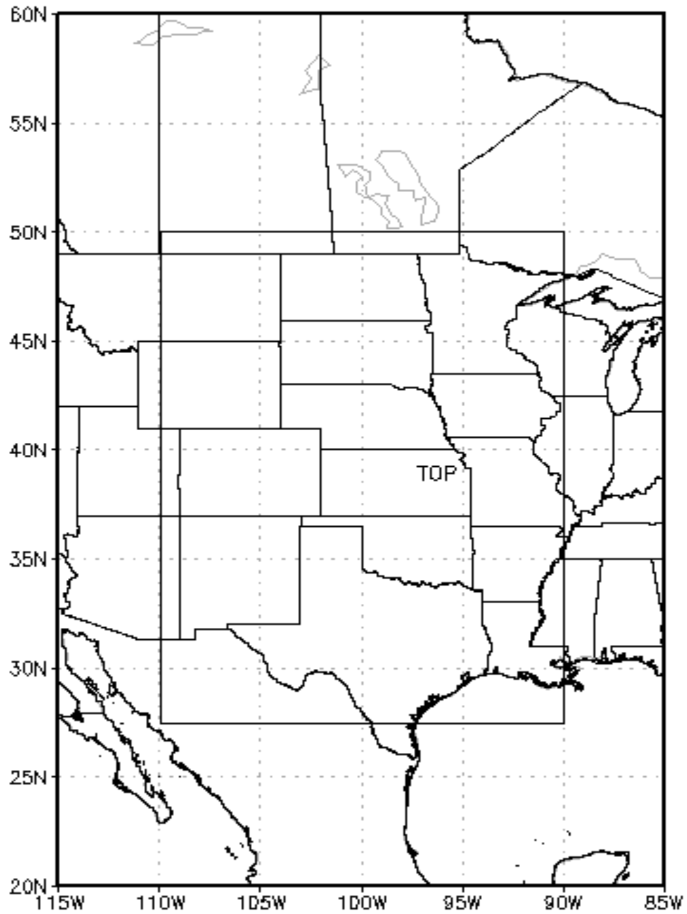


Figure 3.02: Study area used to analyze geographic LLJ characteristics indicated by the box; Topeka used a reference.

The primary research tool utilized for obtaining and analyzing synoptic maps came from the NCEP/NCAR Reanalysis Dataset. The reanalysis dataset first originated in the 1990s when the National Centers for Environmental Prediction (NCEP) and the National Center for Atmospheric Research (NCAR) created a retroactive 51 year (1948-1998) record of atmospheric fields. In order to produce an improved assessment of the entire state of the atmosphere, the joint project compiled data from the land surface, ship, rawinsonde, pibal, aircraft, satellite as well current data. The data gathered from the sources are then assimilated using a modern global data assimilation system. Data are then combined with a short-term forecast from an NCEP numerical weather prediction model using an Ensemble Kalman Filter to produce an entire state of the atmosphere. Since its inception in the1990s, the dataset has continued to grow. The first version of the reanalysis continues to be updated on a monthly basis and now stretches from 1948 to 2013 with a spatial resolution of 2.5° latitude by 2.5° longitude (roughly 172.5x172.5 miles). More recently, the second version of the dataset (V2) has been made available and spans from 1871-2010 and has a spatial resolution of 2° latitude by 2° longitude grid (Kistler et al., 1999). Global reanalysis datasets remain a vital tool in today's climate research, particularly with hydrological studies. Previous research has found the error of identification of the LLJ by the reanalysis V.1 dataset is minor (~5%) compared to conventional wind profiler and radiosonde networks (Wang and Chen, 2009). Therefore, the reanalysis datasets are a vital source for studying low-level jet characteristics and how they relate to nocturnal rainfall.

Version One (V1) of the reanalysis dataset contains various atmospheric variables that can be utilized to understand low-level jet characteristics. Regarding wind speed and direction, V1 reanalysis provides the wind components rather than the wind speed and direction. Wind components are usually expressed in terms of v (meridional wind) and u (zonal wind) in meters per second (m/s). The v wind represents the north-south component of the wind, whereas the u wind represents the east-west wind component. Positive v values indicate a southerly wind, Whereas a positive u value indicates winds blowing to the east, and vice versa for both components. Wind speed (*Speed)* and direction (*Direction*) and can be obtained from the u and v wind components using equations 1 and 2 below:

$$\boldsymbol{Speed=(u^2 * v^2)^{-2}} \quad \text{[Eq. 1]}$$

$$\boldsymbol{Direction= 57.29587 - [arctan(v/u)]} \quad \text{[Eq. 2]}$$

The conversion from the wind components to wind speed and direction were completed in Excel. The pressure levels examined include 925 mb (~550 meters), 850 mb (~1450 meters), as well as the 700 mb (~3000 meters) levels based on the latitude-longitude box in Figure 3.2. The reanalysis data produces measurements four times daily at 0z (1800LST), 6z (0000LST), 12z (0600LST), and 18z (1200LST). Given the importance of the nocturnal hours in this research, data for the sequence of hours from 0z (1800LST), 6z (0000LST), 12z (0600LST), and 18z (1200LST) were selected. Though 0z (1800LST) and 18z (1200LST) do not fit the profile of "nocturnal," they nevertheless provide a glimpse into the evolution and eventual dissipation of the nocturnal LLJ. Moreover, analyzing wind characteristics within the latitude-longitude box at 3 pressure levels in the atmosphere during the time frame creates the ability to analyze and classify low-level jets based on their temporal, vertical and horizontal characteristics.

The NCEP/NCAR data comes in the form of a NetCDF (Network Common Data Form) file. NetCDF is a set of software libraries and self-describing data formats that have the ability to create, access and share array-oriented scientific data. The project is hosted by the UniData program at the University Corporation for Atmospheric Research. A number of software programs have the ability to analyze NetCDF files, including the Grid Analysis and Display System (GrADS) software. GrADS is an interactive desktop tool that allows for the manipulation and visualization of scientific data. Data may be displayed using a variety of graphical techniques. Given the importance of wind to LLJ portions of this research, the GrADS software was used to visualize contours, streamlines, and wind vectors in association with the southerly winds and the larger synoptic environment. Streamlines and wind vectors for the southerly wind components were created and analyzed for 6z (0000LST). The time slot was chosen after analyzing the data and determining that 6z (0000LST) best represents the most common time, among the hours available, for peak jet occurrences. Visualizing and describing the low-level jet offers a workable qualitative methodological approach for a developing a synoptic classification, analyzing the raw data generated from the NCEP/NCAR Reanalysis V.1 outputs will allow for a more objective understanding into the nature of the southerly winds. Fortunately, the reanalysis modeling provides the output that can be examined using a variety of programs and statistical software. With its ease of use, ArcGIS was the primary program used to obtain the output data from the reanalysis.

3.5 *Low-level Wind Maxima Criteria:*

Similar to identifying criteria for a nocturnal event, selecting the criteria for defining a a wind maxima involves subjective decision making due to their varying nature and associated features. The methods utilized in a particular study vary greatly depending on the overall goal of the research. While some approach the phenomenon from a theoretical point of view, others emphasize its climatological characteristics, or focus on its importance to convection and moisture transport. (Blackadar, 1957; Mitchell *et al.,*1995; Walters and Winkler, 2001a). The specific goals of the researcher create different conditions for defining a low-level jet, making it problematic to utilize previous criteria as a framework. At the same time, a difficult-to-define criterion also allows the researcher to be inventive and creative which leads to new classification of LLJ categories based on their data and results. Given the encompassing nature of the Walters and Winkler's (2001a and 2001b) studies, the wind maxima criteria outlined in the research provided a useful reference to analyze southerly winds for each nocturnal event. Though, after the data had been analyzed, their categories had to be slightly tweaked to fit the goals of this research. Moreover, in their study, the authors looked at the shear components of the wind as shear give a distinct characteristic of a jet profile. This research did not analyze shear. As a result, the term "wind maxima" replaced the term"jet," as wind maxima do not require a shear component. Important to note, however, is that research wind maxima still contributes to GPLLJ research as it analyzes the configuration of the southerly wind component. Regardless, certain specifications outlined in the Walters and Winkler's study (2001a) provided a tentative framework for defining the low-level winds. Most important for this study, the data from each grid cell within the study area had to have wind directions between 90˚ and 270˚ and exceed 8 m/s (17.9mph) to be considered a wind maxima.

The first steps in analyzing the winds involved taking a grid cell south of Topeka and determine the wind characteristics at that location. Analysis of the grid cell at 37.5˚N and 97.5˚W, located south and west of Topeka (which will be referred to as Wichita, KS grid cell), provided a glimpse into the low-level wind characteristics south of the region experiencing heavy rainfall. It was assumed that if one grid cell showed the characteristics of a wind maxima, grid cells in nearby locations more than likely mimicked the Wichita grid cell. A similar study (Higgins *et* al, 1996) used the one grid cell approach to analyze the vertical profile of the low-level jet. This assumption by Higgins et al. was later proven true using a composite analysis approach. Naturally, low-level winds on a case-by-case basis more than likely vary greatly in their evolution, intensity, duration, as well as geographically and vertically throughout the nighttime hours. Therefore, creating categories based on the similarities and differences of low-level winds for rainfall events provides a more comprehensive nature into the variability of the wind maxima. The following questions were posed during analysis of the data for the Wichita grid cell, as well as the entire study area:

Wichita grid cell:

Are winds >8 m/s?

At what time? (0z, 6z, 12z, 18z)

At what level?

Do winds increase from 0z to 6z?

Do winds increase from 6z to 12z?

Do winds decrease from 0z to 12z?

Entire Study Area:

Vertical coverage (where is the jet in the vertical?)

Horizontal coverage (Where is the latitudinal and longitudinal position of the jet axis?

How do the winds veer with latitude?

How do the winds veer with time?

How do the winds veer in the vertical from 925 mb to 700 mb?

Where is the max wind (horizontally and vertically)?

Where is Topeka in relation to the jet axis?

Using Excel, the characteristics of the low-level winds were analyzed using the wind data provided for each event. Visual analysis occurred through the images produced using the GrADS software and the reanalysis data. The visual examination of the winds included where Topeka resided in relation to the wind maxima axis, the geographic coverage of the wind maxima for all three pressure levels, the streamline configuration of the winds as well as any other features that could be attributed to wind maxima and related nighttime rainfall.

Initially, the data were compared on a month-by-month basis that allowed for both a long-term climatological understanding of the southerly winds as well as an understanding of the similarities and differences on a month-to-month basis. However, based on the data analyzed for each month and each time slot, it became apparent that the wind characteristics at the Wichita grid cell contain a number of distinct characteristics that could be classified. The diverse nature of the winds led to the development of a classification scheme with five categories based on data from the Wichita grid cell. The number of levels with wind maxima characteristics provided the basis for developing the classes. Each time slot has both a wind direction and speed for each of 3 pressure levels. If a pressure level exhibited winds fitting the criteria of greater than 8 m/s (17.9 mph) and having a direction between 90˚ and 270˚, a 1 was given to that particular pressure level. Conversely, a slot received a 0 if winds did not fit the profile for that level and time. Therefore, for each time slot totals could range from as low as 0 to as high as 3. Taking the event in its entirety, totals could amount to as high as 12. A score of 12 means each level from 0z (1800LST) - 18z (1200LST) had the characteristics of a southerly low-level wind maximum. From these totals, distinct classes emerged based on the wind characteristics during heavy rainfall events at Topeka.

3.6 Larger-Scale Synoptic Assessment:

An additional chapter discusses the various atmospheric mechanisms that generate nighttime convection in northeast Kansas. Analysis includes a discussion of the upper-level features most common to nighttime convection for May, June and July. However, the three-month time frame represents a period where rapid changes occur in the upper-level circulation. The Northern Hemisphere circulation quickly changes from its springtime patterns to the patterns typically seen during the summer during this three-month window, and the changes can be detected on a week-to-week basis. Analyzing a month in its entirety does not provide a full explanation into how the mechanisms for generating nighttime precipitation change from mid-late spring into mid-summer. As a result, the time periods analyzed were divided into half-monthly sections, with two separate sections dedicated to the first and second halves of each month. The primary dataset was once again the NCEP/NCAR reanalysis data V1. From the reanalysis, the 500 mb geopotential height patterns, coupled with the upper-tropospheric wind vectors at 200 mb were extracted to provide insight into the common modes of circulation. The 500 mb height pattern provides a representation of the upper-level westerly circulation in the Northern Hemisphere and the long wave pattern in place over the Northern Hemisphere. On a smaller scale, shortwaves or baroclinic waves moving through the long wave pattern can also be detected by analyzing a 500 mb chart. At 200 mb, however, lies the clearest picture of the upper-level jet stream. The long-wave pattern in place and position of the jet stream are both linked with the uplift needed not only for generating precipitation, but also for fueling a southerly wind maxima in the lower troposphere. The study region for this portion of the research includes parts of the contiguous United States and parts of southern Canada and much of Mexico, or the coordinates of 20˚N-60˚N, and 120˚W-85˚W (Figure 3.03). The reasoning behind expanding the study area for the synoptic section has to do with the larger-scale nature of upper-level circulation patterns. The wind maxima tends to be a feature on the sub-synoptic scale confined to the Great Plains, thus encompassing a smaller region. Upper-level circulation patterns, however; span across the entire United States and throughout the northern hemisphere. In order to fully understand the larger-scale synoptic patterns, the study area had to be expanded to include much of the North American continent.

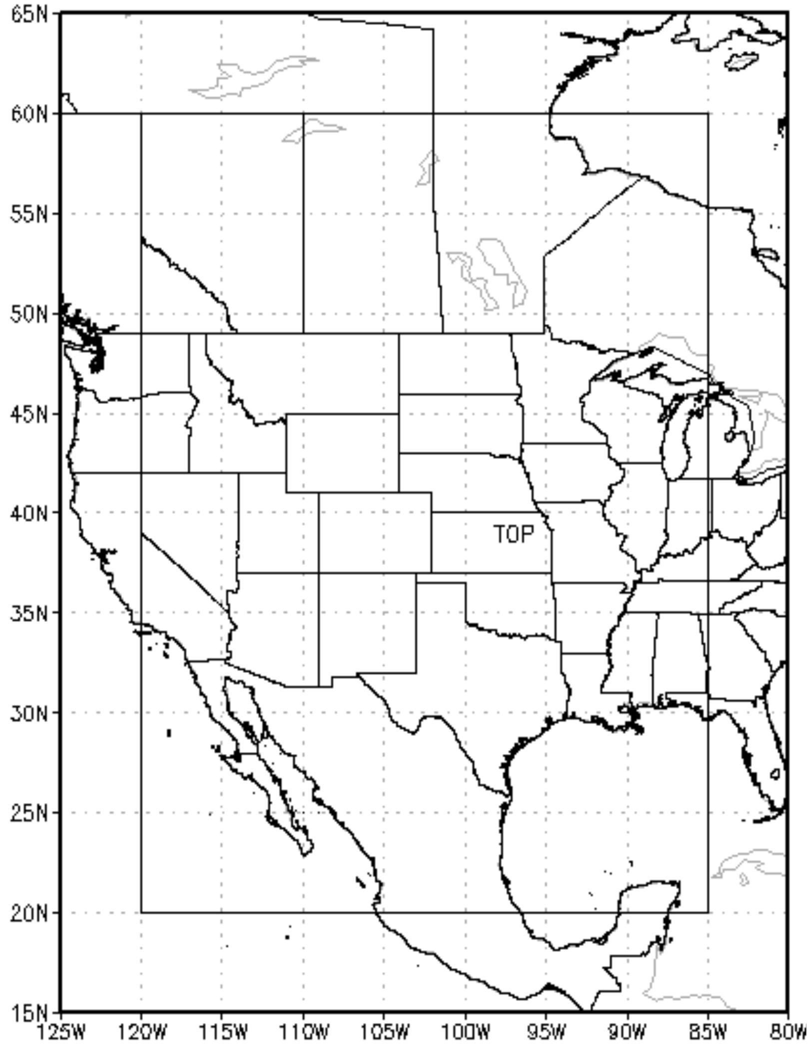


Figure 3.03: Study area in rectangular box used for larger-scale synoptic analysis. Topeka (TOP) provided as a reference point.

Within each half-month, 40 rainfall events were selected at random from the pool of all available nocturnal rain events at Topeka. The time slot of 0z (1800LST) was chosen as the time period of analysis, since in most cases it gives the best representation of the atmosphere prior to nighttime rainfall occurring at Topeka. Output generated for each event from the reanalysis dataset was examined visually, allowing for the development of a synoptic climatology categorization system best representing the most common modes of circulation. Examples of circulation categories included a trough over the western United States, and a ridge in place over the central and eastern United States. Conversely, the opposite pattern frequently showed up as well, with a dominating ridge in the western U.S and a trough in the east. In some cases the flow pattern was much more complicated than a simple trough-ridge or ridge-trough pattern. For categorical purposes, however; the patterns were simplified as much as possible. With 240 events, six categories were then developed based on the most common patterns. The categories include:

Trough-Ridge (TR)

Ridge-Trough (RT)

Closed Low (CL)

Trough (T)

Zonal (Z)

Trough-Zonal (TZ)

CHAPTER FOUR
NOCTURNAL RAINFALL CHARACTERISTICS

"*Rolling as in sleep, low thunders bring the mellow rain.*"
-- John Richard Vernon

4.1 Climatological Characteristics:

Nocturnal rainfall characteristics for May, June and July (MJJ) were analyzed using hourly precipitation data obtained from the Topeka weather station (Topeka Municipal Airport), as well as other stations located south and west of Topeka. On an annual basis, the majority of precipitation occurs during the three-month (MJJ) time span (Table1). Over 1/3rd of the annual rainfall total falls during the three-month time frame, which leads into the first research question: How much is nighttime rainfall contributing to this May, June and July precipitation maximum in Kansas? Using hourly rainfall data from Topeka and stations across Kansas, the following chapter analyzes the various characteristics of nighttime rainfall during May, June and July for the period of 1950-2012. Subsequent sections discuss the timing characteristics of nocturnal precipitation, more specifically when nighttime thunderstorms begin, how long they last, as well as how the intensity of rainfall relates to the diurnal cycle. Additionally, further analysis includes the spatial characteristics of nighttime rainfall using hourly precipitation data from additional stations.

Month	Mean	Extreme	Year
January	0.86	2.67	1973
February	1.32	3.49	1971
March	2.49	8.44	1973
April	3.53	8.69	1999
May	**4.46**	**11.81**	**1995**
June	**5.11**	**10.91**	**1977**
July	**3.98**	**12.01**	**1950**
August	4.24	11.18	1977
September	3.66	12.71	1973
October	3.03	7.24	1980
November	1.85	5.64	1998
December	1.35	4.30	1973
Annual	35.64		

Table 4.01: Topeka monthly average precipitation. May, June and July based on 1950-2013 climatology from this research. Other months based on 1981-2010 normals gathered from NCDC.

Based on the hourly precipitation data for the 63-year period, the Topeka Municipal Airport recorded 463 nighttime rainfall events of 0.25 inches (6.35 mm) or more, an average of 7.37 events in MJJ each year. The number of nocturnal rainfall events for each month did not vary greatly. Topeka recorded 154 events for May, June experienced slightly more with 156, and July had the least number of events with 153 (Table 4.02).

	Recorded nighttime events	Average nighttime rainfall (inches)	Standard Deviation	Extreme Year (highest nighttime rainfall)	# of years with no recorded nighttime rainfall
May	154	2.09	1.63	8.51 (2007)	5
June	156	2.74	2.32	10.32 (1977)	8
July	153	2.47	2.41	9.92 (1950)	6

Table 4.02: Topeka Nighttime Rainfall Characteristics for May, June and July based on the 1950-2012 period.

However, the amount of nocturnal rainfall varied greatly on a year-to-year basis. Nocturnal rainfall varied from as high as 18 inches in 1950, to as low as three inches only two years later in 1952 (Figure 4.01). By far, the back-to-back years of 1950 and 1951 represent the two wettest three-month periods largely a result of the nocturnal precipitation that occurred during the two years. Not surprisingly, this also results in the 1950s decade being the wettest decade on record despite going through the crippling 1950s drought (Figure 4.02). In fact, three out of top ten wettest nocturnal rainfall seasons occur during the 1950s. The variability displayed in the precipitation data epitomizes the Great Plains climate during the warm season, as the region often experiences considerable swings on a year-by-year, and even a month-by-month basis.

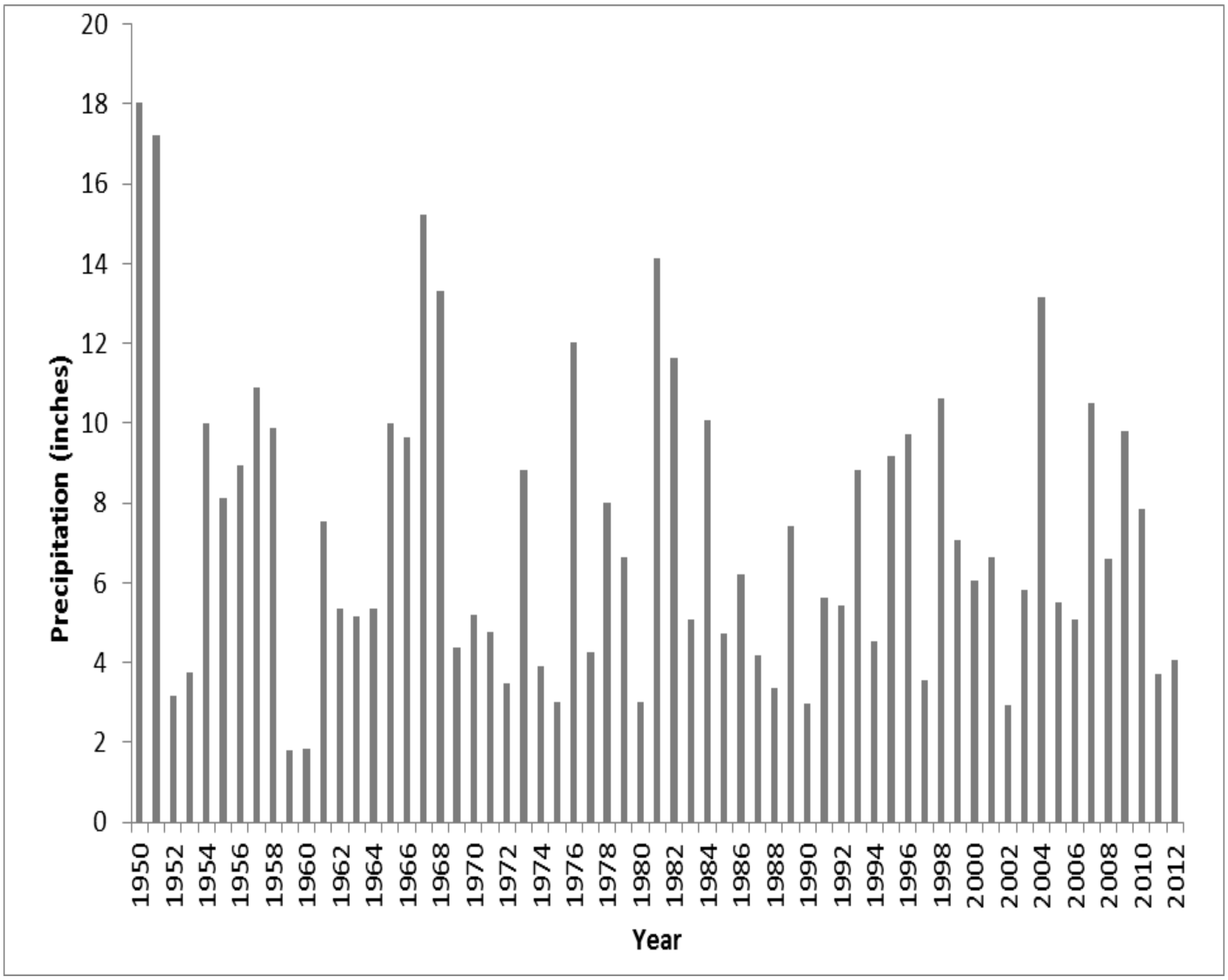


Figure 4.01: 1950-2012 nocturnal rainfall climatology for May-July based on precipitation data from the Topeka weather station.

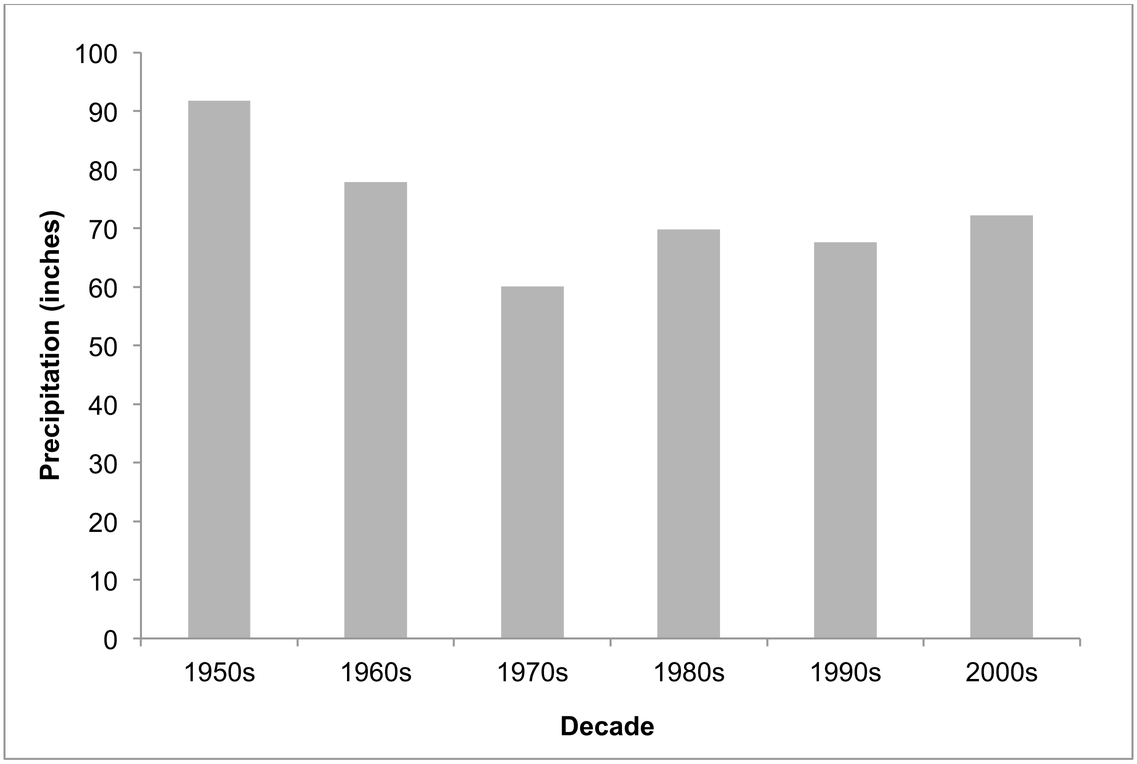


Figure 4.02: Percentage of nighttime rainfall for May, June and July for the six decade study period.

An aspect of climatology is detecting trends and patterns within a dataset. Particularly with long-term data such as precipitation and temperature, analyzing the data for trends and patterns can not only provide an understanding of the historic pattern of the data, but may also provide a projection of what the future will be. With the growing concern of water availability in the Great Plains, understanding precipitation trends, whether it is total rainfall or nocturnal rainfall, becomes of utmost importance. Though the Topeka weather station only records precipitation for one location, analyzing the patterns and potential trends within the data can be a start towards understanding nighttime precipitation in the Great Plains. Analysis of nighttime rainfall indicates that little to no trend exist for any of the 3 months, however; with the exception of July which exhibits a slight downward trend. Though, the trend more than likely results from the two incredibly wet years of 1950 and 1951. Similarly, the seasonal total does not show any significant trends (Figures 4.03-4.06).

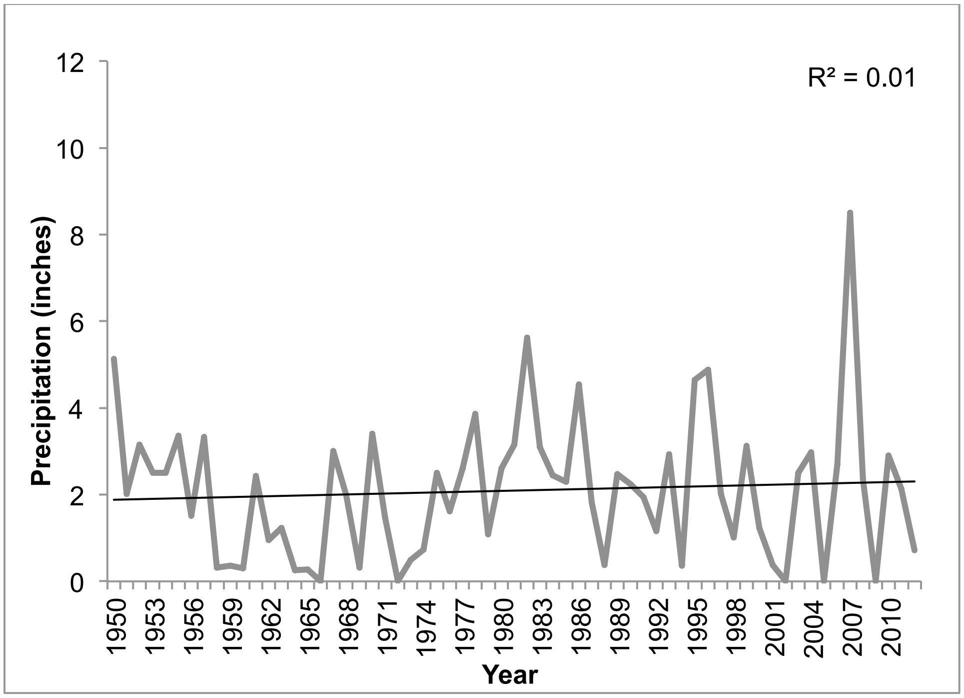


Figure 4.03: Interannual variability and long term trend in the nocturnal precipitation percentage for Topeka for May.

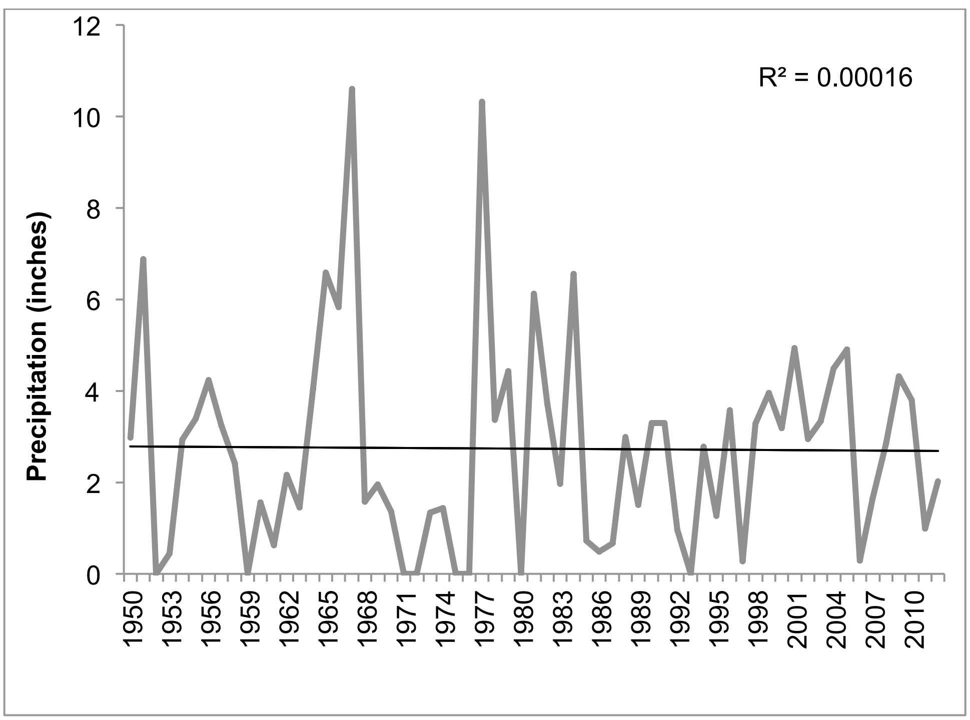


Figure 4.04: Same as 4.03 for June.

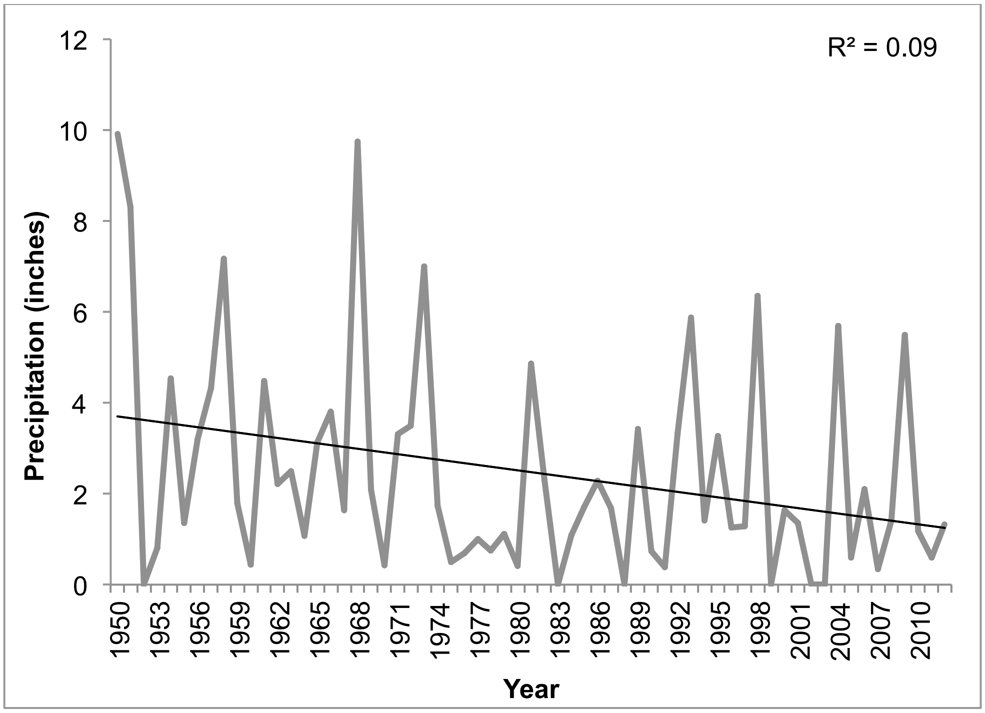


Figure 4.05: Same as 4.03 for July.

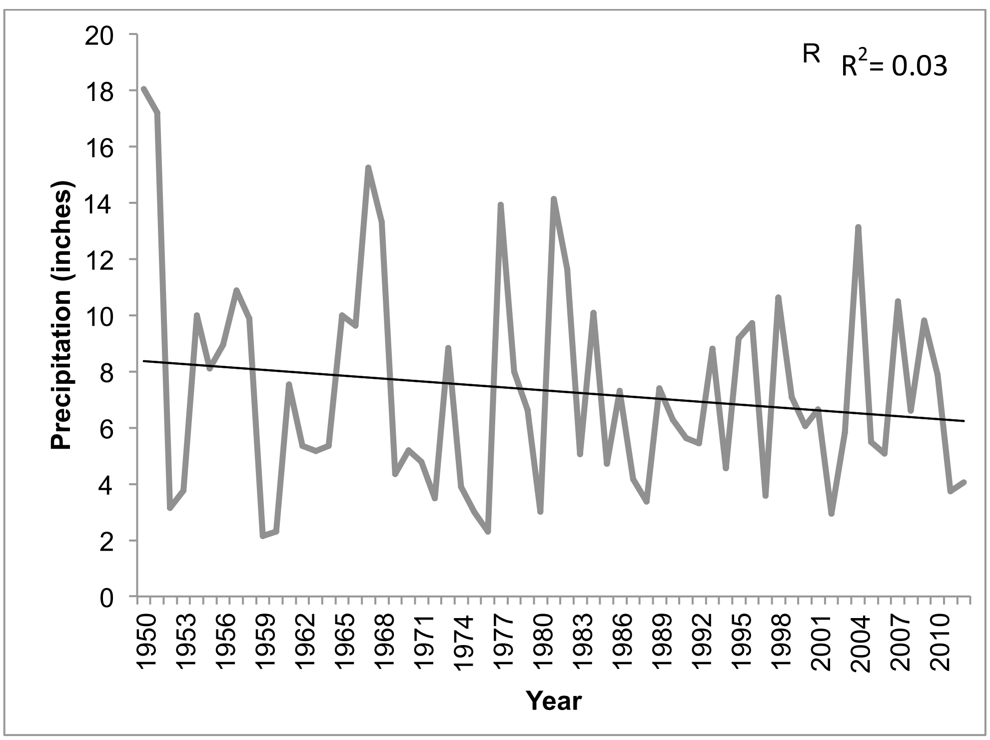


Figure 4.06: Trend of total nighttime rainfall for May, June and July.

Combining the three months, nighttime rainfall contributed an average of 7.55 inches (191.8mm) of the rainfall. The average rainfall total for MJJ is 13.55 inches (344.2mm) based on the 1950-2012 climatology for the Topeka weather station. Therefore, nighttime rainfall accounts for over half (53%) of the rainfall during the three-month time frame. However, the contribution of nighttime rainfall to the overall monthly precipitation total differs by month. For example, during the month of May, nighttime rainfall accounts for 43% (2.09 inches or 53.1 mm) of the monthly rainfall budget. Conversely, June and July have much higher percentages of 56% (2.73 inches or 69.3 mm) and 58% (2.47 inches or 62.7 mm), respectively. Despite the 3 months having relatively similar characteristics in terms of the number of events over the 63-year period, June and July have a stronger nighttime rainfall signature, suggesting a transition in the diurnal nature of rainfall for June and July. The two months also appear to have more variation in terms of monthly nocturnal rainfall totals. For May nighttime rainfall has a standard deviation of 1.63 inches (41.4 mm), whereas June and July have higher values of 2.33 inches (59.2 mm) and 2.41 inches (61.2) (Table 4.02). Increases in variability may result from synoptic influences that can vary greatly and affect the rainfall budget in Kansas during the early summer months.

A relationship also exists between the amount of nighttime rainfall a month receives and how much it contributes to the overall monthly rainfall budget, particularly during the months of June and July (Figures 4.07-4.09). Surprisingly, little to no relation can be detected for May (R-square value of only 0.07), indicating that years with anomalously wet or dry years are not correlated with the amount of nighttime precipitation. However, June and July share similar characteristics, with higher R-square values of 0.22 and 0.24 respectively. Higher and positive R-square values indicate that in years with high monthly rainfall totals, nighttime rainfall more than likely made important contributions to the overall total than during normal or drier years. This relationship is even clearer when analyzing specific anomalous wet or dry years. For instance, 1950 and 1951 were by far the wettest seasons on record, with precipitation totals (both day and night) exceeding 20 inches (no other three month period eclipsed 20 inches). During those years, 75% and 65% of the total rainfall for the 3-month period fell during the nighttime hours. Similarly, during the wet July of 1968, where Topeka recorded 10.17 inches of rain, nocturnal rainfall account for nearly 96% of three month total. Conversely, during incredibly dry years such as 1952 and 1953 (worst drought years of the 1950s), no nocturnal rainfall was recorded in 1952; Whereas in 1953 only 19% of the rainfall fell during the nighttime hours.

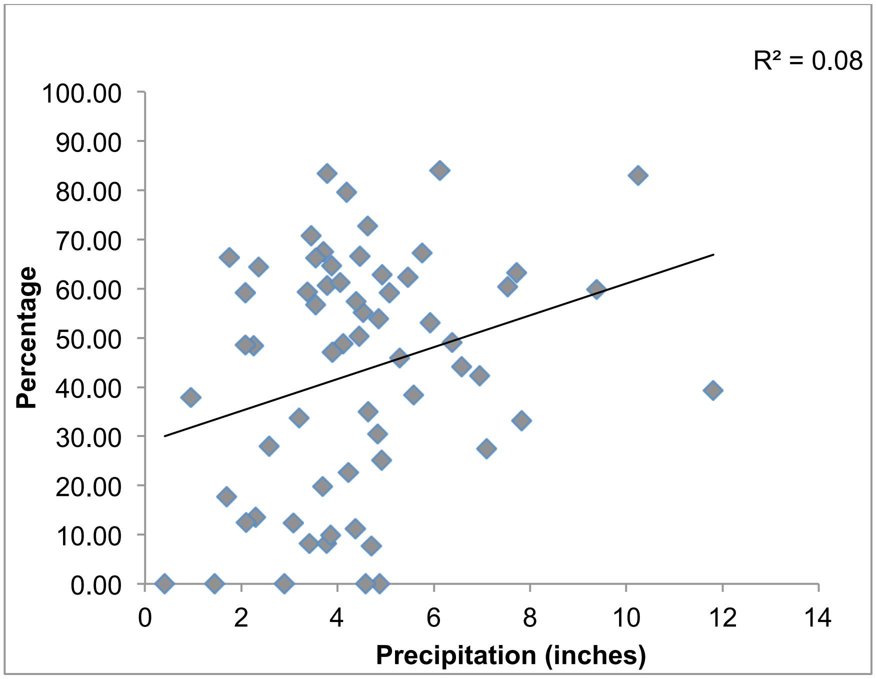


Figure 4.07: Regression model of May precipitation totals on the X-axis and the percentage of nighttime rainfall recorded during that year on the Y-axis.

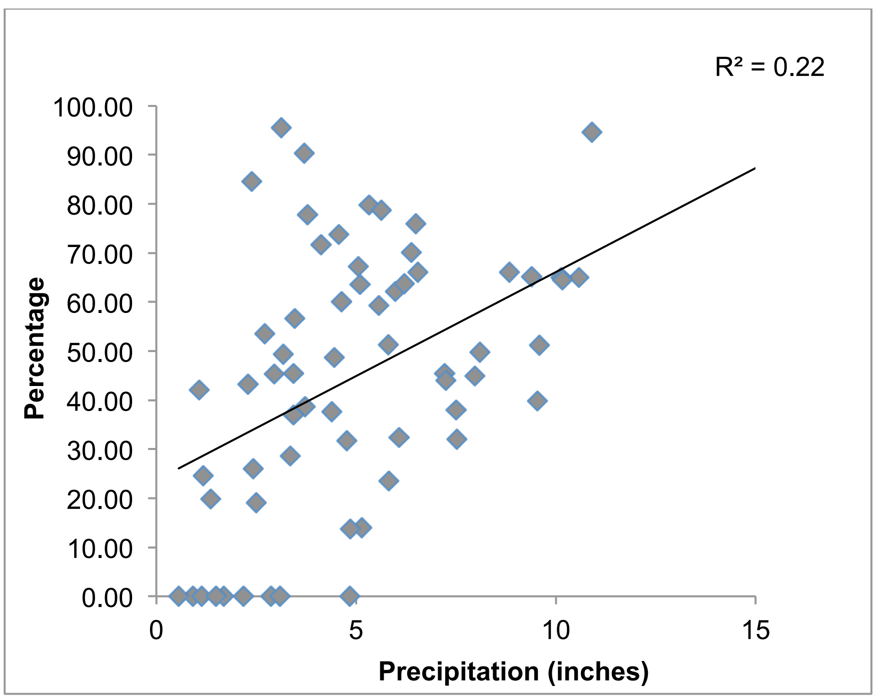


Figure 4.08: Same as Figure 4.07, only for June.

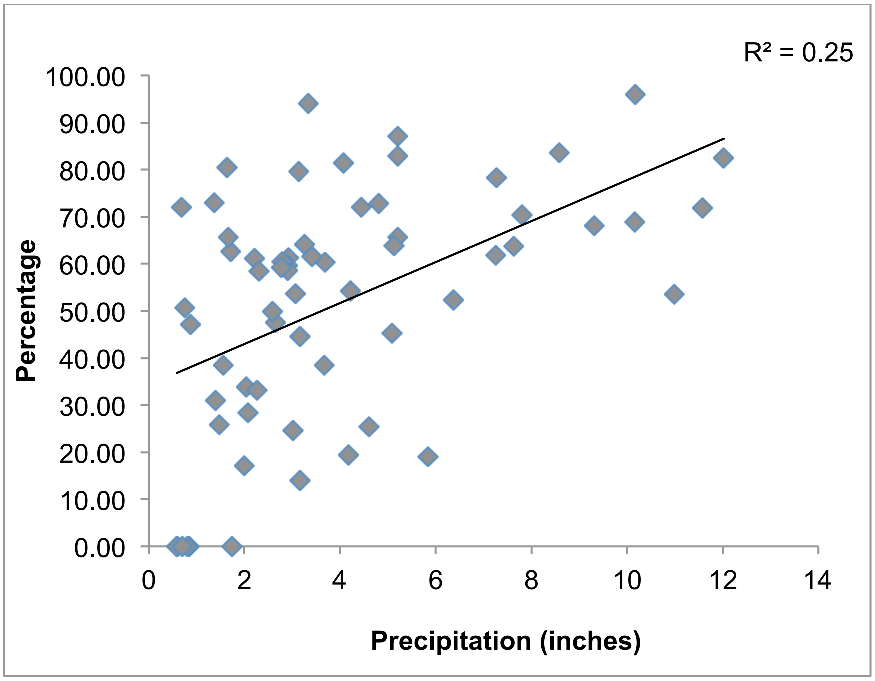


Figure 4.09: Same as Figure 4.2, only for July.

Visual analysis of Figures 4.07-4.09 suggests a possible threshold at about 7 inches (177.8 mm). In months that receive the larger rainfall amounts, the minimum percentage of nocturnal rainfall jumps to at least 30%. An important aspect to remember, however; is that Whereas the hours of record between 0100 and 0900LST represent the study period, a number of events either begin prior to 0100LST or persist past 0900LST. In fact, some events include rainfall that might not be considered nighttime at all, but given the extraordinary characteristics of the event (whether it be its intensity or duration), it was included in the data. Having said that, event start and stop times make analyzing the study period of 0100-0900LST difficult, as subjective decisions must be made regarding what constitutes a nighttime event. Of course, the same issue exists with studies of 24-hour precipitation totals. Weather events can span the time break from one observation interval to the next. Yet, this concern raises important questions about just how much a specific period of time contributes to the overall rainfall budget. For example, the hours of 0100-900LST were extracted from the dataset in order to gauge just how much the nine-hour period contributes to overall rainfall during May, June, and July. Figure 4.10 displays the year-to-year differences among the datasets for the three-month period. Naturally, the 0100-0900LST time slot makes up a smaller percentage of the overall rainfall totals for May, June and July. Yet, the nine-hour period still makes up a significant percentage of the overall rainfall totals, as 38% of rainfall occurs from 0100-0900LST. With convective precipitation facilitated by strong surface heating in late afternoon, one might expect rainfall totals during the night to contribute a smaller percentage. But 9/24ths of the day corresponds to 37.5% of the available time during the day. Nocturnal rains contribute their fair share of the daily total. When one adds in the rainfall from events that begin before 0000LST or extend beyond 0900LST, an additional 15% of the overall rainfall is included.

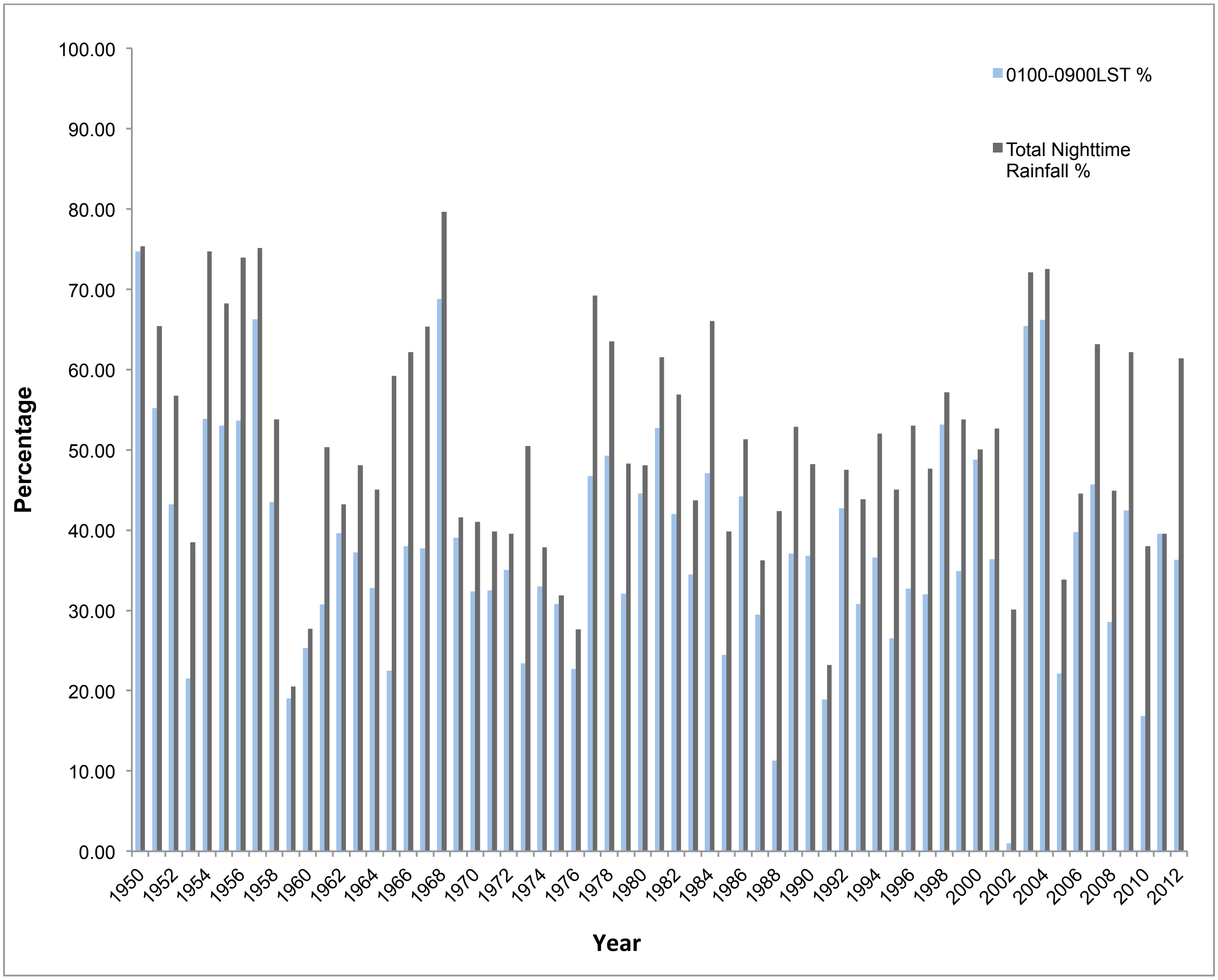


Figure 4.10: 1950-2012 nocturnal precipitation percentage recorded and the total precipitation percentage recorded only during the hours of 0100-0900LST.

4.2: Timing:

Nocturnal rainfall events have distinct start time characteristics based on the month. May, June and July all have different patterns in terms of the timing, duration and intensity of nocturnal precipitation. Analysis of the various temporal characteristics of nighttime rainfall is presented first, with an emphasis on the nature of the start times of nighttime rain and thunderstorm activity. During May, nocturnal rainfall has a distinct tendency for beginning between 0000LST and 0200LST (43% of events began during the 2 hour time frame) (Figure 4.11). An earlier start to nocturnal events compared to June and July (Figures 4.12 and 4.13) is evident. A small percentage of the events begin prior to 0000LST, largely the result of events recorded in May. May is the only month with a relatively high percentage of events beginning prior to 0000LST (19% of May events began before midnight, with roughly 11% beginning at 2200LST or 2300LST). In fact, the Topeka station recorded a number of events that started before the 2300LST (29 out of 463) time slot. Whether these rain producing systems are associated with a mid-latitude cyclone and a migratory surface front or some form of slow moving convective complex is the subject of subsequent presentation of research findings. The relationship between lengthy precipitation events and synoptic-scale forcing is presented in Chapter Six. In the morning hours from 0400LST to 0900LST, the distribution of events did not vary greatly, ranging anywhere from 5-7% per hour and making up 33% of the total start times for May (Figure 4.11). Dividing May events by the first half and second half of the month reveals more surprising facets about initiation times. Only eight of the 69 events (11.5%) occurring from May 1-15 begin prior to the local midnight hour (0000LST). In fact, only 3 events are recorded prior to 2300LST. Comparatively, 85 events are recorded for the latter half of May, with 21 initiating prior to 0000LST and 12 starting prior to 2300LST. The smaller number of events greater than .25 inches (6.5mm) and later start times at the beginning of the month indicates that the majority of events beginning in the evening occur in the latter half of the month.

Nocturnal rainfall in June (Figure 4.12), however; shows a clear tendency for events starting at 0100LST, with 23% (36 out of 156) of rainfall events beginning during that time. The hours of 0100-0300LST make up the majority of start times, as approximately 35% of all thunderstorms began during the 3-hour time frame. An interesting component of June data has to do with the steep decrease in rainfall events beginning at 0200LST, 0300LST, and 0400LST. The number of events beginning the evening before 0000LST decreases from May into June, as only 16% of all June events begins prior to midnight. Interestingly, the number of events beginning just a few hours before midnight more than doubled in June compared to May (Figure

4.11 and 4.12). June rainfall rarely began in the early evening hours and persisted well into the morning hours, suggesting that either June rainfall did not fall as often in the late afternoon or early evening hours, or if it did the mechanisms were not in place to maintain the storm into the early morning hours. While noticeable differences were present from the first half of the month to the second half of the month in May, almost no differences are present in June. Each half almost has the same number of events initiating prior to 2300LST (12 for the first half and 11 for the second half), with no stark differences in the number of events as well (81 events for the first half and 77 for the second half).

July nocturnal rainfall events are more likely to begin in the early morning hours (Figure 4.13). One event initiated at 2000LST (five hours later than the earliest record event in May), and only 10% (15 out of 153) of the total events began before 0000LST. Much like June, thunderstorms appear to randomly cluster in 1 hour over the others, as the largest percentage of events (19%) in July began at 0000LST bin. The percentage is much lower in the hour prior and hour after 0000LST, (2300LST made up only 4% of the total, whereas 0100LST only accounted for 8% of all events). Since the midnight hour accounted for nearly a 1/5 of all the events recorded for that month, a decrease occurs in percentages for the 0100-0900LST time slot. With sunrise occurring around 0500LST, it is not surprising that the frequency of storms starting later in the latter half of the analysis period declines. Only 63% of July nocturnal events began during the study time window, compared to the 73% and 72% for May and June. Reasons why July events tend to begin at the 0000LST time period will be further analyzed and discussed in the Chapter Five and Chapter Six. Taken together, the rain events during the late spring-early summer precipitation maximum clearly display a time period where the majority of events begin between the hours of 0000 and 0300LST (Figure 4.14). Also noticeable from analysis of the data is the gradual step-down of percentages from 0100LST onward. Aside from 0900LST, every hour after 0100LST decreases in the number of events beginning during that time, suggesting the key time for rainfall to move into the area occurs during a four-to-five hour time window. Similar to June, no dramatic differences can be seen between July 1-15 and July 16-31 in terms of initiation. Though, the majority of 0000LST events occur in the first half of the month (18 compared to 11 in the latter half)

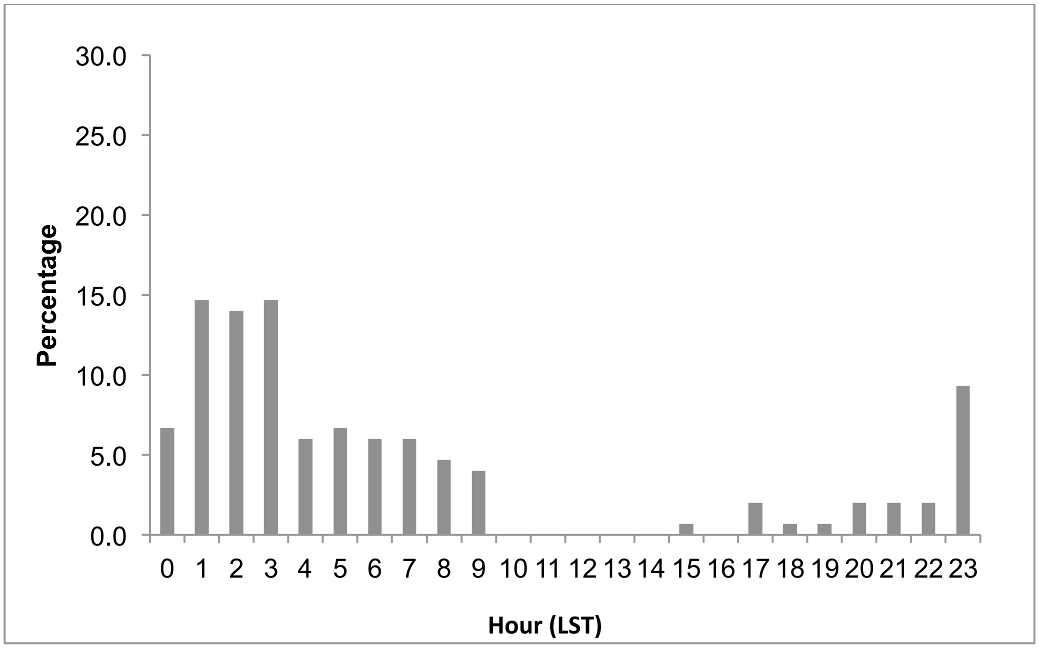


Figure 4.11: Percentage of nighttime rainfall start times for May. Values are listed as percentages.

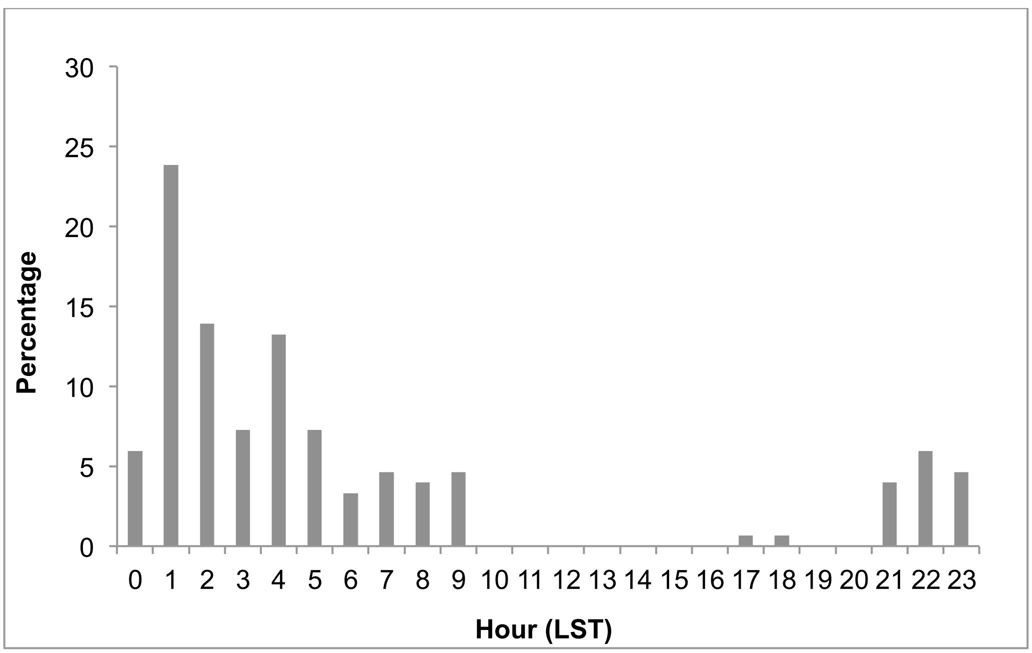


Figure 4.12: Same as 4.12 but for June. Values are listed as percentages.

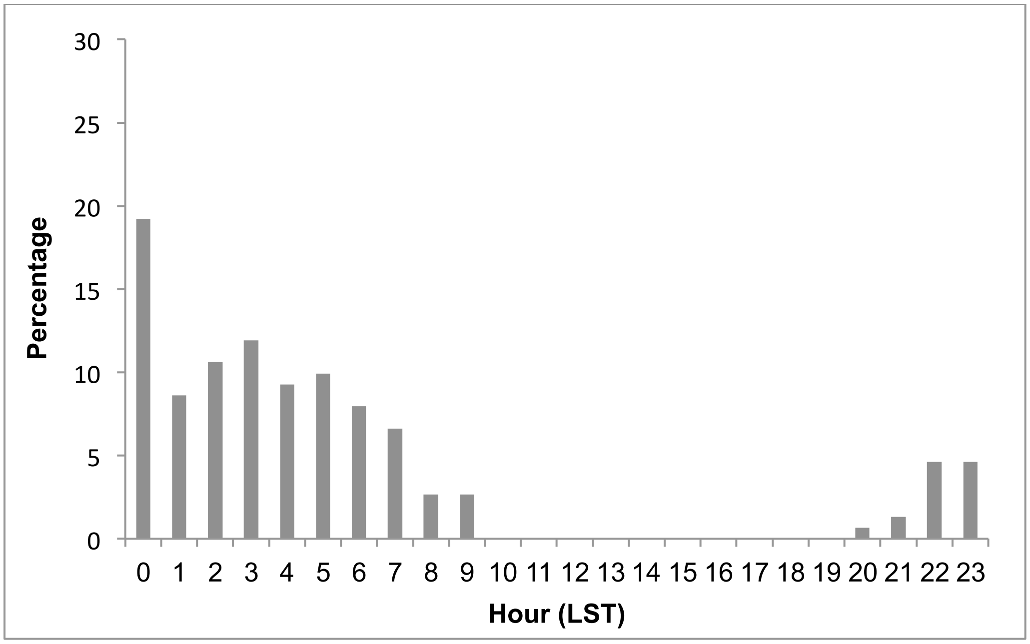
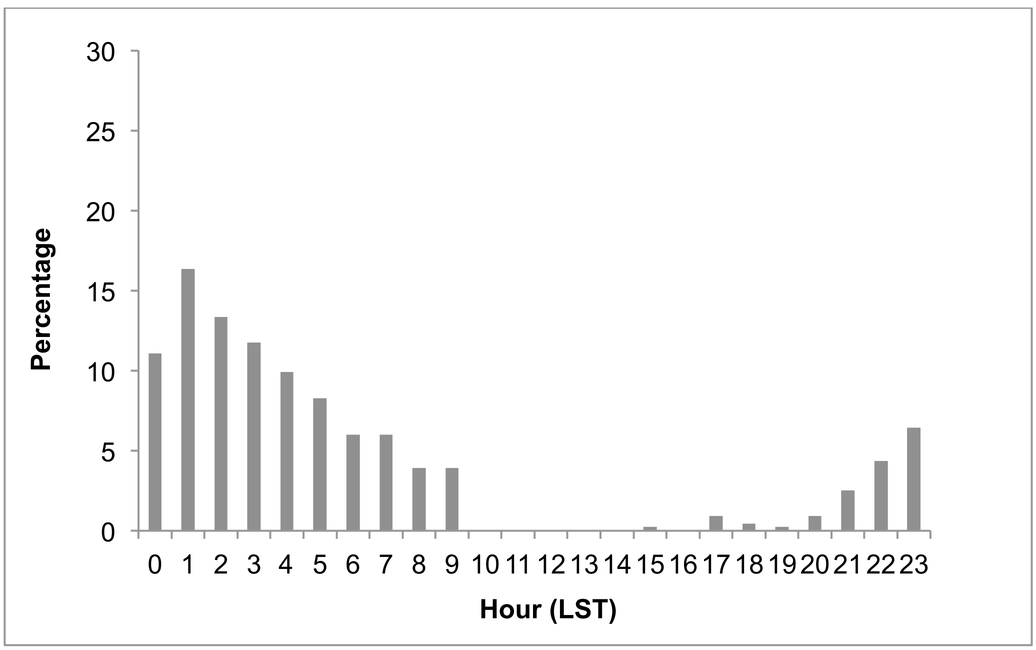


Figure 4.13: Same as Figure 4.12 but for July. Values are listed as percentages.



Figures 4.14: Same as Figure 4.12 but for the combined study period. Values are listed as percentages.

4.3 Duration and Intensity Characteristics:

The duration of nocturnal events was also a key component of the statistical analysis. Event duration did not tend to vary by month, as the average length of precipitation events was 4.9, 4.8, and 4.7 hours for May, June and July. Perhaps the most revealing characteristic of the data has to do with the long duration of events initiating prior to 0100LST. These events display a distinct durational characteristic, as many of them persist well into the early morning hours. On average, events beginning prior to midnight last for over 7.5 hours.

The intensity of the precipitation by hour (Figures 4.15-4.17) is at its highest in the early morning hours, with the highest average precipitation occurring during the period of 2300-0200LST. As a subjective decision for analyzing the hourly averages for precipitation, the hours had to have more than ten recorded observations for each month for the entire study period. The data for May includes the hours of 2000-1400LST, Whereas June and July include the hours of 2200-1200LST. May naturally has a wider time frame due to the earlier tendency and longer lasting nature of storms during the month. For May, the highest average occurred the evening before at 2000LST (0.27 inches or 6.9 mm). Interestingly, a sharp drop off occurs from 2000 to 2100LST, indicating that storms which move through the area tend to be at their heaviest during the 2000LST hour. Yet, a sharp increase occurs at the 0100LST hour.

Similarly, June precipitation tends to be the heaviest at 2300LST as well. In fact, the 2300LST hour during June has the highest precipitation average in the dataset, averaging over 0.40 inches or 10.2 mm for the entire study period. Precipitation tends to gradually decrease into the early morning hours, however. July also shows a tendency for the 2300LST hour, though similar to May an increase occurs at 0100LST as well. Why May and July have an increase in precipitation during 0100LST raises important questions regarding its possible relation to an increasing influence of low-level wind maxima and will be discussed in the following chapter. An interesting aspect of the data is a spike in hourly precipitation averages in the morning hours from 0600LST to 0800LST. In May, the increase begins at 0700LST and reaches a maximum at 0900LST. June reaches a minimum in the morning at 0500LST, yet a sharp increase occurs over the next two hours. Though not as noticeable as May and June, July displays also has an early morning increase in precipitation as well, occurring at 0500LST and 1000LST.

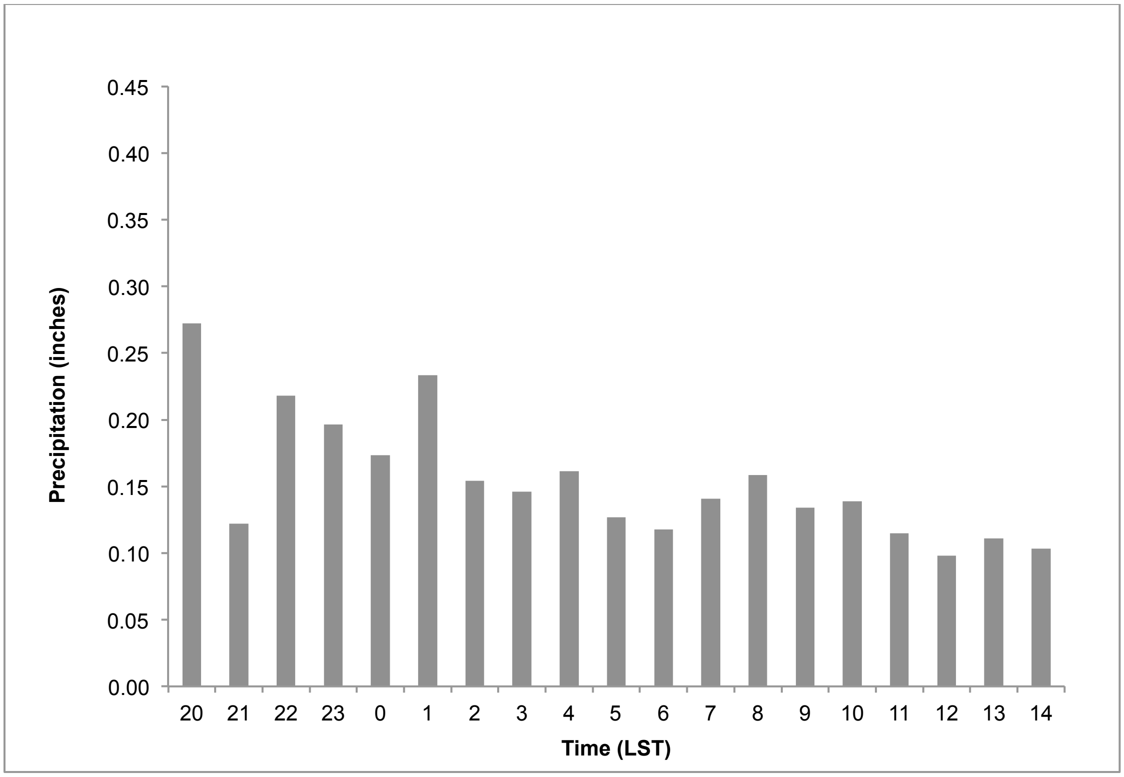


Figure 4.15: Average precipitation by hour for May. Note some hours are excluded from analysis due to their small sample size

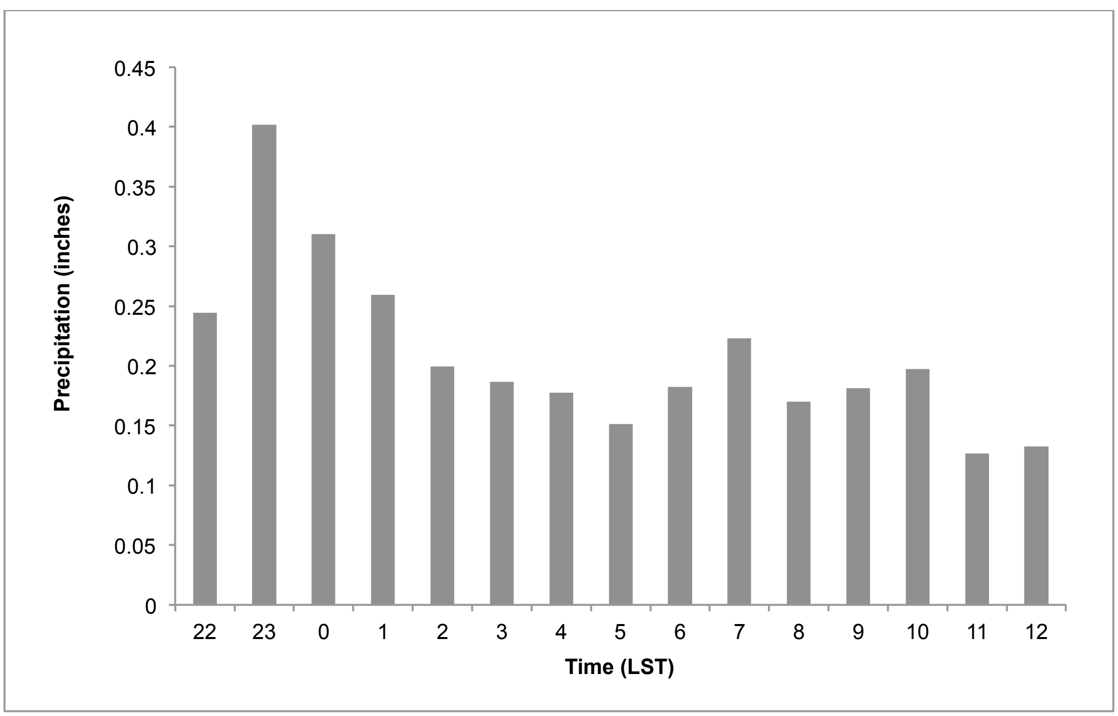


Figure 4.16: Same as 4.15 but for June. Note some hours are exlcuded.

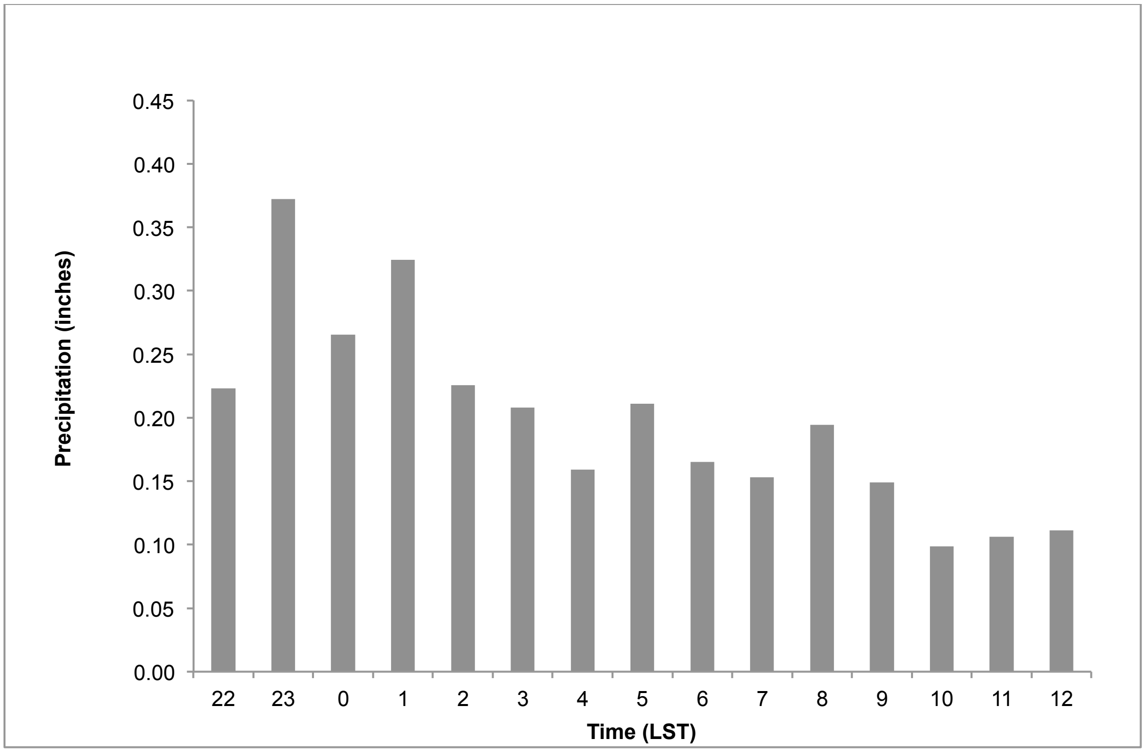


Figure 4.17: Same as 4.15 but for July. Note some hours are excluded.

Both the duration and the intensity correlate with the timing of the event. Not surprisingly, the events that last for longer periods of time have larger precipitation totals associated with them. In fact, nocturnal events greater than 1.50 inches (50.8 mm) on average last 8.2 hours. Conversely, events less than 1.50 inches (50.8 mm) persist for an average of 5.65 hours. As a result, many of the events that begin before 0100LST have the largest rainfall totals. Over 18% of all events were over 1.50 inches (50.8mm) (86 out of 463) and these major rain events occur much more often in June and July (Table 4.03). Nocturnal events greater than 1.00 inches or 25.4 mm increase from around 25% in May to near 35% in June and July. Moreover, nearly 25% of all rainfall events in June are over 1.50 inches (50.8 mm)

	May	**June**	**July**	**MJJ**
.25-.49 inches	58 (37.7)	41 (26.1)	45 (29.7)	144 (31.3)
.5-.99 inches	57 (37.0)	62 (39.5)	51 (33.6)	170 (36.6)
1.00-1.49 inches	21 (13.6)	15 (9.5)	27 (17.8)	63 (13.6)
>1.5 inches	18 (11.0)	39 (24.9)	29 (19.1)	86 (18.5)

Table 4.03: Frequency (and percentage) of nocturnal rainfall events.

The most noticeable feature of the data is the consistent presence of large events initiating prior to 0100LST, though the bulk of the precipitation during those events falls between 0100-

0900LST. Thirty-nine out of the 86 events initiated prior to 0000LST. Dividing the categories even further, 30 out of 46 events greater than 2.00 inches (50.8 mm) began at 0000LST or before, with half (15 out of 30) being recorded at 2200 and 2300LST. Therefore, the occurrence of extremely heavy rainfall event at Topeka generally equates to the storm beginning the evening before and persisting well into the morning hours. Moreover, only 12 times during the entire study period did a storm event recorded during the hours of 0200LST and 0900LST exceed 2.00 inches (50.8 mm). Subdividing events greater than 1.5 inches (50.8 mm) by month, June experienced 39 in the 63 year study period, whereas July has had 29 and May 18 (Table 4.04). Though a good majority of the longer duration events begin in the evening, it does not necessarily equate to the highest precipitation values occurring prior to 0100LST. When analyzing each hourly time slot by its average precipitation for events greater than 1.5 inches (50.8 mm), the highest precipitation amounts tend to be recorded from 2300 and 0200LST (Figure 4.18). Slight increases also occur during the morning hours of 0400-0700LST. More than likely, the development and eastward propagation of MCSs through the nighttime hours results in their arrival at Topeka roughly around sunrise.

	# >1.5in	**Average Start Time**	**Average Duration (Hours)**
May	18	2230LST	9
June	39	2330LST	7.46
July	29	0015LST	8.17

Table 4.04: Duration characteristics by month.

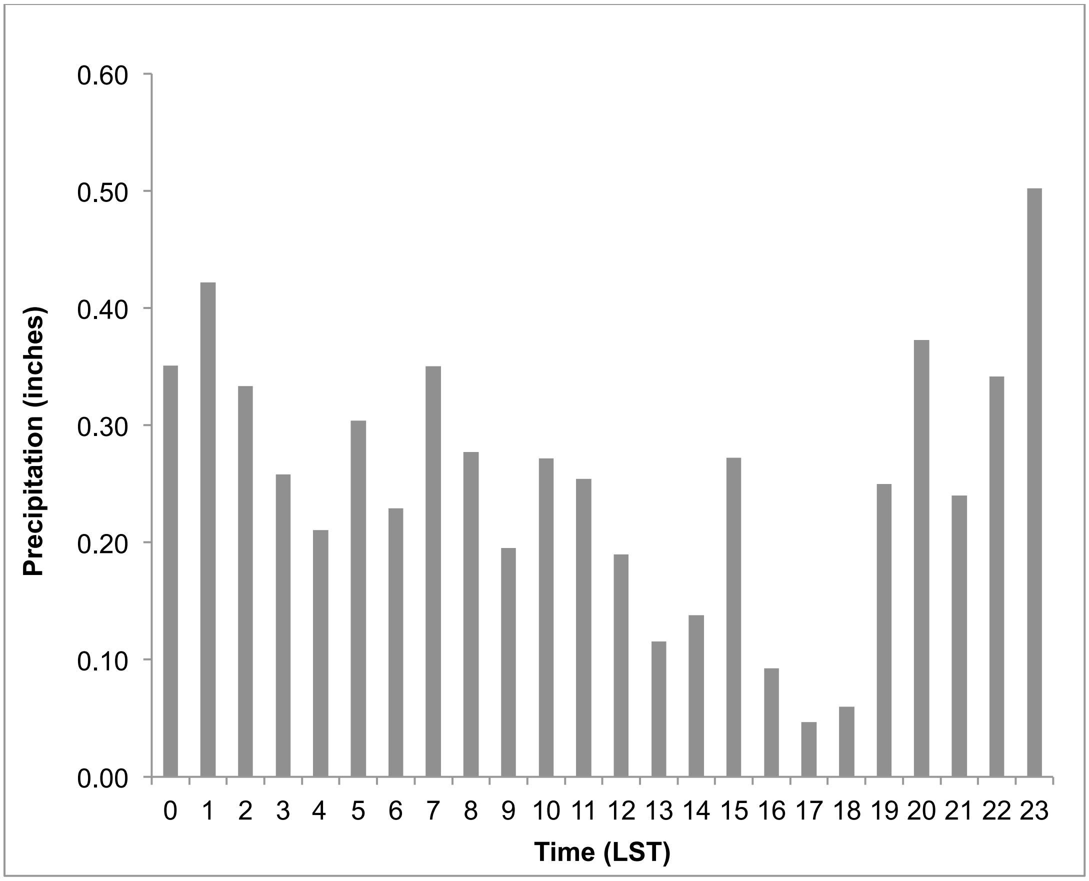


Figure 4.18: Average precipitation for each hour for the 86 events >1.5 inches (38.1 mm) (46 events).

4.4 Spatial Characteristics:

In addition to the temporal characteristics of nighttime rainfall, the spatial characteristics of nighttime events were analyzed using additional stations across Kansas. The spatial analysis can be divided into two parts. The first section analyzes the spatial properties of nighttime rainfall based on all 463 events recorded at Topeka. The subsequent section then examines the climatological characteristics of nighttime rainfall for each of the selected stations, answering questions about the similarities and differences in nocturnal rainfall between Topeka and stations to west. Any rainfall recorded at a station exceeding 0.10 inches (2.5mm) and occurring within a four hour window prior to the first precipitation recorded at Topeka, or two hours after the first recorded precipitation counted as a related event.

	May (154)	June (156)	July (153)	Total (463)
Oakley	49	45	40	134
Hays	71	57	59	187
Kanopolis	96	80	73	249
Concordia	98	100	96	302
Hutchinson	64	55	48	167
Wichita	81	55	43	179

Table 4.05: Number of recorded events at additional stations when Topeka reported >0.25 inches or 6.35mm of rainfall.

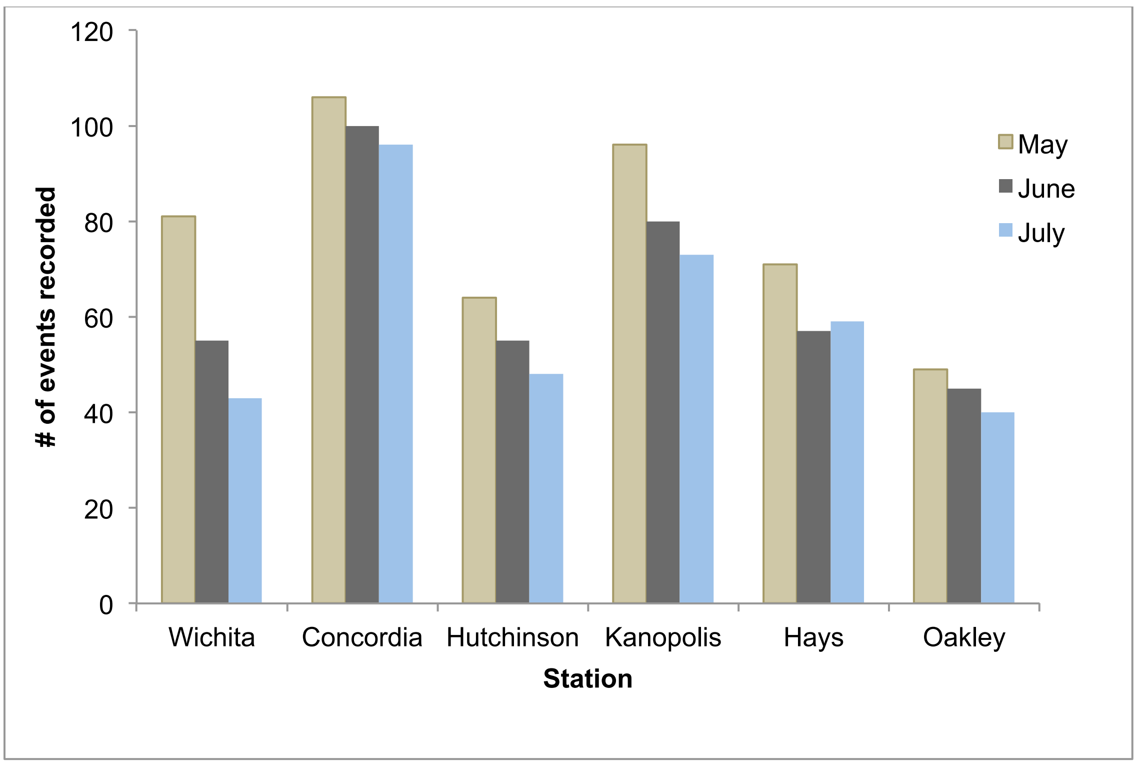


Figure 4.19: Number of times rainfall was recorded at other stations coinciding with rainfall reported at the Topeka Station.

Clearly, nocturnal rainfall activity reported in Topeka shares a much stronger relationship with areas more directly west rather than the southwest., as Concordia recorded more events every month than Wichita despite being similar distances apart from Topeka (Table 4.05). In fact, Concordia rainfall coincided with rainfall recorded at Topeka 294 out of 463 times (roughly 63% of the time). Similarly, Kanopolis reported more events than Wichita as well, though slightly less than Concordia (249 out of 463). Suprisingly, Wichita and Hutchinson had fewer events than Kanopolis and Concordia. Many storms during the late spring and early-summer tend to be steered by southwesterly flow aloft and would migrate in a northeasterly direction. Perhaps these southwest-to-northeast moving storms tend to be different in character and do not qualify as nocturnal events. Not surprisingly, the station in the far western part of the state had the lowest correlation with Topeka, as rainfall at Oakley only matched up with Topeka 29% (134 out of 463) of the time. Regardless, storms during May tend to be more widespread events, as every station recorded the most events during the month.

As mentioned in the first section of the chapter, the distribution of events between the months recorded at Topeka did not vary greatly. Yet, almost every station displays a decrease in the number of events recorded for May-July that coincide with rainfall recorded at Topeka (the exception being Concordia for May and June and Hays during June and July). Therefore, the progression from late-spring into early summer must result in a decrease in the spatial extent of nocturnal precipitation. Causes of the decreasing coverage of storm rainfall may be the result of fewer frontal boundary-related events. A broad zone with multiple areas of thunderstorms associated with an approaching cold front would naturally increase the strength of the relation between Topeka and additional stations. Perhaps storms in June and especially July form at night as a result of surface convergence associated with a southerly wind, creating more isolated clusters of storms rather than linear squall lines.

Rainfall data for each station spanning the 1950-2012 period for May, June and July also provided important insights into nocturnal rainfall based on latitude and longitude of each station. Places in western areas of Kansas do not experience nighttime rainfall as often as locations such as Topeka, due in large part to the differing diurnal cycle of thunderstorm development in the central Great Plains. Hourly rainfall data clearly displays the cycle, best depicted in Table 4.06 and 4.20, 4.21, 4.22 and 4.23. A general decreasing trend can be seen in both total precipitation and nighttime precipitation moving along the east-to-west transect from Topeka out to Oakley in the far western part of the state. Total precipitation decreases by 4.97 inches (126.2 mm) between the two stations, whereas total nighttime rainfall decreases by 3.72 inches (94.5 mm). However, the east-west nighttime precipitation gradient does not hold true for all months. For instance, during May, areas to the north and south of Topeka (Wichita and Concordia) on average, record more nighttime rainfall, despite Topeka having a higher average of overall precipitation. Moreover, in May and July the precipitation totals across the state do not vary as much compared to June. In fact, the largest precipitation gradient (both total and nighttime) exists during June. June also displays the clearest precipitation gradient out of the all the months, as every station to the west of the other stations recorded less nighttime and total rainfall.

	May (N)	May (T)	June (N)	June (T)	July (N)	July (T)	Total (N)	Total
Topeka	2.09	4.46	2.74	5.11	2.47	3.98	7.30	13.55
Wichita	2.25	4.11	2.49	4.58	2.17	3.56	6.91	12.26
Concordia	2.33	4.19	2.14	4.07	2.23	3.74	6.70	12.00
Hutchinson	1.49	4.04	1.98	4.44	1.61	3.71	5.08	12.18
Kanopolis	1.63	3.87	1.71	3.82	1.87	3.50	5.21	11.19
Hays	1.33	3.27	1.27	3.26	2.00	4.00	4.33	10.20
Oakley	1.20	2.98	1.06	2.30	1.34	3.29	3.56	8.58

Table 4.06: Rainfall averages by station. (N) Represents nighttime rainfall, whereas (T) represents total rainfall for each station by month.

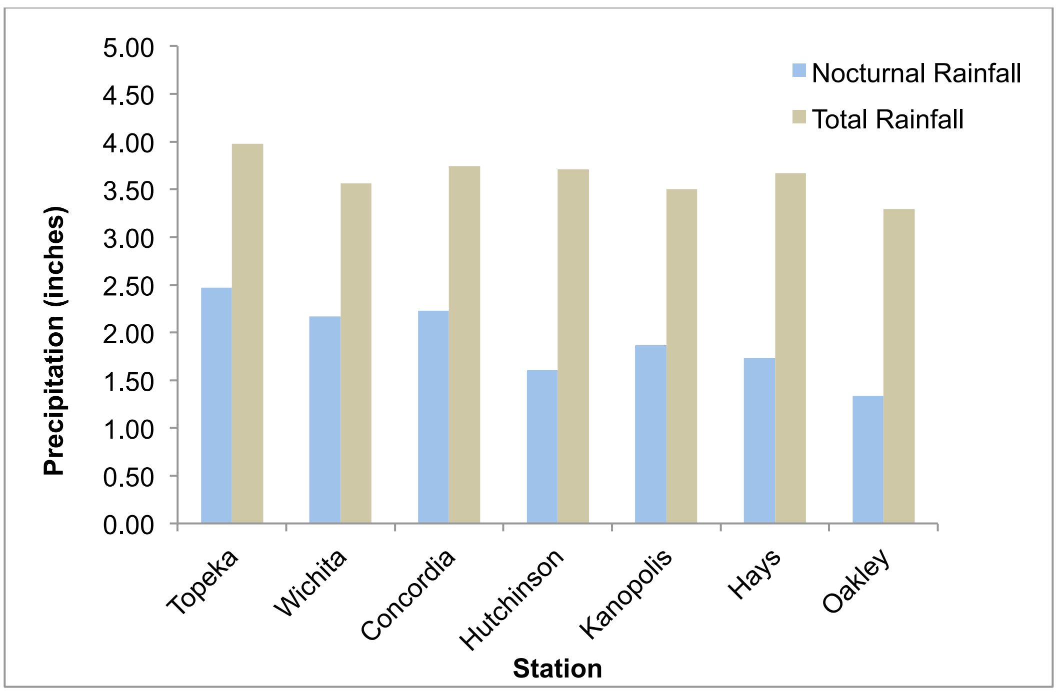


Figure 4.20: Rainfall characteristics for May for each station.

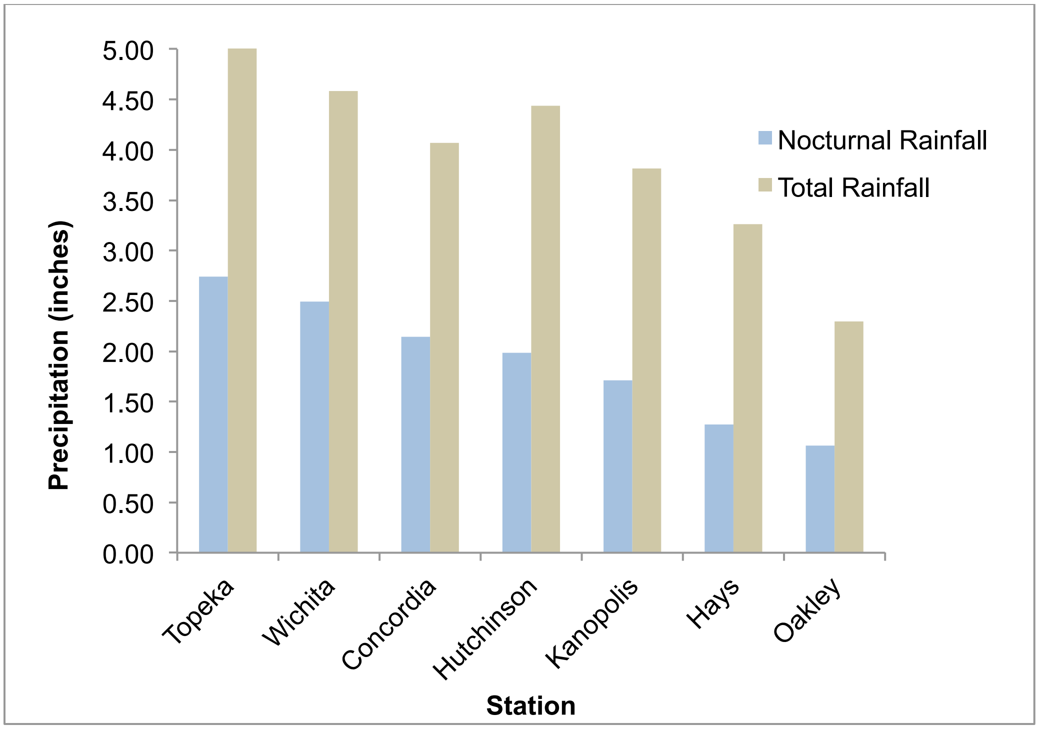


Figure 4.21: Same as 4.20 but for June.

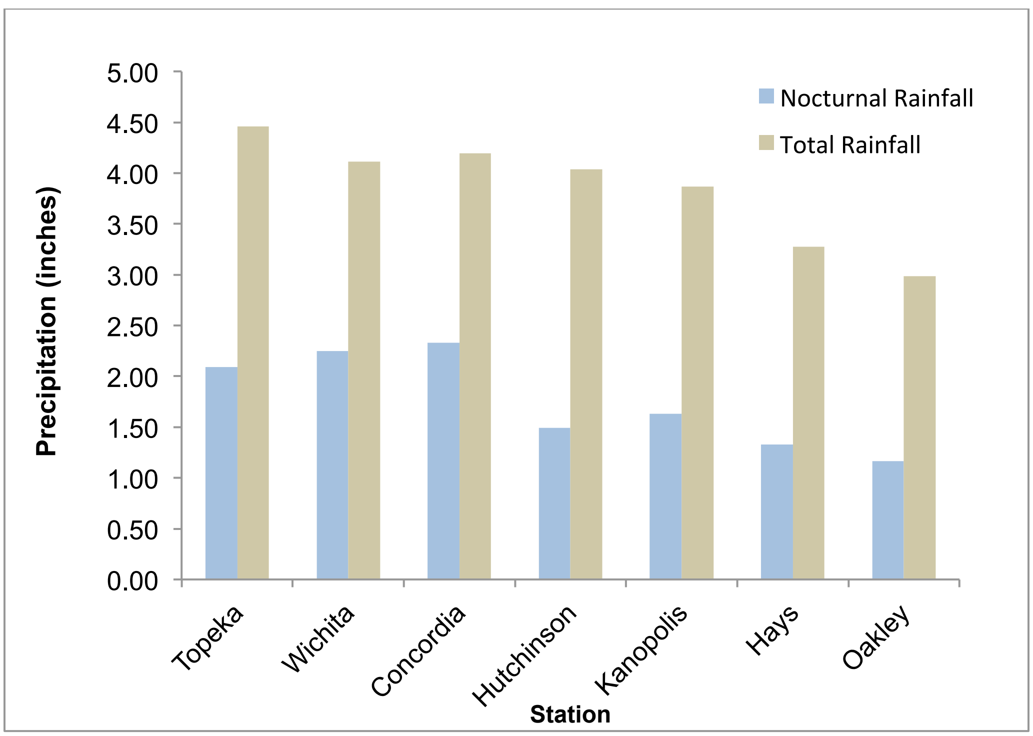


Figure 4.22: Same as 4.20 but for July.

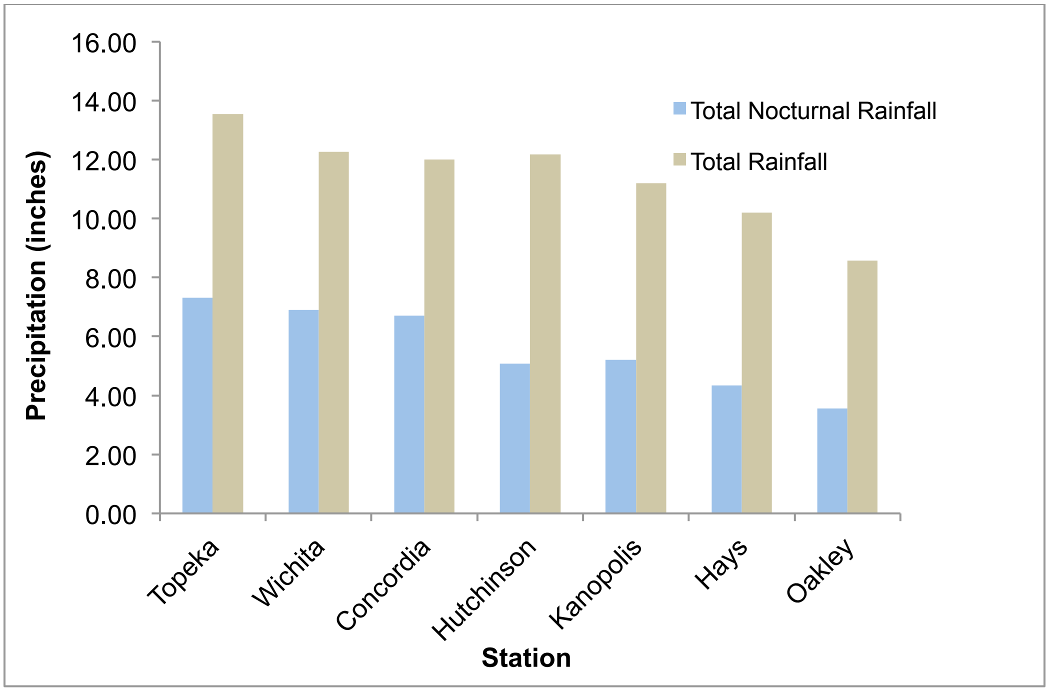


Figure 4.23: Rainfall characteristics for May, June and July for each station.

When broken down by percentages, the east-west gradient can once again clearly be detected (Figure 4.24). Whereas nighttime rainfall in the three eastern locales (Topeka, Wichita, Concordia) makes up between 40-50% of the overall rainfall total each month, nighttime rainfall in areas to the west makes up a lower percentage of the overall total. Even though Wichita and Topeka had a relatively low correspondence between nocturnal events, Wichita still receives a similar amount of nocturnal rainfall. This observation suggests that differing atmospheric processes may account for east-west differences in noctural rainfall. Total nocturnal rainfall percentage decreases by nearly 14% from Topeka to Oakley. The largest difference between stations occurs during the month of June, where over 50% of rainfall occurs at night in Wichita, whereas only 30% occurs in Oakley. In addition, whereas the eastern stations see an increase in nocturnal rainfall from May to June, nocturnal rainfall in western Kansas decreases. During July, the east-west difference decreases, and approximately 15% separates the highest percentage at Topeka (51.6%) from Oakley (36.3%).

July appears to be the month where the largest amounts of nighttime rainfall occurs across the state. Five out of the seven stations record their largest percentage of nighttime rainfall during July. Nocturnal rainfall increases by as much as 7% and 8 % from June to July in Hays and Oakley, perhaps indicating that a mechanism which promotes nocturnal storms increases from June to July. Perhaps daytime convective heating in eastern Colorado along the Colorado Front Range promotes thunderstorm activity during the afternoon hours. The afternoon thunderstorms in eastern Colorado may help destabilize the atmosphere in western Kansas and propogate eastward, as proposed by Hales (1977).

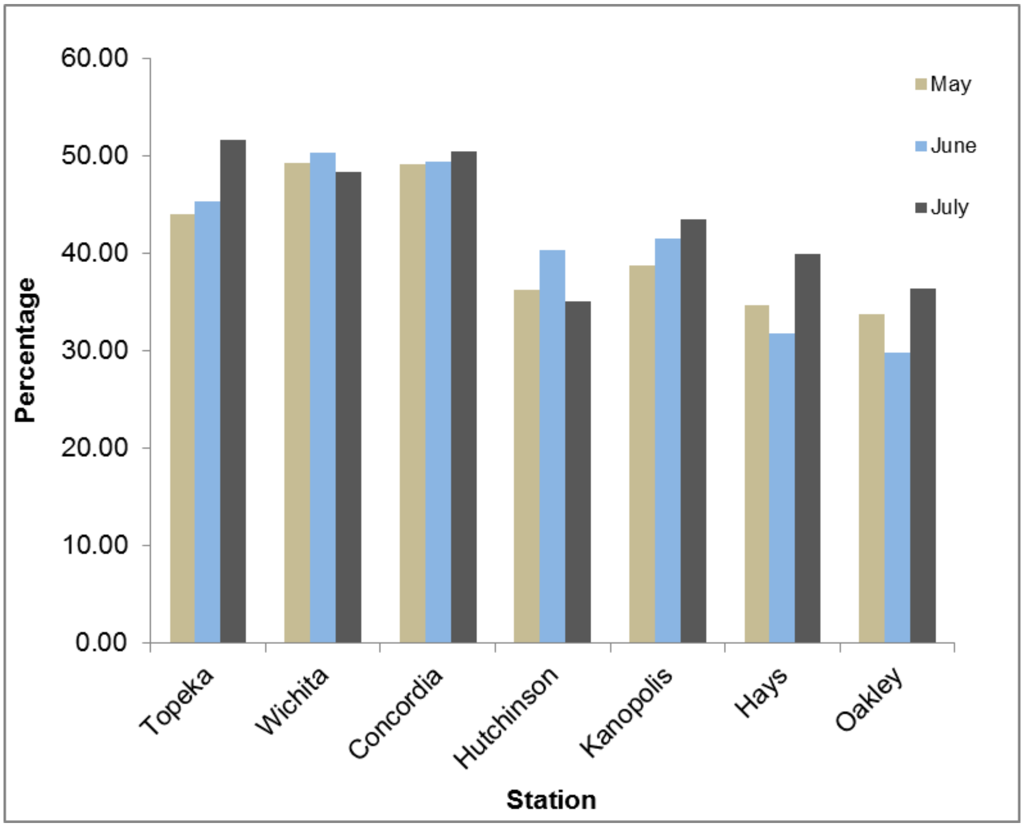


Figure 4.24: Nocturnal rainfall percentage for each station for May, June and July (1950-2012).

4.5 Discussion:

Based on the hourly weather data from Topeka and additional stations, nighttime rainfall significantly influences the hydroclimatology in Kansas. Nocturnal rainfall contributed anywhere from 30% of the total rainfall during May, June and July in the western locations, to as much as 50% of the rainfall in areas of central and eastern Kansas. The nocturnal rainfall contribution shows a clear increase as the warm season progresses from May to July. However, although nocturnal rainfall increased from May to July, the overall coverage of storms decreases. Every station displayed a decrease in the number of events recorded coinciding with events at Topeka from May to July. These characteristics of nocturnal rainfall raises a number of questions about the processes that influence these rain events, in particular how these rains relate to the wind maxima in the southern Great Plains. The next chapter will discuss the importance of low-level wind maxima to nocturnal rainfall, analyzing its characteristics based on the rainfall data presented in this chapter.

CHAPTER FIVE

ANALYSIS OF LOW-LEVEL WIND ORIENTATION AND SPEED MAXIMA FOR HEAVY NOCTURAL RAIN EVENTS AT TOPEKA

"The substance of the winds is too thin for human eyes, their written language is too difficult for human minds, and their spoken language mostly too faint for the ears."

~John Muir

The results presented in this chapter illustrate some of the characteristics of the Great Plains low-level wind maxima at the time of a heavy precipitation event in Topeka, Kansas. Output from reanalysis of atmospheric conditions was the input for these analyses, since that reanalysis modeling provided data on wind speed and direction every six hours for the three desired levels of the lower troposphere (925 mb, 850 mb, and 700 mb). Results are presented first for an examination of temporal and spatial patterns of wind speeds and direction at a specific point or grid location to the south and west of Topeka. A second set of studies examined the regional character of the low-level wind flows using a classification system designed specifically for this study. The classes of low-level wind maxima were designed to separate events into categories based on the degree to which the specific event maintained classic low-level wind characteristics. Both of these sets of analysis provide findings that address the second major research topic addressed in this thesis: describing the character of the low-level wind maxima during large nocturnal rainfall events in northeast Kansas.

5.1 Low-level winds associated with large precipitation events: *Wind Directions at 37.5°N-97.5*°W:

The low-level wind characteristics for nighttime rainfall events were analyzed using criteria discussed in Chapter Three. Heavy rainfall events, those exceeding 1.5 inches (38.1mm) were selected from the data, resulting in a total of 86 events analyzed for May, June and July. For comparison purposes, the 86 events were divided by the month to in order to understand both the monthly and climatological characteristics of nighttime wind maxima. From the criteria discussed in Chapter Three, common modes of circulation associated with the low-level winds were extracted for heavy rainfall events recorded at Topeka. This chapter will begin by discussing the results for heavy nighttime rainfall events for the study period, using a grid cell located at 37.5°N-97.5°W (Wichita grid cell) as a basis for interpreting low-level winds.

Low-level winds during events greater than 1.5 inches (38.1mm) have distinct characteristics based on the month of their occurrence; therefore each month has a separate results section Beginning with May, the Topeka weather station recorded a nighttime rainfall greater than 1.5 inches (38.1mm) 18 times during the 63-year period. Based on the criteria selected, low-level wind maxima characteristics were present during many of the rainfall events for the Wichita grid cell. In all cases, a southerly wind (90˚- 270˚) greater than 8 m/s (17.9 mph) was present at the grid cell in at least one of the time slots (0z, 6z, 12z, 18z). In six out of the 18 cases, winds were greater than 8 m/s (17.9 mph) at all 4 time slots at 925 mb, 850 mb and 700 mb. Though, in three cases no low-level wind profile could be detected among the four time slots. However, the winds greatly differ throughout the diurnal cycle. At all time slots, the most easterly oriented winds were present at the lower levels. In fact, during the month of May the average wind direction for the 700 mb level was greater than 200˚, indicating southwesterly flow aloft. Though, as the night progressed, the winds shifted towards the southwest, particularly at 925 mb and 850 mb. A wind that changes orientation with shifts in a clockwise direction is described as a veering wind. At 700 mb, however; little veering was present, and in fact the winds actually backed slightly eastward from 0z (1800LST) to 6z (0000LST). Regardless, winds tended to veer with height, as winds at 925 mb on average blew from a more easterly direction compared to winds at the other two pressure levels (Table 5.01).

During June, 39 rainfall events occurred which exceeded 1.5 inches (38.1mm). At the Wichita grid cell, for seven out of 39 times assessment of reanalysis data suggested low-level wind maxima were essentially non-existent. The average direction of the winds once again veered vertically and throughout the study period time, with the exception of 700 mb where the west-southwest wind directions remained relatively the same from 0z (1800LST) to 18z (1200LST). Compared to May, the wind directions veer much less and are oriented in a more southerly direction. Additionally, though the wind directions in May continued to veer strongly from 12z (0600LST) and 18z (1200LST), winds turn very little at those times during June. Similarly, however; the largest change in wind direction over time occurs at 925 mb, as wind shift from the south-southeast at 0z (1800LST) to south-southwest at 18z (1200LST) (Table 5.02).

For the 63-year period, 29 nocturnal events greater than 1.5 inches were reported at Topeka during July. In only two cases for July did low-level wind maxima persist from 0z (1800LST) to 18z (1200LST) at the Wichita grid cell. In an additional four cases, no nocturnal low-level wind maxima could be seen in the data and were excluded from analysis.

Additionally, in a number of July events the wind characteristics could be deemed non-jet events. Often times wind speeds would exceed 8m/s for only one level and one time slot, thus raising questions on how many times do low-level wind maxima characteristics have to appear in order to be considered a wind event. Only after examining the rest of the study area were these borderline events able to be classified as a wind maxima or non-wind maxima event. Of course, the opposite held true as well. Whereas the Wichita grid cell exhibited wind maxima, the rest of study area did not. Further classification of jet and non-jet event will be discussed in subsequent sections of the chapter. Similar to May and June, the average wind directions for July have a westward progression with height and throughout the study period. Yet, the magnitude of the veering decreases again in July. In fact, at 925 mb and 850 mb, winds actually shift back to the east from 12z (0600LST) to 18z (1200LST) (Table 5.03).

	925 mb	**850 mb**	**700 mb**
0z	132°	166°	223°
6z	159°	176°	217°
12z	214°	210°	226°
18z	238°	254°	236°

Table 5.01: Average wind direction for May rainfall events >1.5in at 37.5°N-97.5E°.

	925 mb	**850 mb**	**700 mb**
0z	155°	181°	237°
6z	173°	197°	235°
12z	188°	215°	250°
18z	194°	217°	239°

Table 5.02: Same as Table 5.01 for June.

	925 mb	**850 mb**	**700 mb**
0z	160°	181°	247°
6z	171°	196°	223°
12z	208°	235°	251°
18z	187°	226°	260°

Table 5.03: Same as 5.01 for July.

5.2 Average Wind Speeds at 37.5˚N-97.5˚E:

The average wind speeds at each of the three pressure levels also differ both vertically and temporally for the three months. In May, the average fastest winds are located at 700 mb, the exception being 6z (0000LST) where the fastest winds shift to 850 mb. Nevertheless, a clear pattern of increasing winds from 0z (1800LST) to 6z (0000LST) can be seen at all levels, followed by a general decrease from 6z (0000LST) to 18z (1200LST). Interestingly, the largest decreases from 6z (0000LST) to 18z (1200LST) occur at 925 mb and 850 mb, Whereas winds only decrease slightly at 700 mb. In fact, wind speeds actually increase at 700 mb from 12z (0600LST) to 18z (1200LST). The greatest changes in wind direction were evident at the lower levels, Whereas 700 mb remained relatively the same throughout the 18-hour period (Table 5.04).

During June, average wind speed increases from 0z (1800LST) to 6z (0000LST), followed by a general decrease through the night and into the morning at 925 mb and 850 mb. Moreover, winds reach their max averages at 6z (0000LST) at both 925 mb and 850 mb, and their low at 18z (1200LST). Conversely, at 700 mb, wind speed averages reach their minima at 6z (0000LST) and peak at 18z (1200LST). The variation in wind speed is also more prevalent at the lower levels, with both 925 mb and 850 mb having the starkest differences in average wind speed from 6z (0000LST) to 18z (1200LST). At 700 mb, wind speeds vary little from 0z (1800LST) to 18z (1200LST)(Table 5.5). The most noticeable difference from May to June is the general decrease in wind speeds at all time slots and all levels, with the exception of 925 mb at 6z (0000LST). Particularly at 700 mb, wind speeds on average decrease by as much as 3 m/s (6.7 mph) (Table 5.05).

Comparing the 3 months, average winds were by far the weakest at July across all time slots and levels. Only four out of a possible 12 times did wind averages exceed 8 m/s (17.9 mph) (Table 5.06). Similar to other months, however; wind speeds at 925 mb and 850 mb reach their max at 6z (0000LST) and decrease significantly to 18z (1200LST). July has steepest decrease in wind speed averages at 925 mb and 850 mb from 6z (0000LST) to 18z (1200LST), decreasing by over five m/s at 952mb and 4.8 m/s at 850 mb. At 700 mb, however, winds increase slightly at every time period but remain their weakest out of the three months. Weaker low-to-mid level winds suggests that the mechanisms present which caused faster winds earlier in the warm season disappear come the middle of the summer in the Great Plains (Table 5.6). The weaker winds likely result from the northward contraction of the upper-level westerlies as the warm season progresses. In May, the westerlies can remain relatively far south with the influence of the solar declination and the movement of storm systems and colder air masses. Yet, in July the sun is at one of its highest northern latitudinal extents of the year. Also, rarely do the type of storm systems common in May influence the region in July.

	925 mb	850 mb	700 mb
0z	8.8	10.2	**12.9**
6z	10.6	**13.6**	13.3
12z	9.1	11.5	**12.5**
18z	7.8	9.9	**13.1**

Table 5.04: Average wind speed (m/s) for May rainfall events

	925 mb	850 mb	700 mb
0z	8.0	9.5	**9.8**
6z	11.0	**13.0**	9.2
12z	7.5	**10.3**	9.8
18z	6.5	8.1	**10.2**

Table 5.05: Same as Table 5.04 for June.

	925 mb	850 mb	700 mb
0z	7.5	**8.2**	7.1
6z	10.5	**11.6**	7.1
12z	7.3	**9.6**	7.9
18z	5.2	6.4	**8.1**

Table 5.06: Same as Table 5.04 for July.

5.3 Wind Characteristics for 0z, 6z, 12z and 18z:

0z (1800LST):

Analyzing the individual time slots can provide further understandings about the wind speed and directions at the three pressure levels. Looking at the 0z (1800LST) time slot, the maximum wind at the Wichita grid cell during May was most frequently located at 700 mb, as 11 out of the 15 wind maxima were located at the level (three events were excluded from analysis due to no jet profile from 0z (1800LST) – 18z (1200LST). In contrast, three wind maxima occurred at 850 mb and only one at 925 mb. Though, in one instance none of the three levels exhibited southerly wind maxima characteristics for 0z (1800LST). In addition to pinpointing where wind maxima occur in the vertical, the extent of the winds in the vertical provides another avenue of low-level wind characterization at a particular time slot. In 11 out of the 15 events at 0z (1800LST), low-level winds greater than 8 m/s (17.9 mph) were present at all levels of the atmosphere. In the other three instances, low-level wind maxima were present at two levels of the atmosphere. The one event with no wind maxima characteristics at 0z (1800LST) was excluded from analysis at this time slot.

During June, the maximum winds at 0z (1800LST) are most common at 850 mb. Seventeen out of 32 wind maxima occur at 850 mb, Whereas 13 occur at 700 mb, and only two at 925 mb. Similar to May, eight events are excluded due to no wind maxima profile. The vertical profile of the low-level winds changes from May to June as well, as low-level wind maxima at all levels at 0z (1800LST) were present in 21 out of 32 events, a lower percentage compared than May. Five events had low-level wind maxima present at two levels, Whereas four were located in only one level. However, in two cases no level of the vertical exhibited maxima characteristics.

At 0z (1800LST) for July, maximum winds in 18 out of the 25 at the Wichita grid cell occur at 925 mb or 850 mb. Four events are removed from analysis due to no low-level wind profile from 0z (1800LST) to 18z (1200LST). For the first time however; a number of events display no wind maxima at 0z (1800LST). In fact, five times at 0z (1800LST) no wind maxima characteristics can be detected at any level of the atmosphere; largely a result of winds being below 8 m/s (17.9 mph) but still blowing from a southerly direction. The increase in non-jet winds at 0z (1800LST) provides further evidence that low-level wind maxima during July nocturnal rainfall events either do not form in the evening, do not form in southern Kansas, or occur later in the diurnal cycle. The lack of southerly wind maxima also strengthens the notion that low-level wind maxima during the latter portion of the late spring-early summer precipitation maximum can best be seen in the lower levels of the atmosphere. Whereas in May and June a majority of wind maxima could be detected at all three levels, in only eight cases does that occur at 0z (1800LST) in July. For six cases a low-level wind maximum exists at two levels of the atmosphere. In the other six cases, the wind maxima exist at one level of the atmosphere.

6z (0000LST):

Similar patterns can be seen at 6z (0000LST) as well, though the average directions based on height shift more westward at 925 mb and 850 mb with only slight veering at 700 mb. Maximum wind speeds, however; shifts to lower elevations. In 10 cases maximum winds at the Wichita grid cell at 6z (0000LST) occur at 850 mb, Whereas only four occur at 700 mb and one at 925 mb. In 13 of the 15 cases, low-level wind maxima can be seen at all three levels of the vertical. In the other two cases, low-level winds are present at two levels of the atmosphere. Similar to May, the maximum wind at 6z (0000LST) in June frequently exists at 850 mb (25 out of 31 events), compared to only three at 925 mb and three at 700 mb. In 21 of the cases wind maxima profiles persist at all three levels. An additional 10 events have winds that fit the wind maxima criteria at two levels of the atmosphere, with no events displaying low-level wind maxima at one level.

Low-level wind maxima occur at much lower levels at 6z (0000LST) during July, as 24 out of the 25 wind maxima measured for the time slot are located at 850 mb and 925 mb. The lack of wind maxima at the 700 mb reinforces that notion that low-level winds during July form and evolve throughout the night at the lower levels. The highest number of wind maxima at 925 mb occurs at 6z (0000LST) for any month and any time period as well, with seven wind maxima present at that level. The increase in wind speeds from 0z (1800LST) to 6z (1200LST) also led to an increase in low-level wind coverage, as 13 events had a nocturnal maxima present at all three levels. Two levels with wind maxima increased to 10. In the case of wind maxima at two levels, eight out of 10 cases the winds occur at 925 mb and 850. In the remaining two cases, one event display wind maxima at one level, with the other event having no wind maxima.

12z (0600LST):

The presence of faster winds in the upper-levels continues into 12z (0600LST) in May, as 14 out of 15 possible low-level wind events occur at either 850 mb or 700 mb. In eight instances, the max wind resides at 700 mb, followed by 6 at 850 mb and one at 925 mb. Though, in three cases with a 700 mb max wind, the wind directions at all levels contain a strong westerly component (>195° at all 3 levels). Therefore, it begs the question if those events at 12z (0600LST) should be considered "southerly low-level wind maxima," given their westerly nature. The number of jet occurrences at all levels also greatly decreases at 12z (0600LST), as only eight out of 15 events encompass all three levels of the vertical. Events with two levels of wind maxima increases, however; as four events have wind maxima at two levels of the atmosphere, followed by only one event with one level.

At 12z (0600LST) in June, wind maxima in the vertical are most prevalent 850 mb (20 out of 31 events), though the number of wind maxima at 700 mb increases from only three at 6z (0000LST), to 11 at 12z (0600LST). Interestingly, the number of events that have no jet profile increases greatly from 6z (000LST) to 12z (0600LST), suggesting that some jet events during June reach their maximum around the 6z (0000LST). Unfortunately, due to the coarse temporal resolution of the data, it cannot be determined whether wind maxima during nocturnal rains in June happen before or after the local midnight hour. In six cases, winds at three pressure levels displayed no southerly-wind characteristics. Two cases the winds have a southerly wind component but their speeds do not exceed 8 m/s (17.9 mph). The other four times winds speeds are below 8 m/s (17.9 mph) and blow from a northerly direction. Three-level wind maxima decrease from 6z (0000LST) to 12z (0600LST), with only 15 cases exhibiting a southerly wind maxima in all levels of the atmosphere. Only six times could wind maxima be detected at two levels of the atmosphere, with the final five occurring in only one level (the six with wind maxima characteristics were excluded). The decrease in vertically pronounced wind maxima from the midnight LST to the morning hours of 12z (0600LST) more than likely results from sunrise would weakening the nocturnal forcings in place during the night thus lowering wind speeds at all levels of the atmosphere and in June. Events that maintained wind maxima into 12z (0600LST) may have other mechanisms in place that allows the wind maxima to persist into the morning hours. Regardless, the dramatic drop-off in wind speeds from 6z (0000LST) to 12z (0600LST) helps document the importance of the nighttime hours for low-level wind maxima.

At 12z (0600LST) in July, wind maxima continue to persist in the lower levels, perhaps favoring the 850 mb height range more so than earlier hours. One of the more noticeable features of 12z (0600LST) is the number of events that possess no wind maxmia profile at the Wichita grid cell (seven out of 25 events). In most of the non-jet profile measurements, the wind characteristics still have a southerly/southwesterly component but do not have speeds greater than 8 m/s (17.9 mph). This phenomenon suggests that the same mechanisms may be present which create a southerly wind component, but their ability to create a fast, wind maxima within the vertical weakens. It also displays the importance of the nocturnal mechanisms to the low-level wind maxima formation later in the warm season. The low-level winds clearly reaches its peak between 6z (0000LST) and 12z (06000LST) and rapidly decreases in strength in both June and July; corresponding with the rising of the sun. Not surprisingly, the lack of wind maxima profiles also decreases the number of comprehensive jets in the vertical, as only six times could a robust wind be seen in the data for July. The majority of events, therefore; comprised only one or two levels of the atmosphere (nine and three, respectively).

18z (1200LST):

At the final hour of measurement, 18z (1200LST), wind maxima continue to occur at 700 mb at 18z (1200LST), with 12 out of 15 wind maxima occurring at 700 mb, followed by three at 850 mb. In two instances no level in the vertical exhibited southerly wind maxima. Though 18z (1200LST) was included in the study time period, the winds examined at this time should not be considered a "nocturnal wind maxima" given the time in the diurnal cycle. Although low-level wind maxima may continue to occur, the mechanisms driving west-southwest winds in the lower levels are likely not the same mechanisms that resulted in a nocturnal wind maxima and subsequently nighttime precipitation the previous night/morning. Nevertheless, including 18z (1200LST) in the study period allows for useful comparison of the winds from the nighttime hours into the early afternoon hours. Not surprisingly, the number of comprehensive jet events greatly decreases at 18z (1200LST). In only seven cases did a low-level winds encompass the entire lower troposphere; Whereas only two exist at two levels and four at one level (the two cases with no wind maxima were excluded). Therefore, the dominate mode of southerly winds was restricted to only 700 mb, strengthening the idea that low-level wind maxima have dissipated by the early afternoon hours following heavy rainfall events.

Wind maxima profiles continue to decrease from 12z (0600LST) to 18z (1200LST) in June as well, as only 25 out of 31 events have wind maxima-like characteristics in at least one level. However, wind maxima at 700 mb level increase from the previous hour of measurement, with 20 out of the 31 wind maxima present there. Only 10 wind maxima reside at 850 mb; the lowest of the four time slots. Only once do winds reach their peak at 925 mb. However, it must be noted that in the 31 events analyzed at 18z (1200LST), in eight cases maximum wind speeds are below 8 m/s (17.9 mph), or have a non-southerly orientation. The number of low-level wind maxima comprising all three levels decreases as well, as only 12 events have such characteristics. In four cases wind maxima occur at two levels, seven at one level and the remaining eight contain no low-level wind maxima profile.

The lack of wind maxima continues at 18z (1200LST) during July as well, as 13 out of the 25 events display no low-level jet characteristics. Compared to other months, low-level wind maxima at 18z (1200LST) basically cease to exist during July for the Wichita grid cell. In only six out of 24 cases where wind speeds exceed 8 m/s (17.9 mph) did winds at another pressure level exceed 8m/s (17.9 mph) and contain southerly wind component.   As a result, the number of events where a low-level could be detected at all three levels at 18z (1200LST) drops to three in July, followed by six for two levels of the atmosphere and three for one level.

5.4 Low-Level Wind Classification:

As discussed in the methods section, a low-level wind classification scheme was developed based on the variation recognized in the examination of the monthly characteristics of the southerly winds. The first class of winds comprises those that persist at all levels for all four time slots, or "Robust Class." The second category, which includes those events that have wind maxima at all three levels for at least three time periods, is the "Semi-Robust Class." The totals for Semi-Robust class range from 9-11. The third category of winds consists of those with totals of 6-8. Generally, these events tend to be confined to 0z (1800LST) to 6z (1200LST). The fourth category of LLJs consists of those with totals 3-5, generally exist to one level and persist throughout the four time slots. The fifth and final category includes those that exhibit little to no wind maxima with totals ranging from zero to two (Table 5.07).

Robust Class	Semi-Robust Class	0z-6z Class	3-5 Class	Non-Wind Maxima
15	22	18	16	15

Table 5.07: Wind Classes by category.

Beginning with the first "Robust" class, the maximum averages are by far the strongest of the five classes, with 850 mb at 6z (0000LST) recording the highest average of the entire study period at 18.8 m/s (42 mph). In fact, every level contains wind speeds higher than 10 m/s (22.4 mph). Without a doubt, the Robust Class are a late spring phenomenon, as 12 out of the 15 events occur in May or June. Six out 18 May events fall into the Robust category, whereas July records only three events with “Robust” characteristics (the lowest percentage of any category). The largest increase in winds occurs from 0z (1800LST) to 6z (0000LST), and the largest decrease occurs from 12z (0600LST) to 18z (1200LST). The significant increase and decrease in winds with sunset and sunrise suggest that while a wind maxima more than likely exists prior to sunset and persists past sunrise, the typical nocturnal mechanisms which creates a southerly wind component further enhance the winds, creating the fastest winds during the nighttime hours. Throughout much of the night the winds remain either southerly or southwesterly at all levels, with the most veering occurring at 925 mb and 850 mb. Winds tend to be more oriented toward the southeast at the lowest level, whereas a southwesterly wind direction is more common in the upper levels (Table 5.08).

	925 mb Wind		**850 mb Wind**		**700 mb Wind**	
	Direction	**Speed**	**Direction**	**Speed**	**Direction**	**Speed**
0z	159	11.5	179	13.7	226	**14.0**
6z	172	14.5	194	**18.8**	224	13.8
12z	185	12.3	211	**16.9**	229	13.6
18z	182	10.3	210	13.4	225	**13.9**

Table 5.08: Average LLJ characteristics for Robust LLJs. Wind directions are in degrees, wind speeds are in m/s. Max wind speeds for each time slot are bolded.

The Semi-Robust class of winds shares similar characteristics with the Robust Class, but these winds generally last for at least three rather than four time slots (11 out of 22 events exhibit this characteristic). The Semi-Robust category accounts for the largest percentage of wind maxima events, and by far appears to be a June phenomenon. Twelve out of the 22 Semi-Robust events occur during June, making up roughly 30% of all June events. May numbers drop to only four, Whereas only five occur in July. During these wind events, the region of fastest winds occurred in the upper levels, favoring the 850 mb once again with the fastest averages present at 6z (0000LST) and 12z (0600LST). Though, compared to Robust events, the wind speeds at all levels decrease significantly. Interestingly, much greater veering occurs with Semi-Robust events at 925 mb and 850 mb (Table 5.9), with wind directions changing by more than 30˚ and 40˚ at 925 mb and 850 mb respectively. Once again, wind directions remain relatively the same at 700 mb. Similar to Robust winds, wind speeds increase substantially from 0z (1800LST) to 6z

(1200LST), but decrease the most from 12z (0600LST) to 18z (1200LST), particularly at 925 mb and 850 mb (Table 5.09).

	925 mb Wind		**850 mb Wind**		**700 mb Wind**	
	Direction	**Speed**	**Direction**	**Speed**	**Direction**	**Speed**
0z	158	9.0	182	**10.6**	231	10.3
6z	172	12.3	196	**15.3**	225	10.4
12z	196	9.3	222	**13.2**	235	11.1
18z	204	7.3	227	9.8	233	**11.0**

Table 5.09: Average wind characteristics for Semi-Robust Class. Wind directions are in degrees, wind speeds in m/s. Max wind speeds for each time slot are bolded.

In 18 cases low-level wind characteristics representative of a low-level wind maxima exist in at least six time slots. With eight events, June tends to be the month where 6-8 class winds occur the most, followed by July with six and May with four. Though, from a percentage standpoint, the 6-8 class represents ~20% of the events for each month. In general, the 6-8 class constitute mainly those events where jets can only be seen at 0z (1800LST) and 6z (0000LST). In fact, at 0z (1800LST), LLJ characteristics in at least two levels of the atmosphere happened in 15 out of 18 events at 0z (1800LST), whereas every event at 6z (0000LST) have wind maxima characteristics in at least two levels. Comparatively, in only eight cases does this phenomena occur at 12z (1800LST), followed by only two occurrences at 18z (1200LST). At 18z (1200LST), low-level winds essentially disappear. The increase in westward veering continues to be a persistent pattern with the 6-8 jet class, with changes in wind direction totaling over 50˚ at both 925 mb and 850 mb (Table 5.10).

	925 mb Wind		**850 mb Wind**		**700 mb Wind**	
	Direction	**Speed**	**Direction**	**Speed**	**Direction**	**Speed**
0z	163	8.0	188	9.7	240	**10.5**
6z	180	10.5	205	**12.9**	235	10.3
12z	224	6.4	251	8.9	251	**10.0**
18z	215	5.0	249	6.3	260	**10.0**

Table 5.10: Average wind characteristics for events with jet characteristics in 6 to 8 time slots. Wind directions are in degrees, wind speeds in m/s. Max wind speeds for each time slot are bolded.

The fourth category of LLJs, or the 3-5 class, typically can best be seen at 6z (0000LST). Of the 16 3-5 class, the majority occur in June and July with seven events for each month, followed by only two in May. In only two cases are low-level winds present at two levels of the atmosphere at both 12z (0600LST) and 18z (1200LST) combined. Wind speeds largely average below the 8 m/s (17.9 mph) threshold for all time slots and levels, with the fastest averages occurring at 6z (0000LST) at 925 mb and 850 mb. Interestingly, whereas winds decrease from 6z (0000LST) to 18z (1200LST) at 925 mb and 850 mb, winds actually increase at 700 mb. Westward veering continues to increase from class-to-class, with winds veering at lower levels by over 70˚ and 80˚ at 925 mb and 850 mb respectively (Table 5.11).

	925 mb Wind		**850 mb Wind**		**700 mb Wind**	
	Direction	**Speed**	**Direction**	**Speed**	**Direction**	**Speed**
0z	144	6.7	171	6.9	243	**7.1**
6z	162	9.0	188	**9.1**	228	6.6
12z	203	5.5	207	6.3	252	**7.4**
18z	222	4.1	255	4.8	254	**8.2**

Table 5.11: Average wind characteristics for events with wind in 3 to 5 time slots Wind directions are in degrees, wind speeds in m/s. Max wind speeds for each time slot are bolded.

The final class of events consists of those with little to no characteristics of a wind maxima, or the 0-2 LLJs. Non-Wind Maxima events make up 15 of the 86 cases, with two occurring in May, six in June and seven in July. In some cases, a wind greater than 8 m/s (17.9 mph) and a direction between 90˚ and 270˚ can be seen in at least one level and one time period, but the rest of the event analysis times display no wind maxima. For first the time, the 925 mb recorded the highest wind speed average for at least one time slot (6z (0000LST)). Additionally, the Non-Wind Maxima class also represents the first time where the fastest average wind speeds did not occur at any time slot at 850 mb. In only two cases out of the entire study period do wind maxima occur at two levels (twice at 6z (0000LST), whereas only one time during 18z (1200LST) does any level record a wind speed greater than 8 m/s (17.9 mph). Though substantial veering occurs in the lower two levels from 0z (1800LST) to 12z (0600LST), winds turn back to east at 18z (1200LST)(Table 5.12).

	925 mb Direction	Speed	850 mb Direction	Speed	700 mb Direction	Speed
0z	130	4.5	167	4.6	249	**5.0**
6z	158	**6.3**	173	5.6	225	4.7
12z	190	5.1	208	5.2	263	**5.5**
18z	174	4.5	192	4.3	257	**6.7**

Table 5.12: Average wind characteristics for Non-Wind Maxima. Wind directions are in degrees and on the left, wind speeds in m/s and on the right. Max wind speeds for each time slot are bolded.

Though the Wichita grid cell only represents a small portion of the study area, it nevertheless provides a baseline for characterizing low-level winds during heavy rainfall events recorded at Topeka. Analyzing one grid cell within a study area reveals a number of striking characteristics about the low-level winds during extreme nocturnal rainfall events in northern Kansas for May, June and July. The most noticeable similarities among the three-month is the consistent veering with time and height at all pressure levels and all time slots. In addition, wind speeds reach their maximum at 6z (0000LST) for all three months, followed by decreases in wind speed at 12z (0600LST) and 18z (1200LST). Yet, Whereas these similar patterns exist among the three months, a number of variations are present as well. For one, the strongest and most abundant low-level wind maxima occur during the month of May, with the maximum wind height located at 700 mb. As the warm season progresses, winds decrease on average and the faster winds shift to lower levels of the atmosphere. Given the nature of upper-level winds from May to July, it makes sense the strongest winds exist at the highest levels of the atmosphere in May. Though in the late spring the westerlies continue to weaken and track further northward with the sun's declination, they nevertheless are much stronger in the late spring compared to the mid-summer. As a result, winds at the highest level of the study area would naturally have the most robust winds. Other synoptic influences more than likely influence winds in the low-to-mid levels of the troposphere that cause the low-level winds to extend through a significant portion of the vertical during the early part of the warm season.

The occurrence of low-level wind maxima at 700 mb also decreases from May to July. For May, in 65 cases where the Wichita grid cell has at least one level with wind maxima characteristics at any of the four time slots, 39 times the wind speed occurs at 700 mb, or 60% of the time. Conversely, during June the highest wind speed occurs at 700 mb only 48 out of 122 times (~43%). The percentage decreases even more so in July to 18 out 74 times (~24%). By far 850 mb represents the most common level of the atmosphere for low-level wind maxima during June and July; though stronger winds at 925 mb increase in July. Therefore, low-level wind maxima decreases in altitude as the warm season progresses, suggesting the mechanisms which drive nocturnal LLJs changes from May to July. The fastest winds significantly decrease in June (9.8 m/s in June compared to 12.5 m/s in May). In fact, the average wind at each pressure level at 0z (1800LST) weakens in June. More than likely, May winds have a stronger synoptic rather than nocturnal driver behind their formation and evolution, resulting in stronger and more persistent LLJs.

Additionally, the strength and vertical coverage of May low-level winds exceeds that of June and especially July. With the exception of 925 mb at 18z (1200LST), every level at every time slot averages higher than 8m/s (17.9 mph) during May and June events. In July, however; six out of 16 times wind speeds average below 8m/s. In only six out of the 72 (~8%) total measurements (18 events x 4 time slots) are no southerly wind maxima present for May. In 33 cases out of a possible 156 measurements (39 events x 4 time slots) no southerly wind existed for June events, almost triple the percentage of May (~21%). The percentage increases even more in July as 38 out of 112 (~34%) measurements (28 events x 4 time slots) recorded no southerly winds over 8 m/s (17.9 mph). The average wind speeds also decrease from May to July across the board. May average wind speeds for each level at every time slot exceed that of July. May nocturnal wind maxima exhibit a more eastward component to their direction compared to June and July as well, with the exception of 18z (1200LST) in July where winds shift back in the eastward direction. Excluding the anomaly, all months exhibit a westward veering of the winds from the previous evening into the morning and early afternoon hours the next day.

Although fewer heavy nighttime storms occurred at Topeka during May compared to June and July, the nocturnal southerly winds during May appear to be stronger and longer lasting. Low-level winds were faster at all levels, more prevalent, and had a more southeasterly configuration. Of course, it certainly does not mean low-level wind maxima formation decreases from May to July when heavy rainfalls in north-east Kansas are present. Rather, it may suggest that the wind maxima shift either in latitude or longitude, making it more difficult to detect them at the Wichita grid cell. Only by analyzing the entire study area will a more comprehensive picture of the nocturnal winds during heavy rainfall events in Topeka be unveiled.

The variation in the type of winds present in association with heavy rainfall events at Topeka also stands out as a noticeable feature. The five classes differ significantly in terms of their temporal characteristics, wind speeds and wind directions. Why southerly winds differ dramatically remains a topic for the synoptic mechanisms in the subsequent chapter. However, a complete understanding of the southerly wind phenomenon not only by month but also by class can be gained by analyzing the broader wind characteristics using the geographical study area defined in the methods section. Therefore, the ensuing section analyzes the low-level winds for stretching from 27.5-50N, 252.5-270E. Of particular importance includes the geographical and vertical extent of the winds and wind maxima axis, its location in relation to Topeka, as well as the streamline configuration of the winds from the Gulf of Mexico up to the northern Great Plains.

5.5 Low-level Wind Characteristics by Month:

The low-level winds for the study area of 27.5-50N-252.5-270E are analyzed for all events greater than 1.5 inches (38.1mm). Once again, dividing the events by month provides important climatological and seasonal information about the nature of LLJs from late spring into mid-summer. Commencing with May, low-level winds greater than 8 m/s (17.9 mph) with a direction between 90˚ and 270˚ are present for all 18 events. Maximum winds range from as low as 13.4 m/s (~29.9 mph) to as high as 27.3 m/s (61.1 mph) (Figure 5.01). The maximum winds during May commonly appear at 850 mb, as the highest wind speeds occur 14 out of 18 times at the level. For the other four cases, the maximum winds reside at 700 mb. The highest wind speed reported for all May events (27.3m/s or 61.3 mph) was recorded at 850 mb during the May 26, 1955 event. On average, maximum wind speeds average 20.3 m/s (45.4 mph). In ten instances, the fastest winds occur at 12z (0600LST), Whereas the other eight occur at 6z (0000LST) (Appendix P).

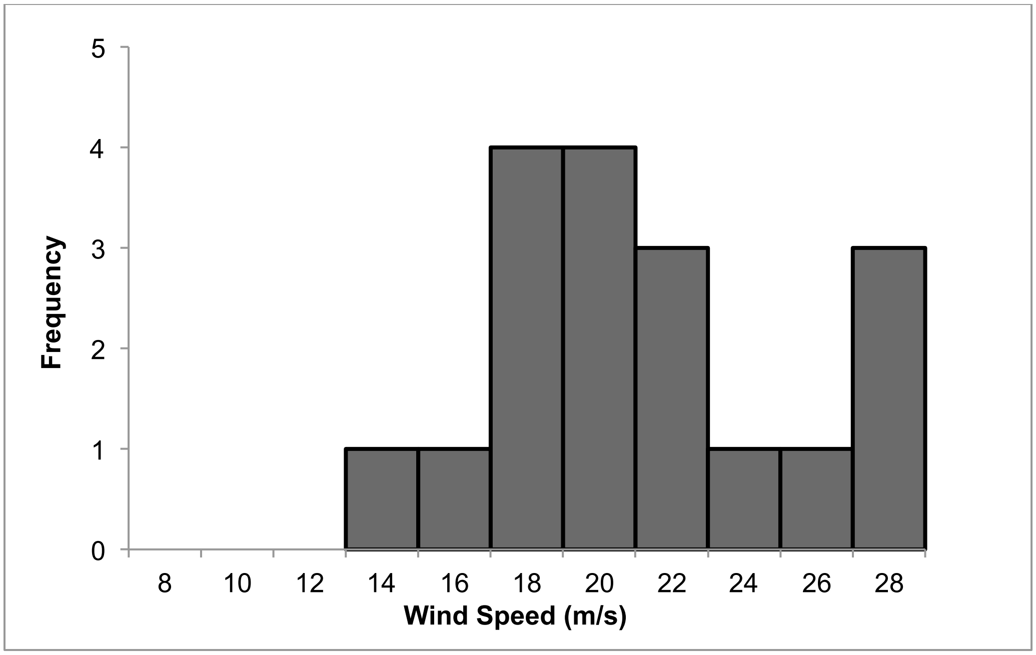
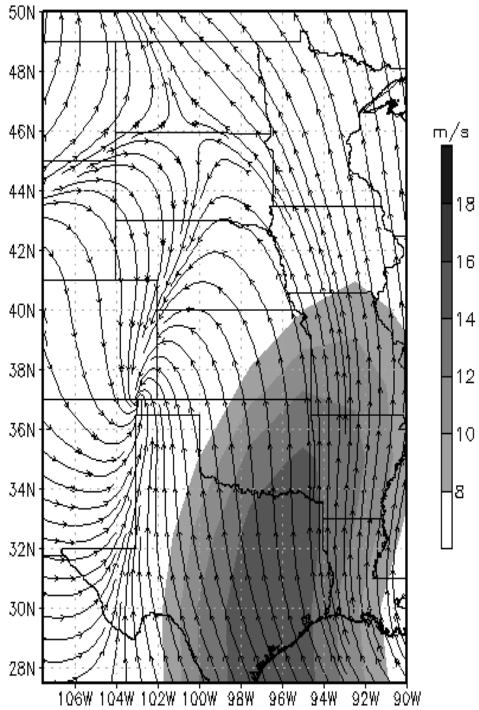
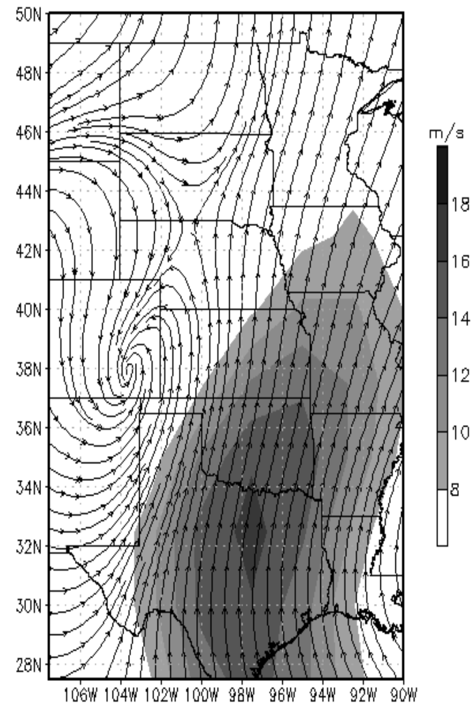


Figure 5.01: Maximum wind speeds for May nocturnal low-level wind maxima.

Though the highest wind speeds can often be seen at 850 mb, wind maxima reside in the other two levels as well, especially from 6z (0000LST) to 12z (0600LST). However, the area of maximum winds at one level rarely exists at the same place in the other two levels. The wind profile of May events has a clear influx of Gulf Moisture, particularly at 925 mb and 850 mb, with the wind maxima axis at 850 mb located in north-central Texas during May events and an average axis location of 35.7˚N-264.9˚E. However, the region of maximum winds at 925 mb and 850 mb differs, as the region of the winds shifts eastward at 925 mb. Regardless, a well-defined area of cyclonic circulation resides in eastern Colorado at 925 mb, along with an area of bifurcating air located to the northeast of the circulation along the Dakota border. At both levels a closed cyclonic circulation can be seen, but the region of bifurcation shifts further southward at 850 mb towards the South Dakota, Nebraska border. Although counter-clockwise circulation dominates the flow pattern at 925 mb from Texas up to the Dakotas, anticyclonic circulation prevails throughout much of the study area at 850 mb. Though, winds located near the center of the cyclonic circulation in Texas, Oklahoma, Kansas, Colorado and New Mexico has a well-defined counterclockwise flow (Figures 5.02 and 5.03).





Figures 5.02 and 5.03: Composite streamlines for May at 925 mb and 850 mb for 6z (0000LST). The shaded contours represent regions of wind speeds >8 m/s (17.9 mph).

The airflow at 700 mb differs from the flow pattern at the two lower levels, hence the reason for separating it from 925 mb and 850 mb discussion. For one, the area of fastest winds shifts westward to central Oklahoma, with a broad area of winds greater than 8 m/s (17.9 mph) stretching from the southern Great Plains up to the Great Lakes region. Wind flows no longer originate from the Gulf of Mexico, but rather are more southwesterly oriented. The 700 mb flow pattern also indicates that a trough in the upper levels digging into the western United States, creating the necessary ingredients for increased southerly flow and moisture influx at the lower levels.

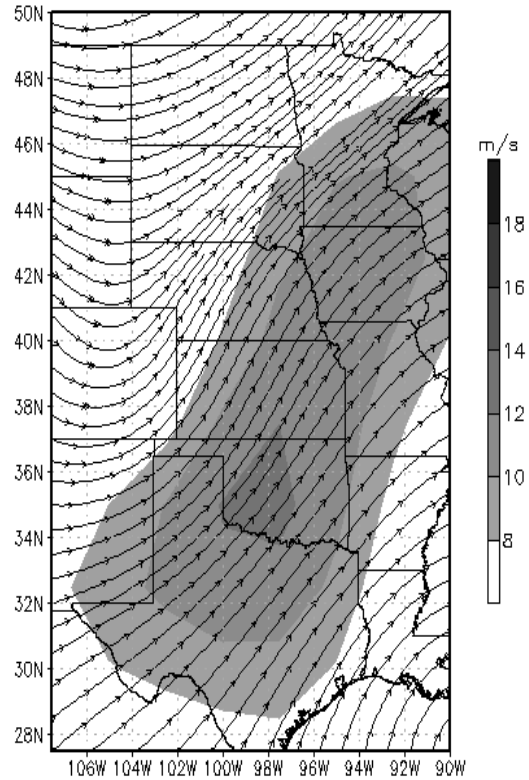


Figure 5.04: Composite streamlines for May at 700 mb at 6z (0000LST). The shaded contours represent regions of wind speeds >8 m/s (17.9 mph).

Convection may form during the nighttime hours in areas of low-level convergence. Therefore, the level of the atmosphere exhibiting a low-level wind maxima profile with a nose-like structure at the northern end of the wind maxima more than likely creates an initiation point for producing nighttime rainfall. Interestingly, a convergence zone appears to be located west of the wind maximum at 925 mb along the New Mexico-Texas border near the circulation center. An area of streamlines very close together east of the wind maxima region stretches from the Gulf Coast up the Upper Midwest at 925 mb as well, suggesting the greatest moisture influx occurs east of the wind max at the lowest levels. Interestingly, the region cannot be detected at 850 mb. With the average location of the wind maxima's nose close to Topeka at 6z (0000LST), it makes sense for heavy rainfall events to occur in the region. In 13 out of 18 May nocturnal storms, Topeka resides to the north or northwest of the wind maxima axis, followed by three cases south or southwest of the jet axis and finally two cases where the wind maxima actually occurs to north or northeast of Topeka.

For June, 850 mb continues to be the most favorable level for the wind maxima axis, though its axis occurs much more frequently at lower levels in June. Unlike May, maximum winds within the study area take place at 925 mb 11 times. Moreover, only once do the highest wind speeds occur at 700 mb. Though, average maximum wind speeds decrease from May to June, as in May winds average over 20m/s, whereas the fastest winds in June average 17.8 m/s. Jet maxes ranged from 9.7 m/s (21.7 mph) to as high as 27.3 m/s (61 mph) (Figure 5.05). Somewhat surprisingly, though the low-level winds in May reach their peak closer to 12z (0600LST) than the 6z (0000LST) hour, in June the opposite happens. In fact, 25 out of the 38 possible maximum winds occur at 6z (0000LST), with only 11 recorded at 12z (0600LST) (Appendix Q). Areas of maximum winds also shift southwesterly in June, with the average location of the axis found at 34.3˚N-97.5˚W.

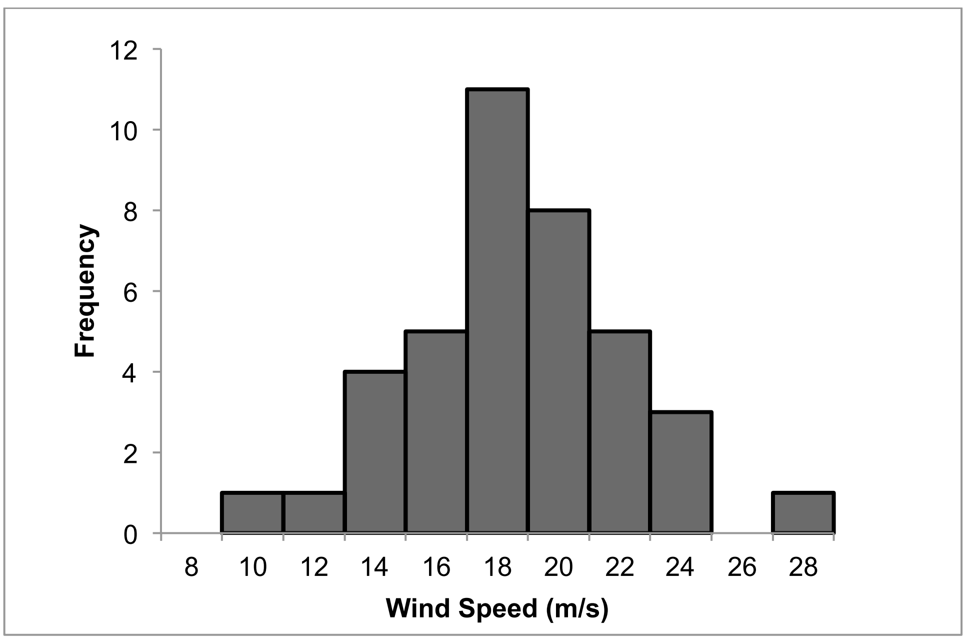
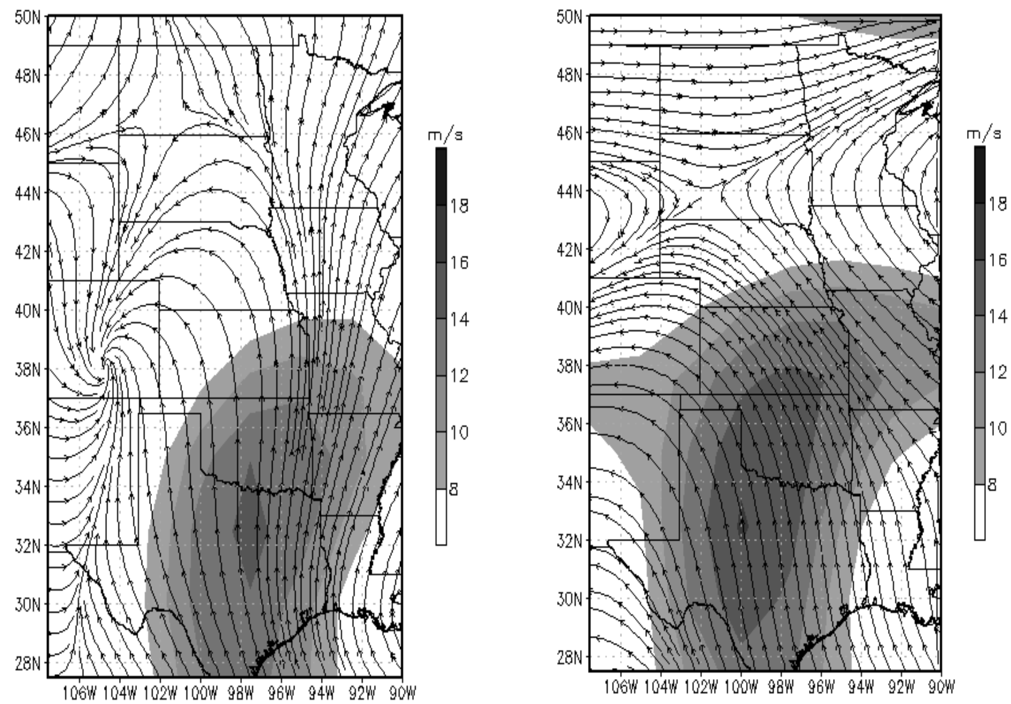


Figure 5.05: Maximum wind speeds for June nocturnal rainfall events.

The wind flow pattern at 925 mb and 850 mb for June heavy rainfall events share a number of similar characteristics to May. The presence of a low pressure region at 925 mb continues once again in June with a closed cyclonic circulation in eastern Colorado. At 850 mb, however; a cylonic circulation center cannot be detected, but winds flow from a southeasterly direction. Bifurcating regions of air are found in the same areas at 925 mb and 850 mb during June. The region of maximum winds in the lower levels continues to be located in the southern Great Plains, though it shifts westward (particularly at 850 mb). The region of maximum winds at 925 mb is once again located to the east of the wind maxima axis at 850 mb. The tightly packed streamlines just east of the jet axis at 925 mb are seen in June as well, stretching from the

Gulf of Mexico up to Iowa before eventually diverging in southern Minnesota. Whereas northeast Kansas lay to the northwest of the wind maxima axis in May, in June the axis shifts westward, thus making Topeka on average situated north to northeast of the jet axis. Interestingly, Whereas anticyclonic flow dominated the pattern at 850 mb in May, cyclonic circulation stretching from the Gulf of Mexico up to the South Dakota region persists in June. A split-like flow pattern persists at 925 mb, with southerly-to-anticyclonic winds present east of the Missouri River Whereas cyclonic flow persists in the western part of the study area (Figures 5.06 and 5.07).



Figures 5.06 and 5.07: Composite streamlines for June events at 925 mb and 850 mb at 6z (0000LST). The shaded contours represent regions of wind speeds >8 m/s (17.9 mph).

The wind flow at 700 mb once again differs greatly from the lower two levels for June events. Although the air flow configurations remain relatively the same in June compared to May, the winds weaken and shrink significantly. Though it appears some of the wind flows originate out of the Gulf Mexico, they turn anticyclonically well south of Kansas. North of the anticyclonic pattern westerly winds dominate the geographic region, with what appears to be slight troughing in Colorado and New Mexico (Figure 5.08).

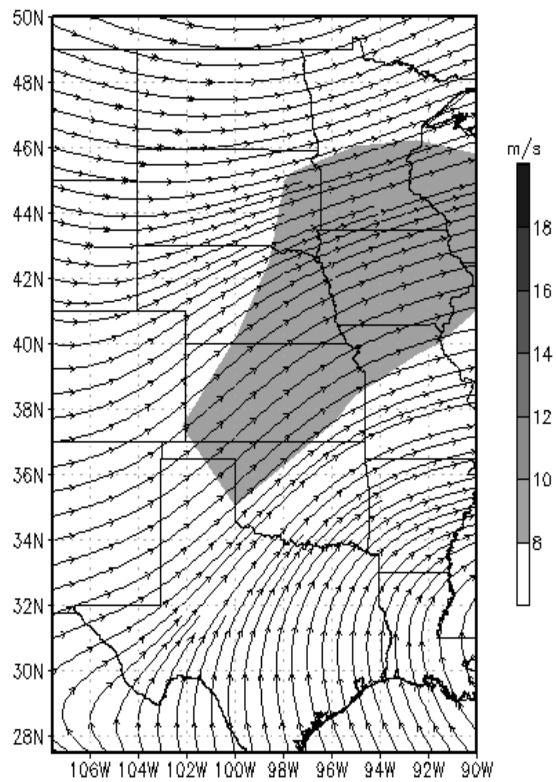


Figure 5.08: Composite streamlines at 700 mb for June at 6z (0000LST). The shaded contours represent regions of wind speeds >8 m/s (17.9 mph).

The trend of lower jet heights and lower maximum wind speeds continues for July nocturnal rainfall events, as the maximum wind speed for July events peaks at 25 m/s (55.9 mph) for the July 16, 1950 event. July also continues the trend of earlier peak jet times, with 23 out of 29 wind maxima recorded at 6z (0000LST) (Appendix R). In July, however; a number of events display little to no jet profile. In fact, in five out of the 29 events, low-level winds essentially do not exist from 0z (1800LST) to 18z (1200LST) at all three levels. Regardless, the southwestward shift of jet maxima continues into July, with the average location of the wind maxima axis located at 32.6˚N-261˚W (Figure 5.09).

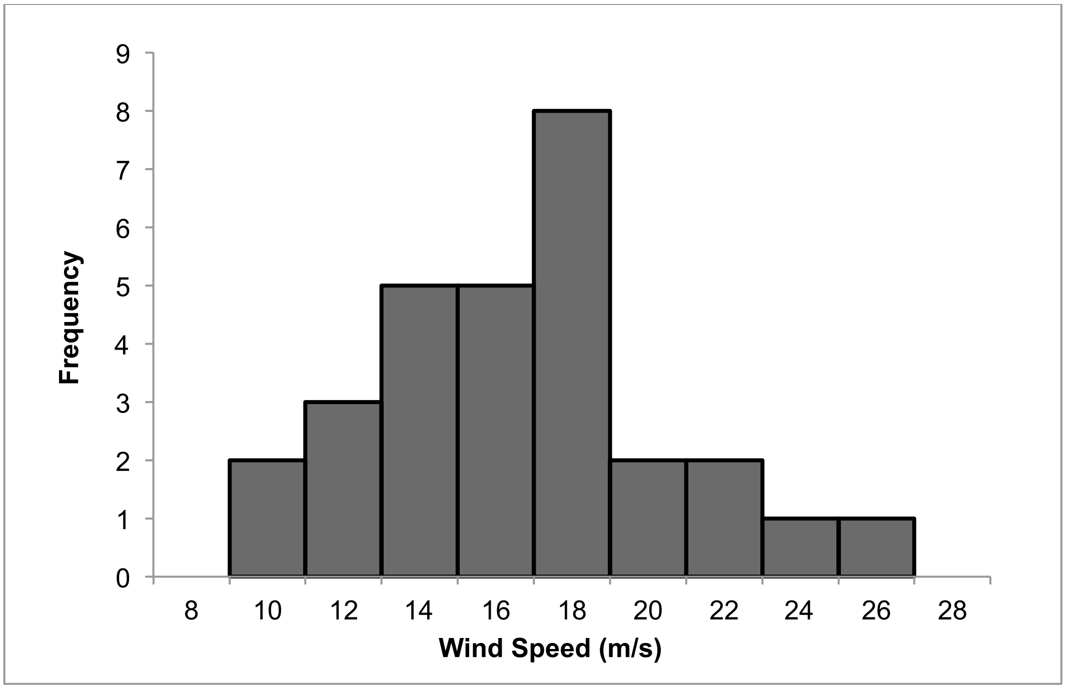
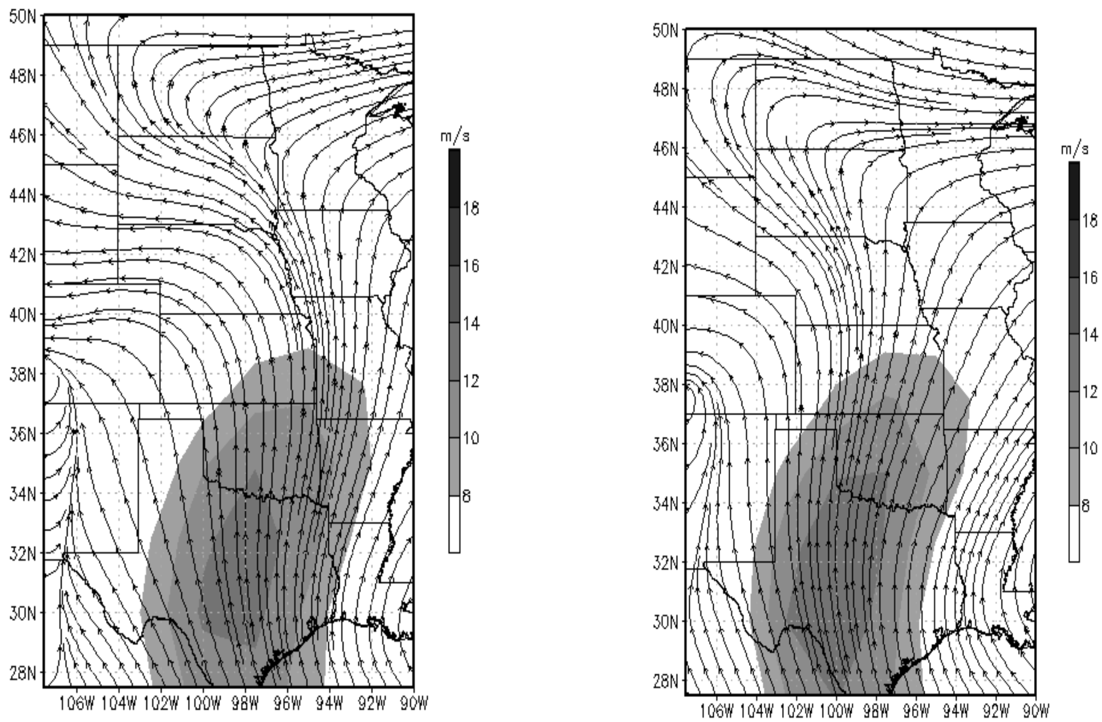


Figure 5.09: Histogram of maximum wind speeeds for July events.

The airflow configurations at 925 mb and 850 mb at 6z (0000LST) change significantly from May and June to July, with wind speeds throughout the study decreasing significantly. The closed circulation present in May and June all but disappears in July, perhaps shifting westward out of the study area as indicated by tight streamlines oriented cyclonically in western Colorado. Though, in areas of New Mexico a clear south-north convergence zone can be seen, an indication that some type of circulation exists in the western periphery of the study area. The region of max moisture influx continues to ride along the eastern fringe of the wind maxima at 925 mb, but a second area of confluence can be seen flowing at the center of the wind maxima. North of the jet axis, an area of divergence exists at both levels, situated along the border of South Dakota and Minnesota at 925 mb, and along the South Dakota-Nebraska border at 850 mb. July low-levels winds also exhibit a very pronounced split-like flow. Winds in the Southern Plains have a south-north streamline orientation, with perhaps slight anticyclonic circulation at 850 mb. However, on the eastern and western portions of the streamlines, the winds turn cyclonically in the west and anticyclonically to the east in the central Great Plains. At the center, winds remain oriented south-north. The anticyclonic turn of the winds in the east suggests the presence of high-pressure region in the southeast United States. More than likely the anticyclonic airflow configuration indicates the presence of Bermuda or Azores high-pressure cell that dominates the north Atlantic circulation during the summer. The Bermuda High brings in Gulf Moisture to the central United States allowing for nighttime convection during the summer months. The slowing down or convergence of winds once again occurs close to Topeka, suggesting that winds converge at north-central or northeast Kansas causing uplift and subsequently nighttime convection (Figure 5.10 and 5.11).



Figures 5.10 and 5.11: Composite streamlines at 925 mb and 850 mb for July at 6z (0000LST). The shaded contours represent regions of wind speeds >8 m/s (17.9 mph).

The decrease in low-to-mid-level winds persists in July, though the streamlines share a similar configuration to the other months. In no portion of the study area do winds exceed 8 m/s (17.9 mph). Perhaps the largest difference in July from May and June is the northwestward progression of the large anticyclonic circulation present in the southeastern United States and Gulf of Mexico. Once again, this more than likely indicates the presence of the Bermuda High (Figure 5.12).

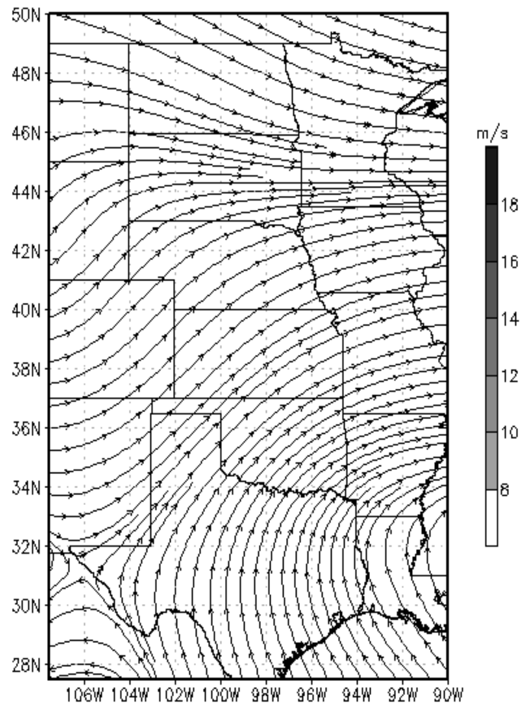


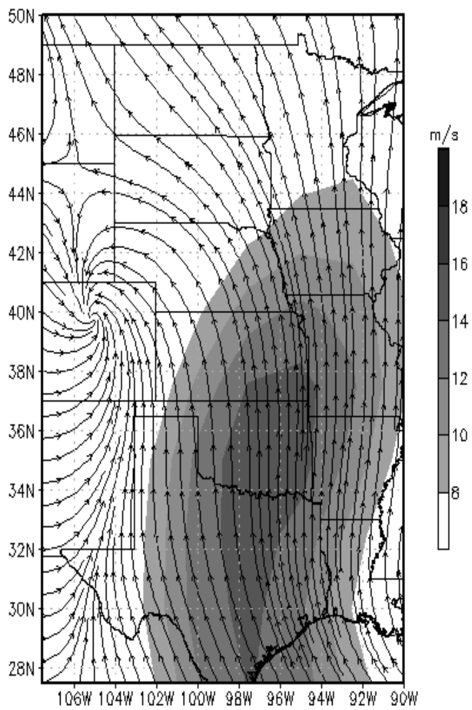
Figure 5.12: Composite streamlines at 700 mb for July at 6z (0000LST). The shaded contours represent regions of wind speeds >8 m/s (17.9 mph).

5.6 Low-level Wind Characteristics by Class:

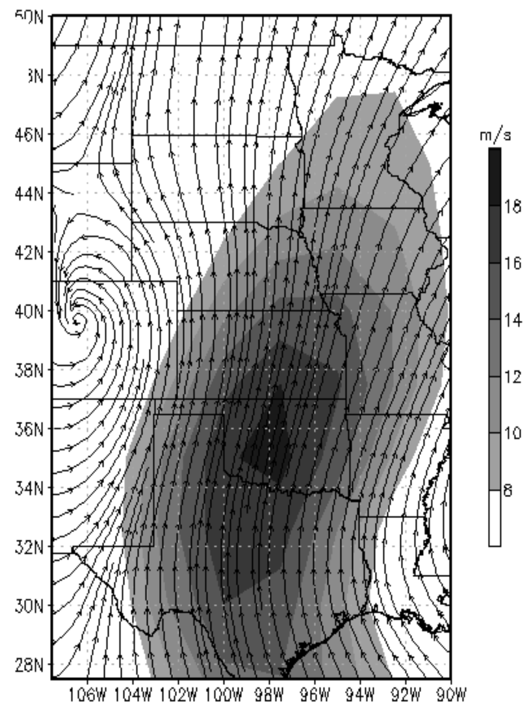
Along with understanding the low-level winds on a monthly basis, visualizing the phenomena based on its strength and intensity from data at the Wichita grid cell can also provide a glimpse into the diverse nature of the southerly low-level winds. The winds can be broken down into 5 LLJ classes as discussed in a previous section of this chapter. The following section analyzes the spatial and vertical characteristics of the five classes of jets for the 6z (0000LST) hour.

Beginning the first class, the "Robust Class", wind maxima exist at all levels of the atmosphere for all four time slots at the Wichita grid cell during these events. The streamline configuration for Robust events show a large area of wind maxima in the lower levels of the atmosphere, especially at 850 mb where wind averages reach as high as 18 m/s (40.3 mph) in areas of central Oklahoma and into south-central Kansas (Figure 5.13). The expansiveness of the winds for Robust events exceeds that of any other class, extending from the Gulf of Mexico up to Minnesota at 850 mb. A multi-level closed cyclonic circulation resides in north-central Colorado, along with an area of bifurcation to the north of the circulation, and an area of convergence along the New Mexico-Texas border. Though both areas show similar characteristics regarding the cyclonic circulation, their streamline orientation through the Great Plains differs. At 925 mb, winds from the southern Great Plains into Kansas turn cyclonically with latitude, turning into the area of low pressure in Colorado. The winds at 850 mb, however; veer anticyclonically with an increase in latitude. In relation to northeast Kansas, at 6z (0000LST) the wind maxima axis actually sits just to the south of Topeka. At 925 mb, Topeka still lies within the wind maxima region, but the region of max winds shifts southeast. The airflow configuration at 700 mb differs from that of the lower levels, but closely represents the flow present during heavy rainfall events in May. An expansive region of southwest-northeast oriented winds greater than 8 m/s (17.9 mph) extends from New Mexico up to the Great Lakes region. Yet, within that region centered on northern Texas, central Oklahoma and southern Kansas, a convergence zone exists where winds from the southwest meet winds from the Gulf of Mexico (Figure 5.13).

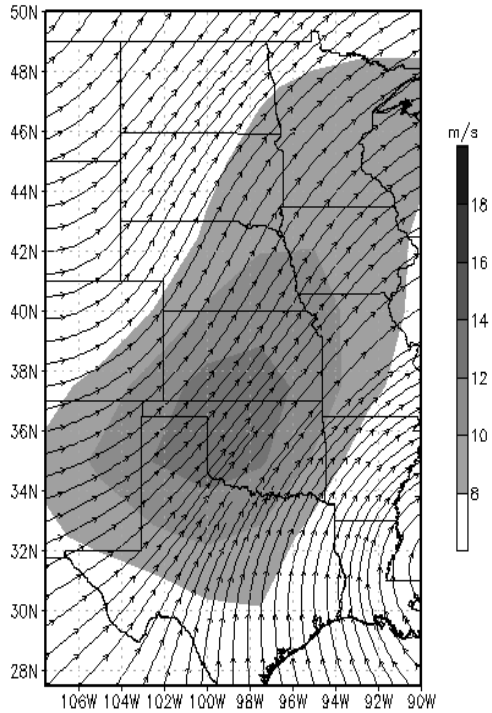
A.)



B.)



C.)



Figures 5.13: Streamline composites for Class One winds for 6z (0000LST). A.) 925 mb composite, B.) 850 mb composite, and C.) 700 mb composite. Shaded regions represent areas where winds >8 m/s (17.9 mph).

The second class of low-level winds, the Semi-Robust class, weakens in strength and horizontal convergence compared to the previous class, but nevertheless shares similar characteristics. The region of maximum winds continues to reside in Texas and Oklahoma, however; the wind maxima no longer stretches up to the Great Lakes region, rather the area of winds greater than 8 m/s (17.9 mph) ceases in northern Kansas and into southern Nebraska. Although a closed cyclonic circulation can be seen at both levels, it shifts southward compared to the first class of LLJs. Moreover, a bifurcation zone no longer exists to the north of the circulation. Interestingly, at 850 mb a region of tightly packed streamlines stretches from the Gulf of Mexico into Kansas and Nebraska. However, in northern Kansas and Nebraska the winds diverge, either turning cyclonically into the low pressure region or continue flowing north before turning anticyclonically in the upper Great Plains. Winds similarly turn anticyclonically in northern regions of the study area at 925 mb, but the streamlines from Texas into Nebraska turn cyclonically into southern Colorado. A west-east convergence zone also exists at 925 mb, stretching from Wyoming into southeast South Dakota (Figure 5.14).

Not surprisingly, the strength and geographical coverage of winds at 700 mb weaken and shrink for Class Two winds, though the streamlines share similar configurations to the first class of winds. The fastest region of winds once again exists over Kansas. The most noticeable feature at 700 mb is the closed anticyclonic circulation residing over the southeast centered in southern Arkansas. The feature may help bring in moisture to the mid-levels of the atmosphere and create a southwest-northeast convergence zone centered on Kansas (Figure 5.14).

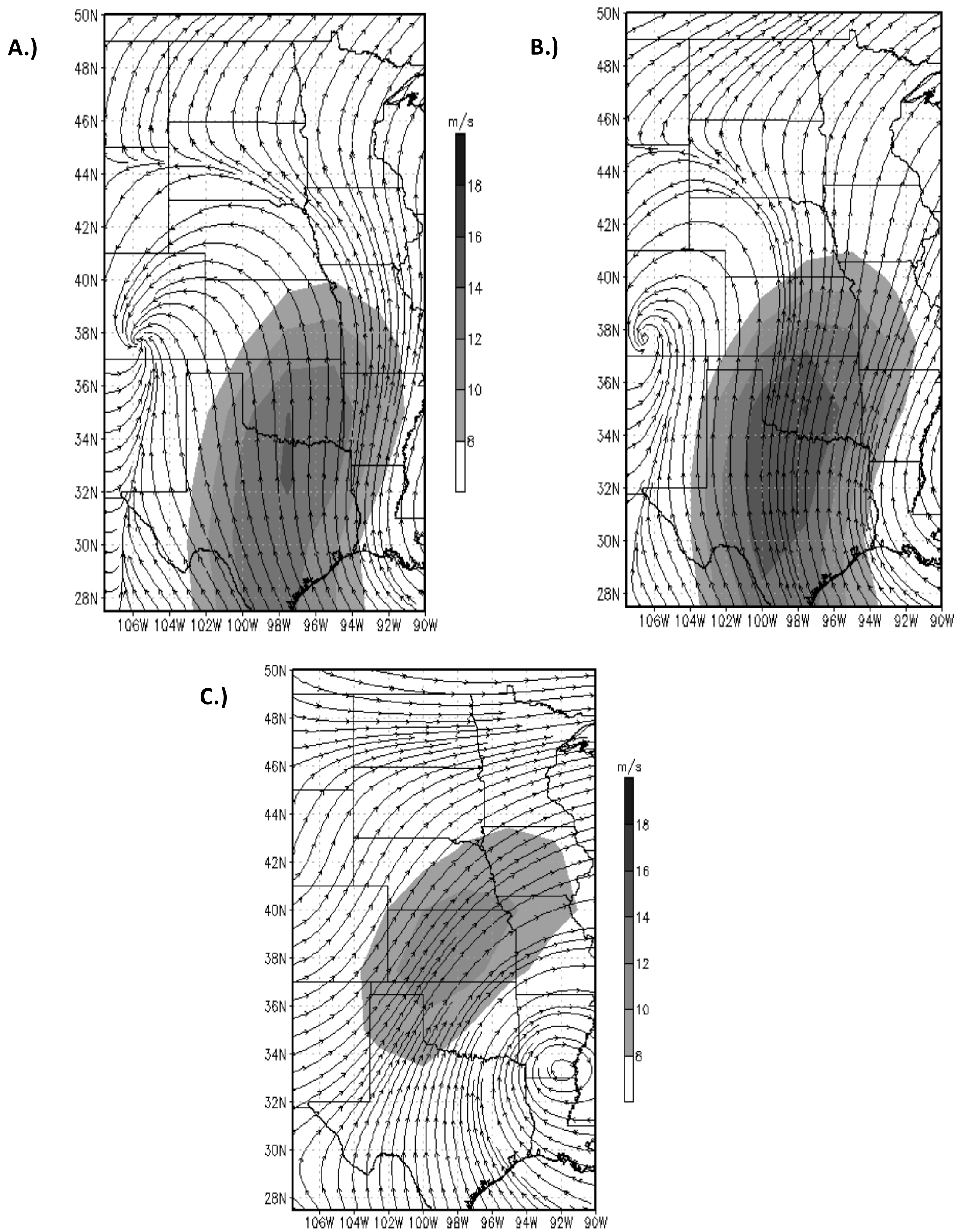
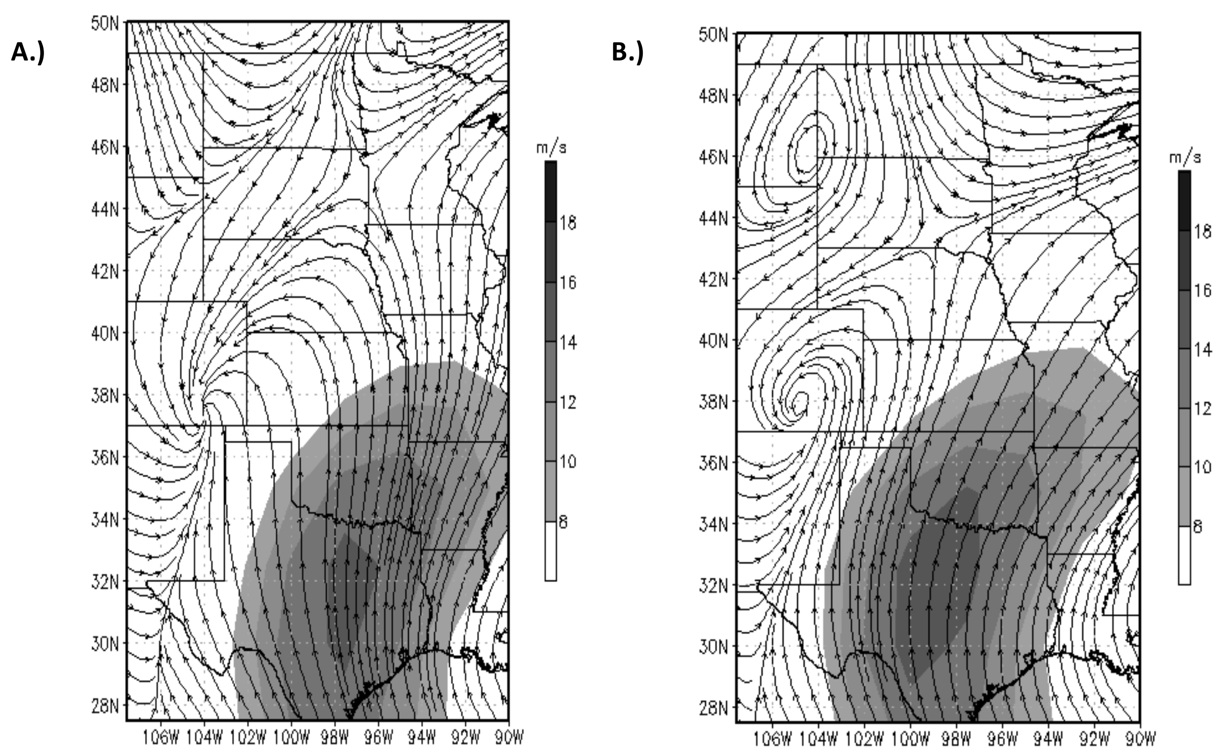


Figure 5.14: Same as 5.13 for Class Two "Semi Robust" winds at 6z (0000LST).

The third class differs substantially from the previous two in terms of its streamline configurations. For the first time, the vertical region with maximum winds occurs not only at 850 mb, but at 925 mb as well. Yet, the wind maxima axis at 850 mb covers a larger spatial area than 925 mb. However, the axis shifts well south compared to the previous first two classes, now occupying areas of southern and central Texas and extending up into central Kansas. Unlike the first two classes, Topeka now lies on the boundary oriented southwest to northeast of areas with wind maxima and those areas not fitting the wind maxima criteria. The cyclonic circulation occurs once again at both levels, though it is better defined at 850 mb compared to 925 mb. A second area of closed cyclonic circulation also resides in eastern Montana and the Dakotas at 850 mb, accompanied by a convergence zone in the same region at 925 mb. Two bifurcation zones are located at both levels as well, both located in central and eastern South Dakota at 925 mb and 850 mb. The winds at 925 mb from the southern to central Great Plains largely turn cyclonically with latitude; however winds to the east of the wind maxima axis flow from a southwesterly direction from the Gulf of Mexico up to the Upper Midwest. Winds at 700 mb continue their decrease for the third class, but troughing persists in Colorado and New Mexico (Figure 5.15).



C.)

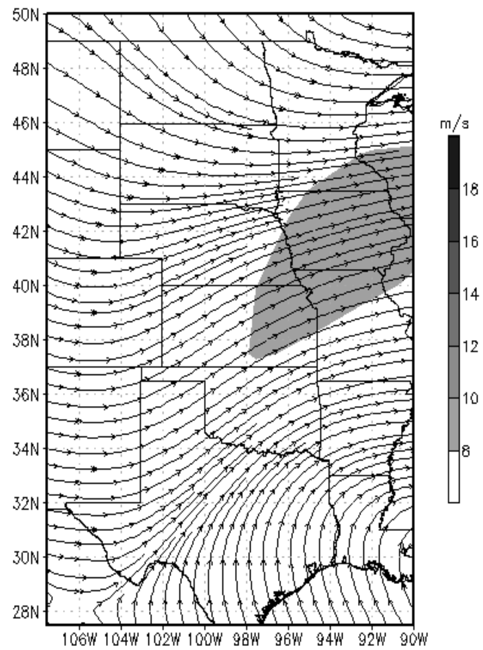


Figure 5.15: Same as 5.13 for Class Three (6-8) for 6z (0000LST).

For the fourth class of LLJs (3-5 class), winds at both levels weaken, along with the wind maxima continuing its westward shift at 850 mb. The region encompassing winds with wind maxima characteristics shrinks once again from the previous class, now only reaching the southern Kansas region at its northward extent. The Topeka area no longer resides in areas where wind speeds reach 8 m/s (17.9 mph) or higher, now located to north and northwest of the LLJ region. The cyclonic circulation in Colorado moves westward for Class Four winds, as do the two bifurcation zones at both levels. Just to the northwest of the 925 mb bifurcation zone a second closed cyclonic circulation exists. The dominant flow pattern from the Gulf of Mexico occurs at 925 mb with a south-north streamline from the Gulf of Mexico up to Kansas, suggesting that the moisture advection needed for nighttime convection in Kansas occurs primarily at 925 mb. Little change occurs at 700 mb, with the perhaps the only difference being a better-defined convergence zone present along the Nebraska-South Dakota border (Figure 5.16).

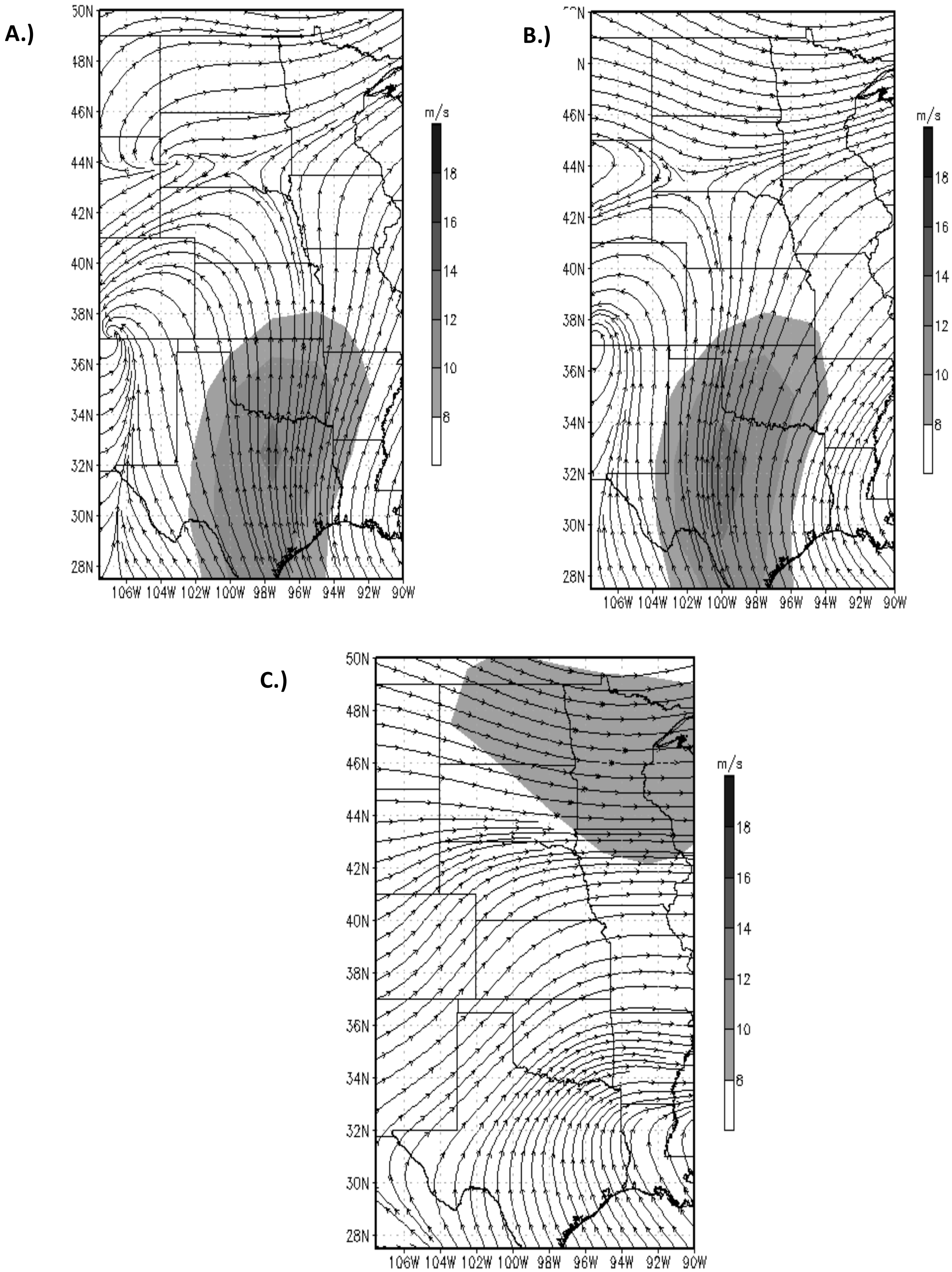
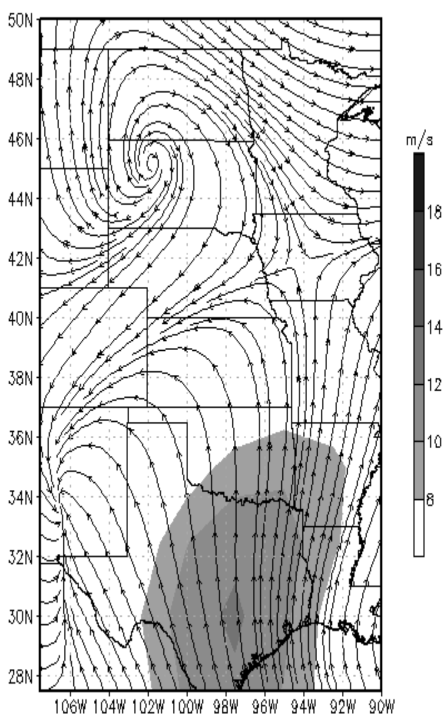


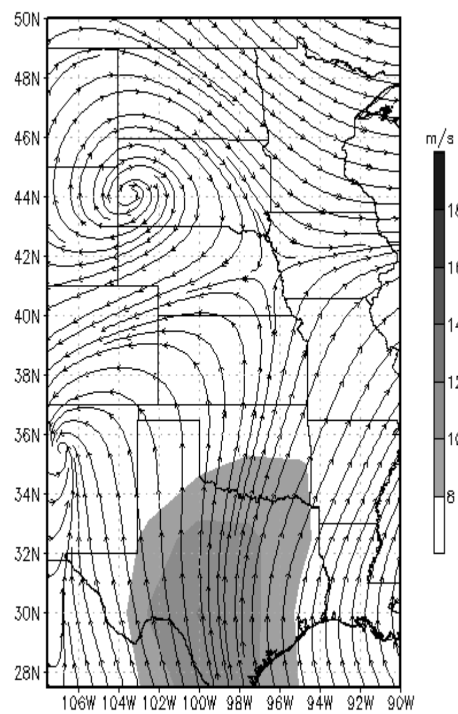
Figure 5.16: Same as 5.13 for Class Four winds (3-5) for 6z (0000LST).

The final class of winds consists of those that exhibit little to no wind maxima qualities for any level at any time slot. Spatially, the low-level winds contain a number of features that greatly differ from all other wind classes. To begin, the spatial extent of the winds reaches its minimum for the five classes, only extending as far as southern Oklahoma at 850b. As a result, Kansas and much of the central Great Plains lie well north of the wind maxima axis. Additionally, a large closed anticyclonic circulation resides over the western Great Plains in South Dakota at both 925 mb and 850 mb. A second area of cyclonic circulation also exists in north-central New Mexico, creating a south-north convergence zone in southern New Mexico. The presence of the two features appears to result in a bifurcation zone in southeast Nebraska at both levels as well as a shift in the southerly-oriented winds in the southern Great Plains, to eastern winds in much of Kansas at both levels. Some streamlines, however; continue to flow northward before eventually turning anticylonically at the bifurcation zone. Even the pattern at 700 mb differs from previous classes, with a slightly amplified trough pattern in Colorado and New Mexico, as well as a bifurcation zone to the south along the Texas-Mexico border (Figures 5.17). Given the location of the wind maxima axis to Topeka, the question remains just how much influence the fifth class of winds has on heavy rainfall events in Topeka. Perhaps others mechanisms, such as the influence of mid-to-upper level synoptic systems or the presence of both an anticylonic and cyclonic circulation to the west in the lower levels play a larger role in generating deep convection in the nighttime hours for the fifth class of winds.

A.)



B.)



C.)

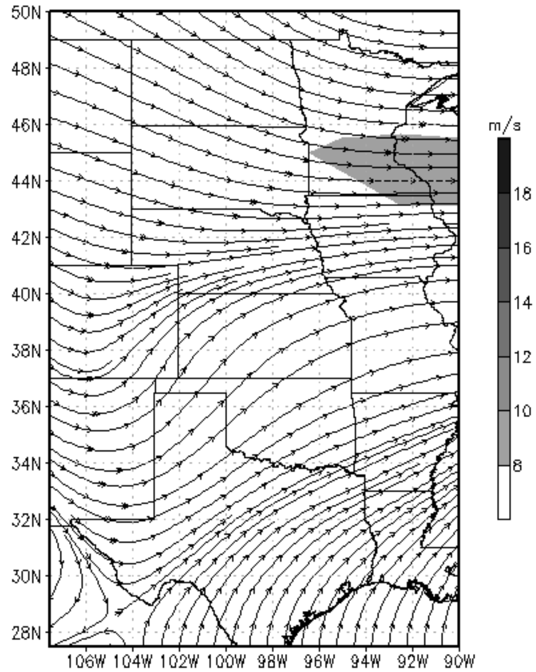


Figure 5.17: Same as 5.13 for Class Five winds (0-2) for 6z (0000LST).

Broken down by class, the streamline configurations of the low-level winds have an incredibly diverse nature when associated with nighttime precipitation. Streamline orientations differ significantly by class, by level, as well as by their location within the geographic study area. Similar to the streamline composites of each month, 850 mb represents the portion of the atmosphere where the axis of the low-level wind maxima resides. Nevertheless, based on the streamline patterns, 925 mb appears to be the level where the largest fluxes of moisture into the central Great Plains occur. Whether the two levels represent multiple airstreams within the wind maxima or simply represent an increase in wind speed with height before decreasing again at 700 mb remains a question for further analysis. Also noticeable from the composites is the diverse nature of the airflow configurations throughout the study region. Areas in the eastern portion tend to be more anticyclonically oriented, Whereas areas in the west turn more cyclonically with latitude as a result of the persistent cyclonic circulation present in Colorado. The changes in airflow with latitude resulted in areas of frequent bifurcation zones present throughout the northern Great Plains. Though, of most importance in the streamline analysis is how the wind flow at the lower levels relates to Topeka. Unsurprisingly, Topeka was frequently located to the north of the wind maxima axis. Depending on the class, however; Topeka was either close to the jet axis (jet axis located in northern Oklahoma or southern Kansas), or hundreds of miles away in southern Oklahoma or Texas. Therefore, the mechanisms needed for a rare heavy precipitation event greater than 1.5 inches at Topeka varies incredibly both vertically and spatially.

5.7 Discussion:

It has been long accepted that the influx moisture from the Gulf of Mexico brought by the LLJ provides the necessary ingredients to generate nighttime convection in the central Great Plains. The results of this chapter further strengthen this concept, as in the majority of the heavy rainfall events analyzed a low-level wind maxima accompanied them. Additionally, this analysis of the southerly low-level winds provides a further understanding into its nature and evolution. At the same time, however; the strength of the low-level winds does not correlate with the strength of a heavy rainfall event. Low-level winds of all strengths and geographic ranges can generate a heavy (>1.5in) rainfall event at a particular location. In 86 cases analyzed, low-level winds could be broken down into five distinct categories; relatively evenly distributed among the categories. In fact, in some instances low-level winds in the study region had little to no wind maxima-like characteristics, further strengthening the concept that other mechanisms must be in play to generate heavy nighttime convection Regardless, whether by analyzing data from one grid cell or across an expansive region, low-level winds vary incredibly temporally, spatially, and vertically.

Though within this variation, it nevertheless is obvious that the southerly winds in the southern and central Great Plains reaches its peak during the hours of 6z (0000LST) and 12z (0600LST). The streamline composites also convey the idea that southerly low-level wind maxima tend to be limited to regions of the vertical below 700 mb. In every streamline composite, 700 mb winds contain a westerly component more indicative of the mid-upper level westerly wind patterns. Unfortunately due to the vertical resolution of the data, it remains unseen at what pressure height winds switch from a “Great Plains Low-Level Wind Maxima” to the upper-level patterns seen at higher levels. Nevertheless, each level of the vertical has distinct characteristics in terms of its wind speeds and locations of wind maxima, with a general anticyclonic turning of the winds with height. Previous research has suggested the idea that multiple airstreams can exist at different levels, especially when associated with mid-latitude cyclones (Bonner, 1966; Dickinson and Neumann, 1982; Kontroni and Lagouvardos, 1993). Winkler and Walters (2001a) validated the hypothesis with their climatological study where they discovered 12 different jet types with various airflow configurations. In a number of the jet types, the authors discovered multiple air streams with the LLJ characteristics and configuration in the vertical. In particular, the study of Walters and Winkler (2001a and 2001b) most closely relates to the methods implemented in this research, making it useful for a cross-study comparison. From their research, Walters and Winkler (2001a) discovered 12 different types of LLJs with various geographical and vertical extents, as well as regions of convergence, divergence, confluence, diffluence and bifurcation. In general, their results show that a number of jet axes occurred much further northward compared to the jet axes in this research. The region of fastest winds tended to be along areas of significant confluence or deformation zones. Though, both studies display that the geographical distribution of jet axes vary greatly. Moreover, the streamline configurations in their research share a number of similar traits to the streamlines analyzed here, particularly their classes referred to as "Cyclonically curved wind maxima (Cc [closed cyclonic] and LCc [Long wind maxima with cyclonic flow and confluence] jet types)," and "Bifurcating wind maxima." An unexpected finding of both studies is the considerable number of jets with cyclonic as opposed to anticyclonic flow. Only 37% of the events analyzed in their study were anticyclonically curving jets. However, the flow at 850 mb in the eastern regions and particularly 700 mb in this research more resembles anticyclonic flow. Similar to Walters and Winkler (2001a), more often than not a south-north convergence zone was set up in the western boundary of the study area. Perhaps the most glaring similarity between the two studies is how many different types of LLJs exist depending on the month analyzed. Walters and Winkler (2001a) were able to associate jet types with their month of most frequent occurrence, reinforcing the idea that forcing mechanisms change by month and consequently change the orientation, strength, and location of the LLJ.

As to what level, what time, and what type of the southerly wind maxima contributes most to heavy rainfall events in northeast Kansas remains an important question for further analysis. Nevertheless, the many of types of LLJs in the Great Plains raises questions on what causes its variation throughout the late spring-early summer precipitation maximum. It must be kept in mind, however; that the LLJ is only one ingredient needed to generate nighttime convection in the central Great Plains. Only by analyzing the entire synoptic picture will a more comprehensive understanding be obtained about the factors contributing to nighttime rainfall in northeast Kansas, as well as the LLJ and its variation. The following chapter delves into the synoptic environment, analyzing the circulation patterns in place during nighttime rainfall events recorded at Topeka.

CHAPTER SIX
THE SYNOPTIC ENVIRONMENT RELATED TO THE HEAVY RAINFALL EVENTS AT TOPEKA

"Realize that everything connects to everything else."

-Leonardo da Vinci

Though southerly wind maxima play a vital role in the development and maintenance of heavy rainfall events in the central Great Plains, it is important to remember that rainfall in the region can result from other mechanisms in the atmosphere. Moreover, the formation and evolution of the nocturnal low-level winds also results from varying atmospheric circulation patterns. It has been well documented that deep convection is often associated with environments which promote upward vertical motion such as upper-level disturbances and mid-latitude cyclones (Carleton et al. 2008). During the warm season upper-tropospheric dynamic features such as divergence and vorticity advection, as well as surface features such as cold fronts, warm fronts and surface cyclones contribute to the central Great Plains hydroclimate. Coupled with the analysis of the low-level winds in the southern Great Plains, analyzing atmospheric circulation patterns can enable a more comprehensive understanding to the generators of nighttime convection during the late spring-early summer precipitation maximum in the central United States.

As the northern hemisphere rapidly progresses from spring in the summer with the increasing northerly latitude of the sun's declination, the upper-level circulation patterns change drastically over the United States. The average circulation patterns in place in May share little to no similarities to the patterns in place during July (Harman 1993). In fact, the transition period from late spring into early summer results in changing circulation pattern not only on a monthly basis, but on a week-to-week basis as well. As has been done in previous chapters, circulation patterns will once again be broken down into temporal subsets. Given the rapid changes in circulation from May to July; however; the analysis and summarization is done on a half-monthly basis. Further explanation of the methods implemented in this chapter can be found in the Methods chapter. The first section of this chapter delves into the statistics of upper-level and surface patterns for each bi-weekly section, analyzing the larger-scale synoptics of the contiguous United States and the frontal boundaries moving through the central Great Plains during nighttime rainfall events at Topeka.

6.1 Statistical Analysis:

The statistical analysis of upper-level and surface patterns gives a glimpse into the common modes of circulation that generate nighttime convection. For each half-month period, 40 random events were analyzed from the evening (0z or 1800LST) prior to the rainfall being recorded at Topeka giving a total of 240 events analyzed. The upper-level patterns for the study area were then placed into six different categories based on their pattern and orientation. By far, the most common circulation patterns responsible for generating nighttime convection in the Great Plains was either a ridge in place in the western U.S followed by a downstream trough in the eastern U.S, or the opposite pattern. In the case of the former, the upper-air pattern above Kansas was often dominated by winds out of the northwest, whereas in the trough-ridge (TR) pattern southwesterly flow prevailed. Interestingly, the trough-ridge (TR) pattern appears to be most common in the late spring, whereas the ridge-trough pattern dominates the flow pattern in the mid-summer months. A closed upper-level low-pressure system affecting an area at or near Kansas was a feature common only in the late spring months, whereas the trough-ridge-trough pattern across the US appears more as a summertime pattern. The presence of closed-lows only and the tendency for pools of cold air to cut-off in late spring makes sense given the further southward movement of storm systems and the position of the westerlies earlier in the warm season. The three least common modes, zonal, trough and trough-zonal (TZ), are relatively evenly distributed from May to July, with perhaps a slight favoring for May and June (Table 6.01).

	May 1-15	May 16-31	June 1-15	June 16-30	July 1-15	July 16-31	Total
Trough-Ridge	13	8	12	**16**	8	4	61
Ridge-Trough	5	4	6	6	14	**24**	59
Trough-Ridge-Trough	0	4	6	8	**10**	9	37
Cut-off-Low	8	**12**	4	0	0	0	24
Zonal	4	4	5	**7**	1	2	23
Trough	3	**6**	2	3	5	1	20
Trough-Zonal	**5**	2	**5**	2	2	0	16

Table 6.01: Common modes of upper-level circulation on a half-monthly basis.

An additional statistical analysis was performed on features at the surface, most importantly the placement of frontal boundaries related to northeast Kansas. Throughout the year, frontal boundaries assist in providing the necessary low-level convergence and other ingredients to create an unstable atmosphere and precipitation in the Great Plains. Though more prevalent in the cooler season, cold fronts pushing their way into the central Great Plains are often responsible for promoting convection both in the day and nighttime hours. Understanding the placement and orientation of these features related to northeast Kansas allows for a comprehensive picture of the mechanisms associated with nighttime rainfall from the surface to the mid- to upper-levels of the troposphere. Listed below is the frequency frontal boundaries, frontal types, and where precipitation occurs at Topeka in relation to the boundary (Table 6.02). In general, most nocturnal precipitation is in some way related to a passing frontal boundary in Kansas, particularly during May. As spring turns into summer, the frequency of non-frontal related precipitation events increases, but nevertheless the majority of nocturnal events in June and July are associated with a frontal boundary. However, the types of fronts moving through the area change based on the time of the year. During the late spring, fronts with a west-east orientation commonly move through Kansas, Whereas come summer the southwest-northeast oriented fronts are more common. Interestingly, the south-north type cold front, or the cold fronts moving from west to east, essentially ceases come the beginning of summer. In a small number of cases, a frontal boundary had a northwest-southeast orientation (NW-SE). Similarly, a few times during each month a warm front was the only detectable boundary on the weather map.

	May 1-15	**May15-31**	**June 1-15**	**June16-30**	**July 1-15**	**July 16-31**
# of fronts	40	39	36	38	32	36
S-N	10	10	8	6	1	0
SW-NE	**15**	7	**18**	12	**18**	**21**
W-E	10	**21**	6	**14**	10	11
NW-SE	3	1	3	4	2	3
Warm Front	2	0	1	2	1	1

Table 6.02: Frontal Boundaries and their orientation for each half-monthly time period.

Visual analysis of both upper-air and surface weather maps reveal a number of features about the synoptic mechanisms of nighttime precipitation in May, June and July. By far, in the upper-levels a single long wave pattern of a trough-ridge (T-R) or ridge-trough (RT) dominated the mode of circulation for many nighttime rainfall events, constituting nearly half (46%) of all circulation patterns. Yet, a number of other upper-level patterns are associated with nighttime precipitation, and it varies based on both the half-monthly and monthly analysis. In a similar manner, the surface boundaries needed for nighttime precipitation also vary depending on the time period examined. An interesting feature from the data analysis is in the increase in both ridge-trough (R-T) and SW-NE oriented cold fronts during early and mid-summer. In most circumstances, a ridge-trough pattern in place over the continental U.S generally meant northwesterly flow in the upper-levels in the central Great Plains. Since air mass movement and related cold frontal boundaries tend to follow the upper-level pattern, it makes sense for cold fronts to orient themselves in this manner during the summer months. Further examination of the synoptics of nighttime precipitation for the late spring-early summer precipitation maximum in the central Great Plains using composite analysis and mapping provide a visual understanding into the common modes of circulation.

6.2 Composite Analysis:

The synoptic patterns for 0z (1800LST) and 12z (0600LST) were analyzed for 40 events occurring during the first half of May. Analysis includes the 200 mb wind field as well as the 500 mb height patterns stretching from Canada down to southern Mexico. Of particular importance is the presence of any upper-level disturbances, the location of jet stream or wind maxima at 200 mb, as well as any other features of the flow pattern worth noting. Looking at the composite upper-level circulation for the first half of May, a clear trough-ridge pattern is in place for the United States. A trough axis is situated over Utah and Arizona, followed by a downstream ridge axis in the upper Midwest. A broad area of faster jet stream flows stretches from the eastern Pacific off of Baja, Mexico into the Great Plains and upper Midwest, with wind maxima located off the southern coast of Baja and over the Great Lakes region. Kansas lies downstream of the trough axis. In many cases, upper-level divergence occurs out ahead of the trough axis, leading to rising motions throughout the vertical and often times cloud development and precipitation. Few changes occur in the 12 hour period from 0z (1800LST) to 12z (0600LST); the only noticeable features being the slight eastward shift in the trough axis and a weakening of the winds at 200 mb. The eastward shift of the axis indicates a feature moving through the long wave pattern with perhaps a slight increase in its amplitude from 0z (1800LST) to 12z (0600LST). Indeed, out ahead of the trough axis a surface low pressure area resides in western Kansas, with the center of circulation situated over western Kansas and the panhandles of Texas and Oklahoma. Kansas also resides close in the area of a jet-streak (or a region where winds are stronger than other areas) at 200 mb. Generally, a jet streak has both an entrance region and exit region, with both ascending and descending motions due to air rising or sinking at the particular location. The entrance region of a jet streak is where winds are accelerating upstream of the jet streak axis, with upper-level divergence (spreading of the air) usually occurring to the right of the jet max (right entrance region). Upper-level divergence causes pressure or pressure heights to fall at the surface and lower levels. Similarly, the exit region also has a region of divergence, generally at its left flank downstream of the jet max. In areas of divergence, upward motions are promoted, increasing low-level flow and promoting instability and subsequently inclement weather (Figure 6.01) (Uccellini and Johnson, 1979).

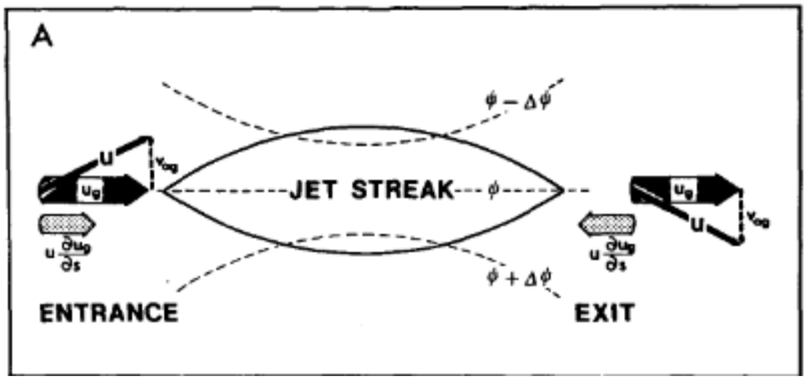
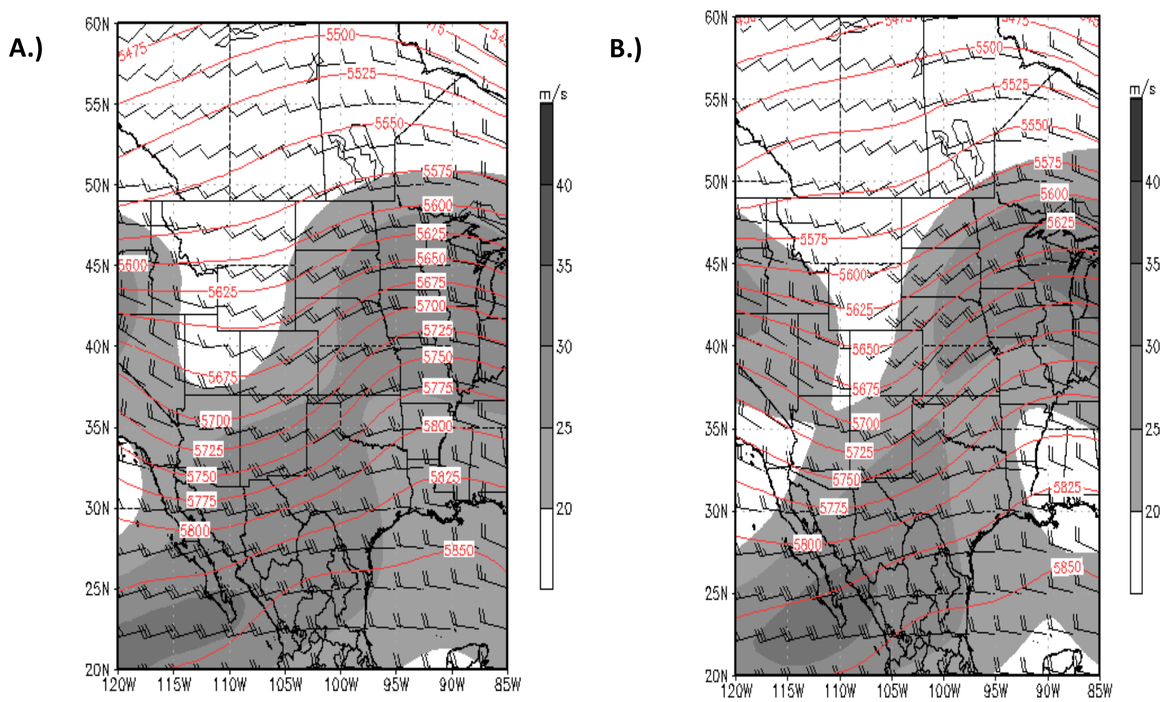


Figure 6.01: Idealized schematic of a jet streak. Right entrance region and left exit region represent areas of spreading air (divergence), left entrance and right exit are areas of air converging (convergence. Reproduced from Uccellini and Johnson (1979).

A clear surge of moisture influx from the gulf can be seen at both the surface and at 850 mb, though winds at the surface take a more cyclonic track compared to the wind flows at 850 mb. Given the general veering with height noticeable in the Chapter Five, southeasterly winds at the surface matches well with previous understandings of clockwise turning of the winds with height (Figure 6.02).



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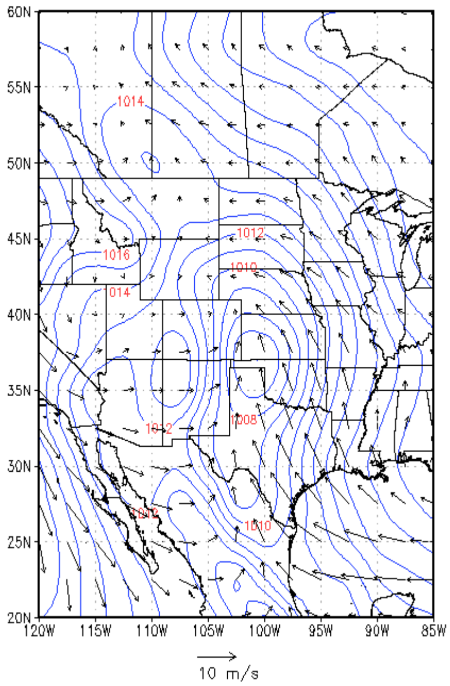

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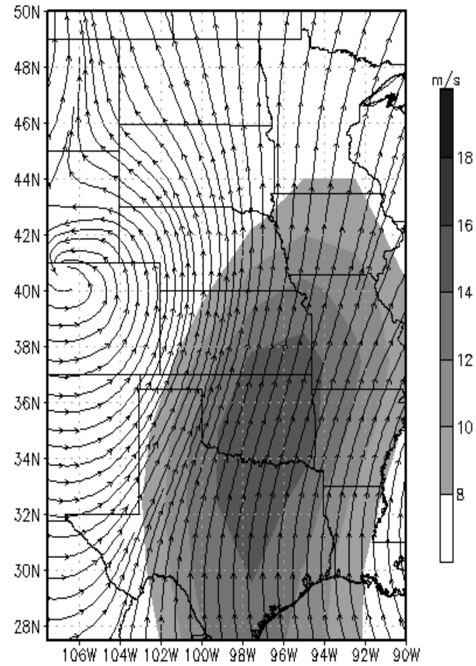
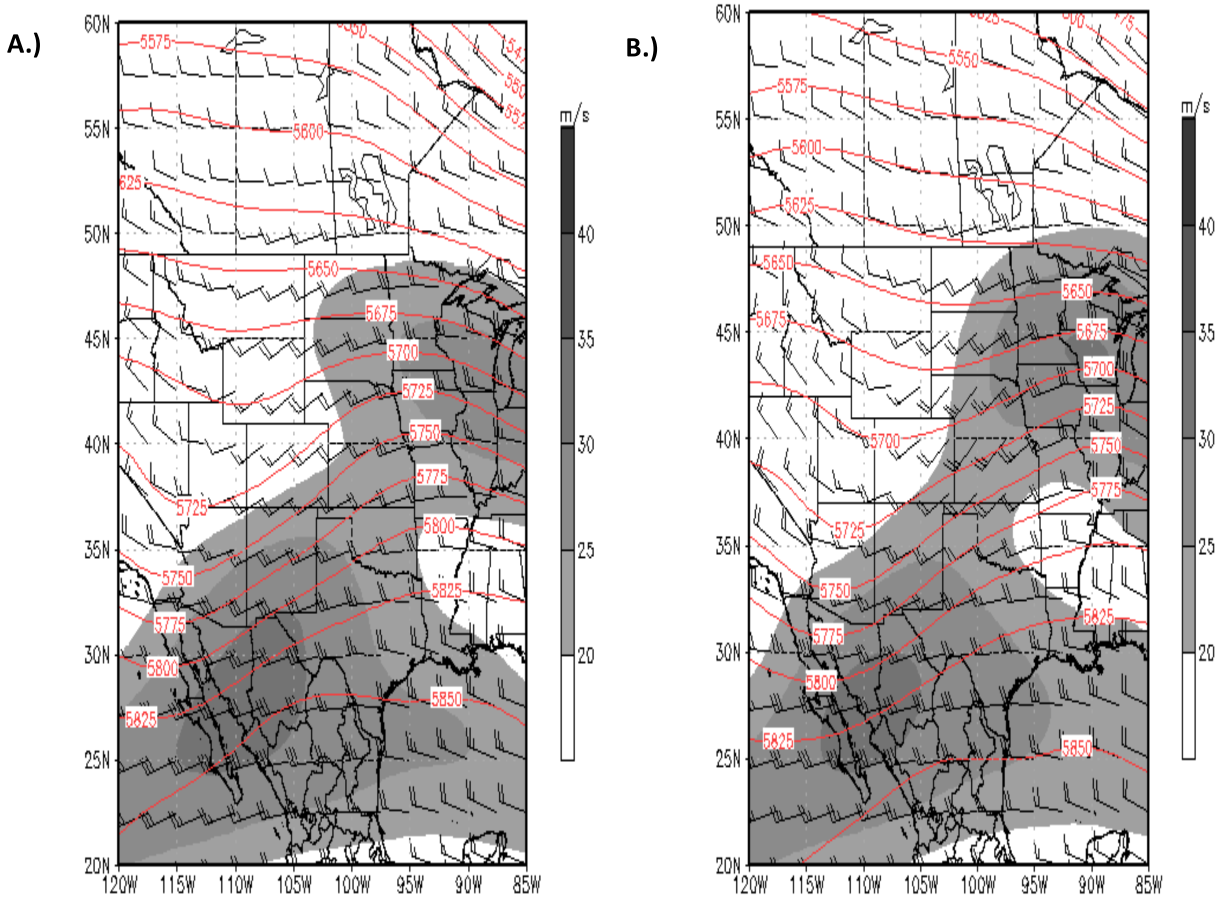


Figures 6.02: A.) 200 mb wind field 500 mb geopotential height for 0z (1800LST) and B.) 12z (0600LST) for 40 May 1-15 events. The shaded regions represent winds speeds >25 m/s at 200 mb. C.) Surface sea level pressure (blue) and wind arrows averaged for 0z (1800LST) to 18z (1200LST). D.) Displays low-level wind characteristics for 850 mb at 6z (0000LST).

The upper-level flow pattern for events in the latter half of May exhibit features which differ from the first half of the month. For instance, the 200 mb wind speeds weaken and contract, though the location of the jet axes remain in the same locations. The 500 mb pattern still displays a trough-ridge pattern, but the pattern shifts westward. The trough axis now resides near Nevada and Arizona, with the ridge axis located over the eastern Great Plains. Height contours also display a northerly shift, indicating an increase in temperatures over the study region. The 12-hour period also sees an increase in amplitude of the trough axis and an indication of the disturbance moving eastward. Surface weather patterns for nocturnal events in the latter half of May differ very little from the first half of the month. Gulf moisture continues to flow into the central Great Plains, along with west/southwesterly flow originating from the western U.S. The mixing of the air masses in mid-late May is cited as one of the reasons for the peak in supercellular thunderstorm formation in northeastern Kansas (Thompson and Edwards, 2000). Wind flows at 850 mb display a decrease in speed and coverage from the first portion of the month. Therefore, the characteristics of the low-level winds change not only on a monthly basis, but a half-monthly basis as well. Interestingly, surface-level convergence can be seen to the northwest of the low pressure region at 850 mb. With the general eastward movement of storms in the nighttime hours, surface-based convergence may form in areas of eastern Colorado and western Kansas and Nebraska in the afternoon hours and then track eastward into northern Kansas during the nighttime hours (Figure 6.03).



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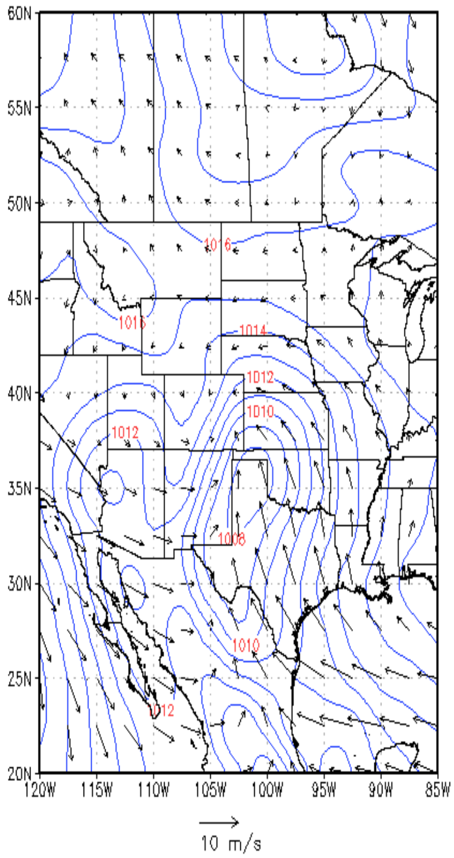

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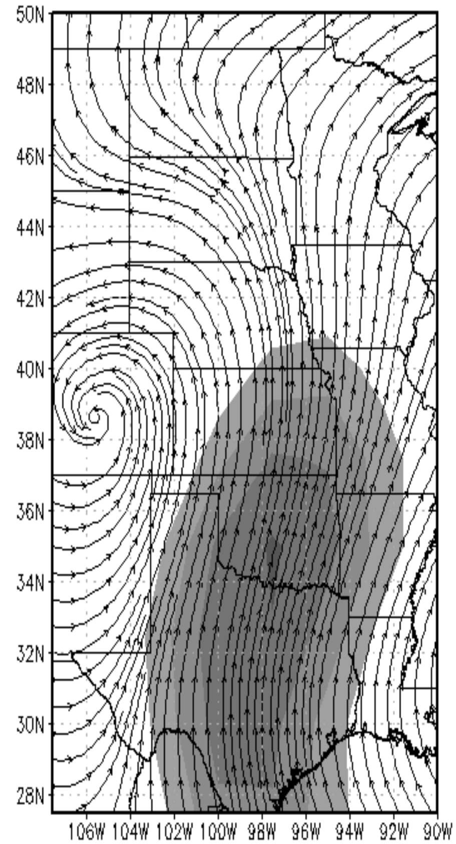
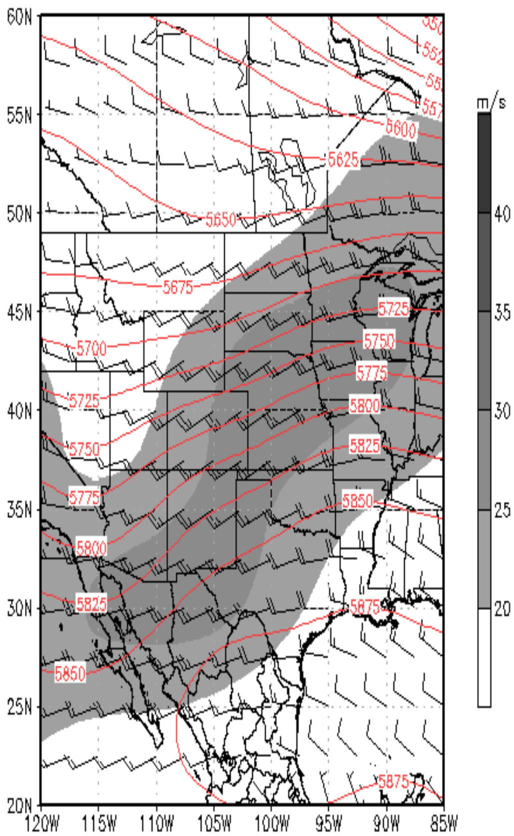


Figure 6.03: Same as Figure 6.02 for May

The trend of decreasing upper-level westerly winds at 200 mb continues into the first half of June, with now only a broad region of winds of ~25 m/s (55 mph) extending from the Gulf of California up to Wisconsin at 0z (1800LST). A slight increase in the wind field occurs at 12z (0600LST). A similar westward shift in the 500 mb height pattern is present in early June as well, with perhaps a slight glimpse into the patterns more commonly seen during the summer in the United States. Regardless, a trough axis continues to reside in the western United States, now shifted to southern California, with Kansas under the influence of southwesterly flow extending out from the trough axis. More than likely the longer wave trough axis in the western U.S. is not the driver of local processes that produce precipitation in northeast Kansas. Smaller upper-level short waves moving through the longer wave train are likely to provide the necessary localized ingredients for uplift and nighttime rainfall. The surface features resemble those of previous analysis times, though the surface low in the panhandles of Texas and Oklahoma weakens slightly in terms of both central pressure and spatial coverage. Interestingly, the 850 mb winds actually display an increase in speed and coverage from the latter half of May into early June. This may be the result of an anomaly in the datasets for either time period, but nevertheless a distinct difference in the wind flow in the lower levels of the atmosphere exists in the transition from late spring into early summer (Figure 6.04).

A.)



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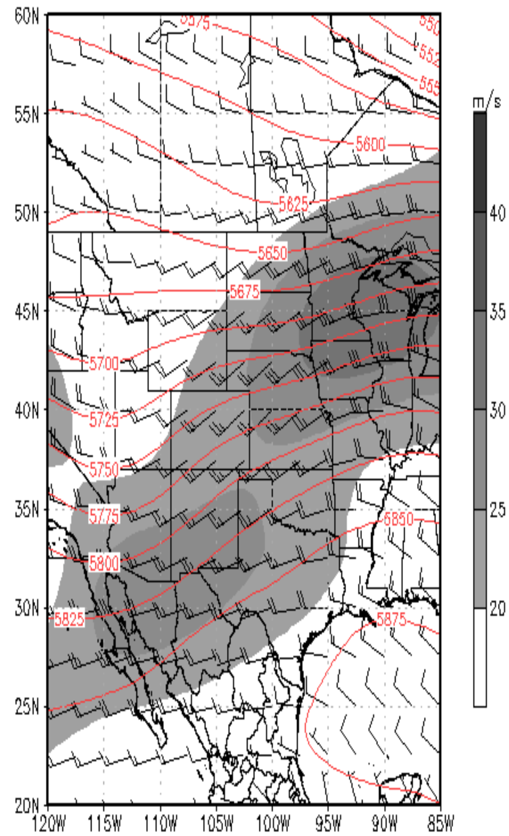
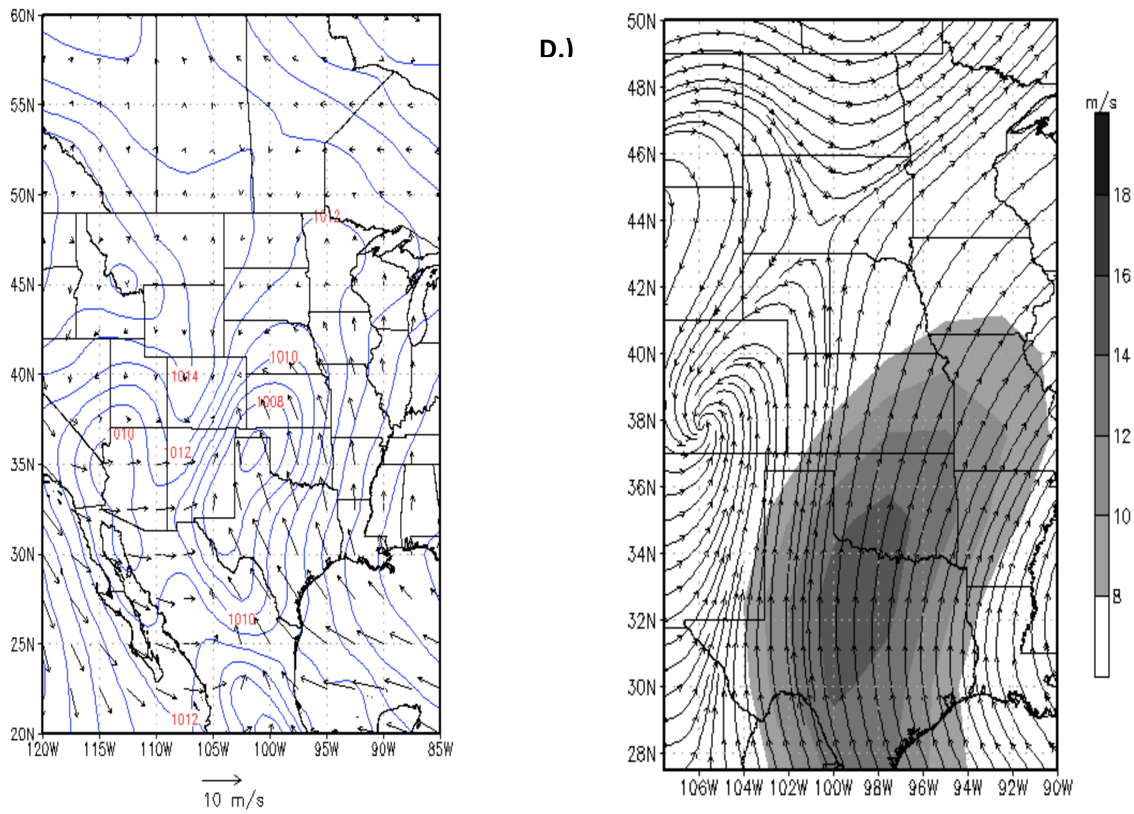


C.) D.)



Figures 6.04: Same as Figure 6.02 for June 1-15.

The pattern in late June most resembles west-east zonal flow across much of the United States, with a slight indication that the upper-level pattern is beginning to shift into the more typical summertime flow where a western Great Plains ridge controls much of the central United States. Similar to early June, coupled with a relatively strong anticyclonically oriented jet at 850 mb and a surface low in Kansas, the main culprit behind nighttime precipitation likely results from smaller upper-level waves moving through the zonal long-wave pattern as opposed to the larger synoptic-scale trough-ridge pattern in the early time periods. However, given the resolution of the data and the reanalysis modeling, small short waves do not show up in the dataset. The 200 mb jet continues its northward track as spring turns into summer in the United States, reaching its peak northerly latitude over the northern Great Plains and upper Midwest. Kansas now resides to the south of the upper-level jet stream. Though weaker, a continued influx of Gulf moisture at the surface steered cyclonically converges in western Kansas and Oklahoma, just south of the 200 mb wind maxima. Surface level-convergence can be seen along the Kansas-Nebraska border, as well as long the Kansas-Colorado border (Figure 6.05).

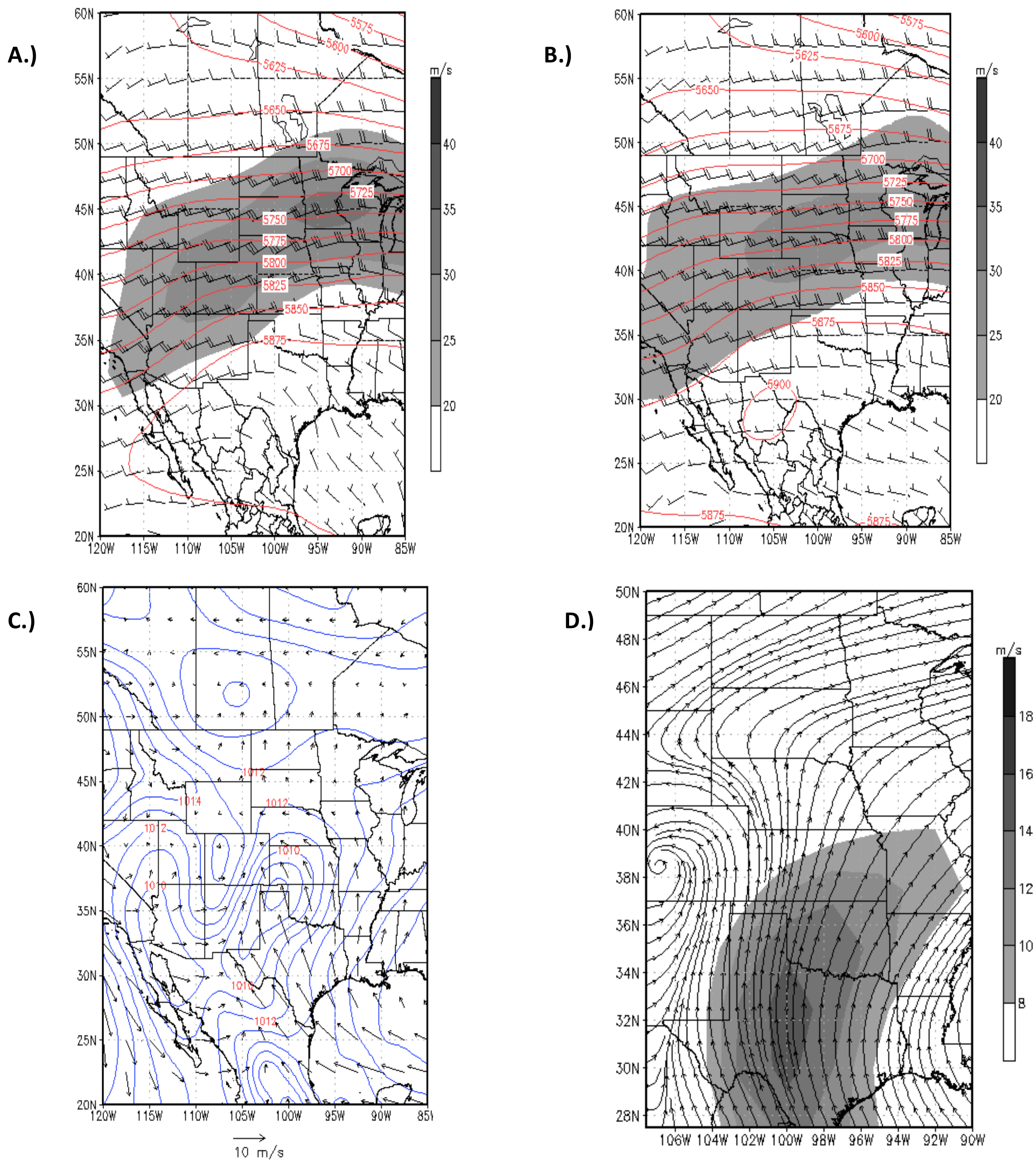
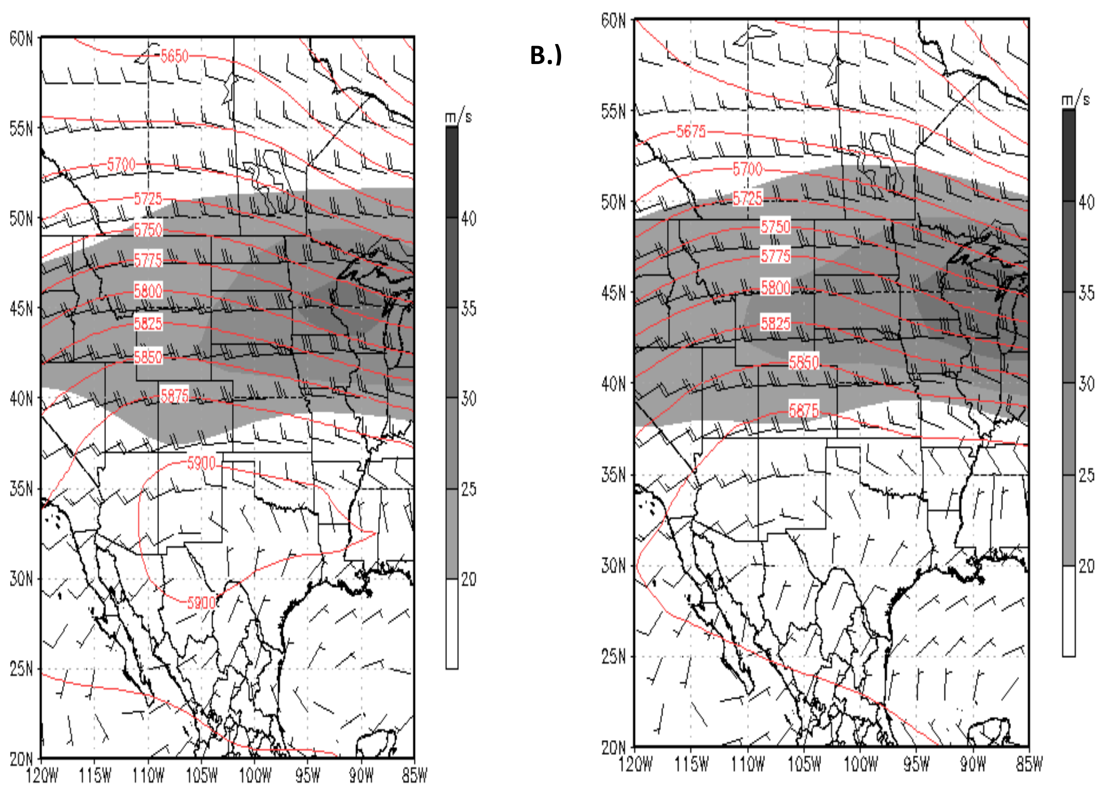


Figure 6.05: Same as Figure 6.02 for June 16-31.

By far, the greatest changes to the upper-level circulation occur from late June into early July. No longer does the more frequent spring season trough-ridge longwave pattern dominate the United States, rather a the shorter wavelength pattern with the ridge over the eastern Rockies or High Plains sets up. In the case of periods of nighttime rainfall, the central North American ridge shifts westward. Kansas now resides to the east of the ridge axis, with predominant northwesterly flow. The ridge axis, however, shifts eastward from 0z (1800LST) to the next morning at 12z (0600LST). Although previous 200 mb wind vectors decrease during the 12-hour period, in the case of early July events winds speed increase. Surprisingly, whereas the upper-levels share little to no similarities to the previous analysis times, at the surface the same patterns continue to persist. Though weaker, the surface low-pressure area and surge of moisture from the Gulf of Mexico continue in the same areas in early July, as does the wind pattern at 850 mb. With the changes in the upper troposphere and persistent patterns in the lower levels, more than likely the mechanisms which promote nighttime thunderstorm activity changes from late spring into early summer. No longer are upper-tropospheric wave features such a trough west of Kansas the more important drivers of nighttime convection. Rather, low-level based convergence in the form of low-level wind maxima now dominates the mode of circulation needed for nighttime rainfall (Figure 6.06).

A.)

B.)



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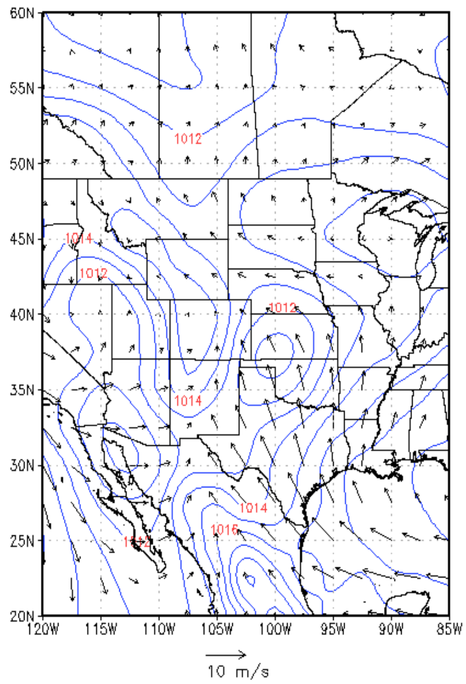

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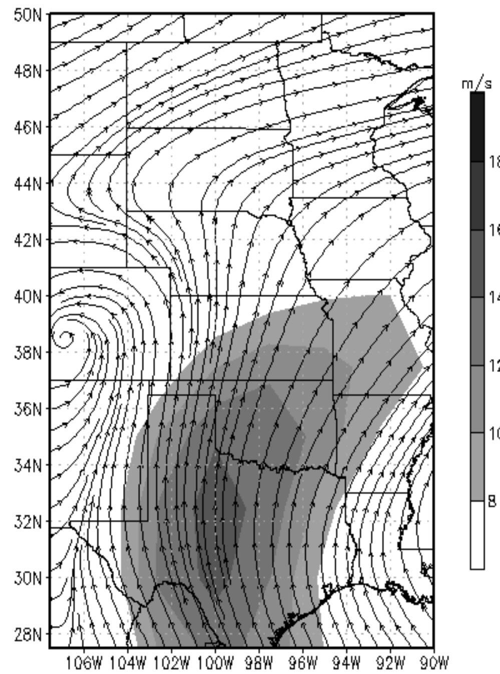


Figures 6.06: Same as Figure 6.02 for July 1-15

The circulation for final two weeks of July closely resembles that of early July. A pronounced ridge persists in western United States, with an axis over Utah, Idaho, Montana and up into Canada. Compared to early July, the ridge shifts slightly westward. Once again, northwesterly flow aloft prevails over a Kansas. The jet stream continues its northward track into the northern United States and into Canada, with the strongest winds continuing to be centered on the Great Lakes region. This again keeps Kansas in the right entrant area of the wind speed maximum and this upper-level dynamic might influence the ability of the atmosphere in northern Kansas to sustain strong vertical motions during the overnight hours. The ridge axis located along the U.S Canadian border shifts from 0z (1800LST) to 12z (0600LST), an indication of some type of upper-level disturbance causing an eastward shift in the pattern (Figure 6.07).

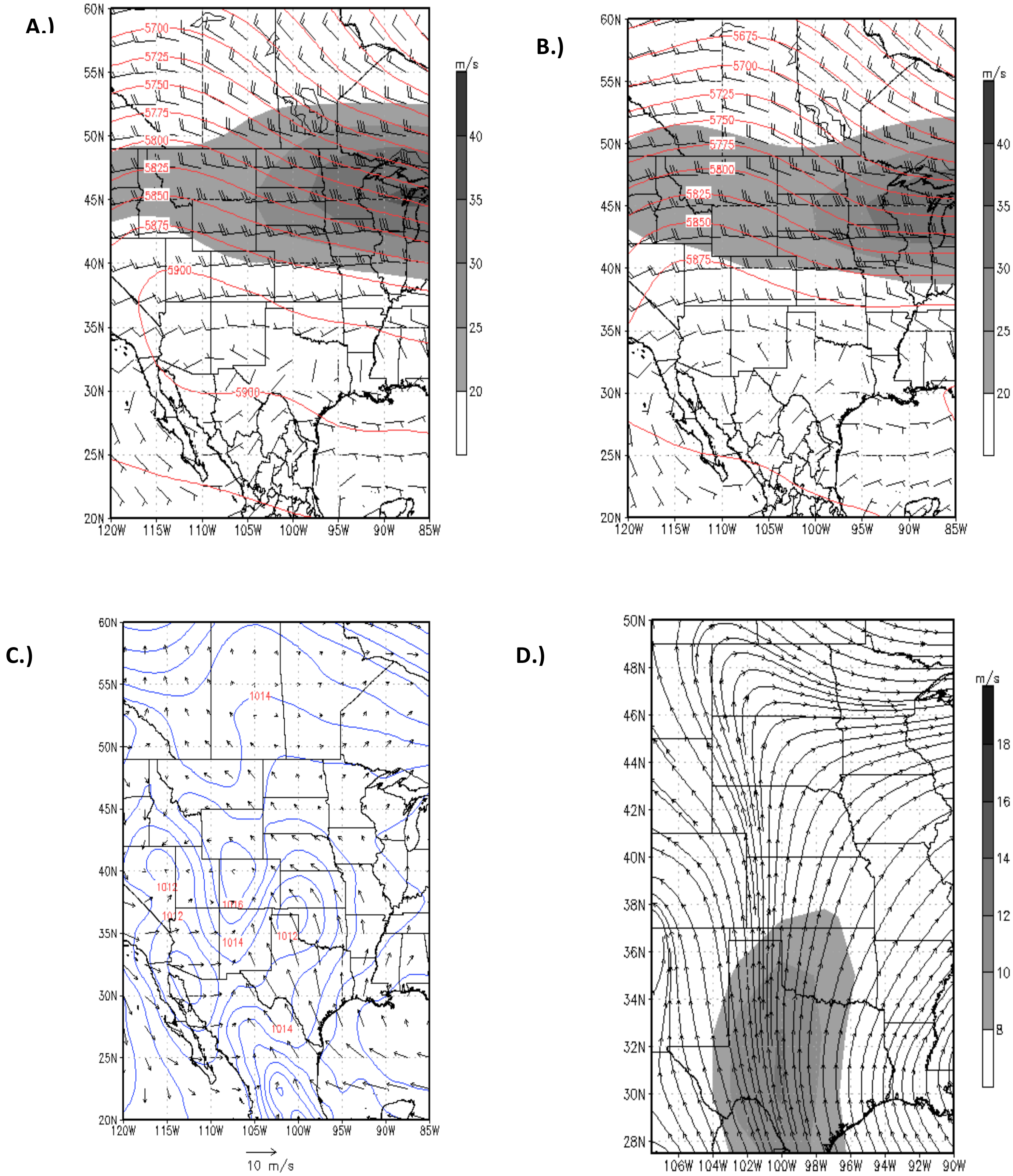


Figure 6.07: Same as Figure 6.02 for July 16-31.

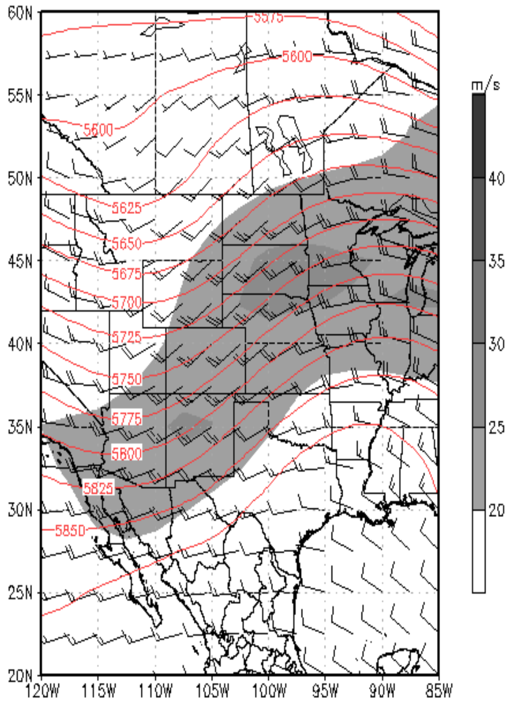
Interestingly, though major changes occur in the higher levels of the troposphere, the same surface features show up in every analysis period and weaken only slightly from the beginning of May to early July. Similar to early July, with the dominant ridge pattern in place and northwesterly flow aloft, more than likely the energy fueling nighttime thunderstorms comes from surface-based or low-level convergence as a result of the northward movement of moisture transported by the southerly low-level winds.

Similar to the statistical analysis, the visual composites of the atmosphere assist in understanding the changes in circulation from month-to-month. The mechanisms of nighttime precipitation vary greatly in the upper-levels from May-July. In May, the large-scale trough in place over the western U.S coupled with a relatively strong upper-level jet helps in creating a favorable environment for nighttime precipitation. However, come summer, the long-wave pattern switches to more zonal and eventually a ridge-like pattern in the central United States. More than likely, large-scale upper-level disturbances in late-spring give way to smaller disturbances moving through the long wave pattern during the summer. Interestingly, however; the features at surface remain relatively the same from May through July. Gulf inflow of moist tropical air can be seen at all time periods, converging on a surface low-pressure feature in the western Great Plains.

From the composite analysis, it is also noticeable how changes in the upper-level patterns can be linked with changes to the structure and strength of the low-level winds. Similar to the findings of the previous chapter, the low-level winds weaken in both strength and horizontal coverage from May-July, largely a result of the changing dynamic forcing related to shifts in the patterns in the upper atmosphere. In May, common modes of circulation include the movement of strong upper-level disturbances and in some cases closed-low pressure systems/mid-latitude cyclones that generate a southerly wind component in the warm sector of the cyclone. The synoptic-induced southerly wind or LLJ is a stronger, more robust feature compared to its summer thermal-induced counterpart. Indeed, the idea of a stronger, synoptically-related southerly wind matches well with previous findings. Chen and Kpaeyeh (1992) found in their examination of 64 cases studies that a developing upper-level baroclinic wave with an upper-level jet stream frequently forms east of the Rocky Mountains in the spring months. The upper-level divergence (lateral spreading of the air) enables vertical motions that lead to an increase in wind speeds at the lower levels ahead of the baroclinic wave. In other words, the large temperature gradient separating the air ahead of the approaching wave and the air behind the wave causes a robust south-to-north wind stream to form out ahead of the wave. The large temperature gradient results from the pressure gradient force (PGF) in the upper-atmosphere caused by a strengthening upper-level disturbance, thus creating a stronger and more robust southerly wind component or LLJ. The stronger of the upper-tropospheric wind field and divergent component, the stronger the vertical motions and resultant low-level winds (Ucellini and Johnson, 1979 Maddox and Doswell, 1982; Achtor and Horn 1986). However, come summer when the stronger upper-level short waves are quite rare in the region, the thermal-induced wind maxima or LLJ is the more common low-level feature. As seen in both this chapter and Chapter Five, the weakest low-level winds occur later in the warm season, since their formation largely depends on the temperature gradient between the western and eastern Great Plains (Bonner, 1968). Using the classes analyzed in the Chapter Five, further composite analysis will reveal the synoptic patterns in place for the varying types of low-level winds in the subsequent section.

6.3 Composite Analyses of Circulation Patterns Related to Low-Level Wind Maxima:

In the previous chapter, five types of wind maxima were identified using data from the Wichita grid cell for events greater than 1.5in (38.1mm). In this section, the circulation patterns most common to five classes of LLJs are analyzed. Similar to the previous section, composite analysis of both the 500 mb height patterns, along with winds at 200 mb are included. Beginning with the "Robust" class of winds, the upper-level patterns most closely resemble the pattern seen in the previous section for May events. A trough digs in the western United States with an axis over stretching from Montana to Arizona, followed by an upstream ridge with an axis over the upper-Midwest into Canada. An expansive jet stream resides at 200 mb, stretching from the desert southwest of the United States up to the Great Lakes region, with a jet max in both New Mexico and the northern Great Plains into Minnesota (Figure 6.08). The coinciding composite of Robust winds shows a strong and extensive region of low-level winds just east of the jet axis, centered over central Oklahoma and southern Kansas (Figure 6.09).



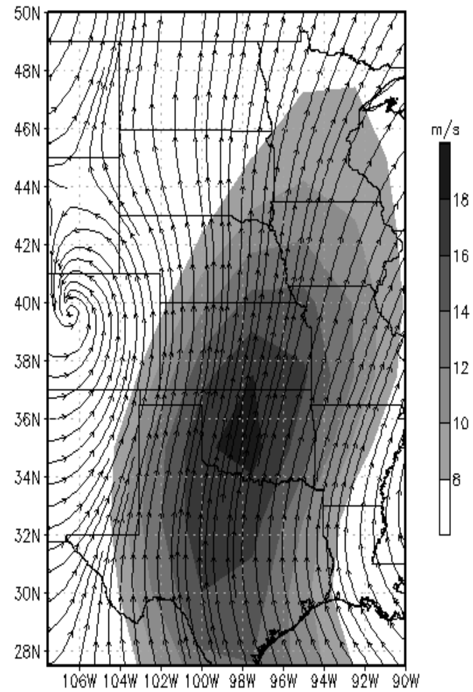
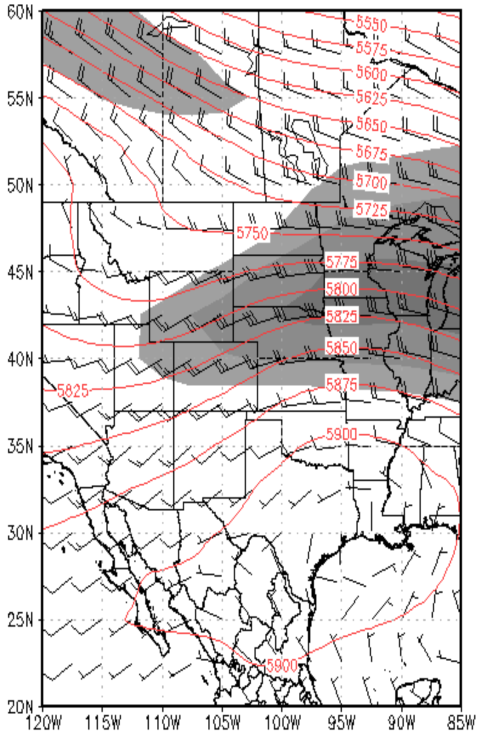


Figure 6.08: Composite analysis of upper-level patterns for Robust Class averaged for 0z (1800LST)-1800z (1200LST). Figure 6.09: 850 mb composite for Robust class of winds at 6z (0000LST).

For the Semi-Robust class of low-level winds, the upper-level patterns exhibit a few changes compared to the previous class. Though a western trough and downstream ridge pattern is still in place, both the trough axis and ridge axis are not as well defined. Moreover, both waves weaken slightly in terms of their amplitude. Nevertheless, upper-level flow in Kansas continues to be from the southwest. The fastest winds at 200 mb are now confined to northern Great Plains and upper-Midwest. The resulting low-level wind maxima subsequently weakens in strength and geographic coverage. Moreover, the 850 mb axis shifts southwestward to central Texas and Oklahoma (Figure 6.10 and 6.11).



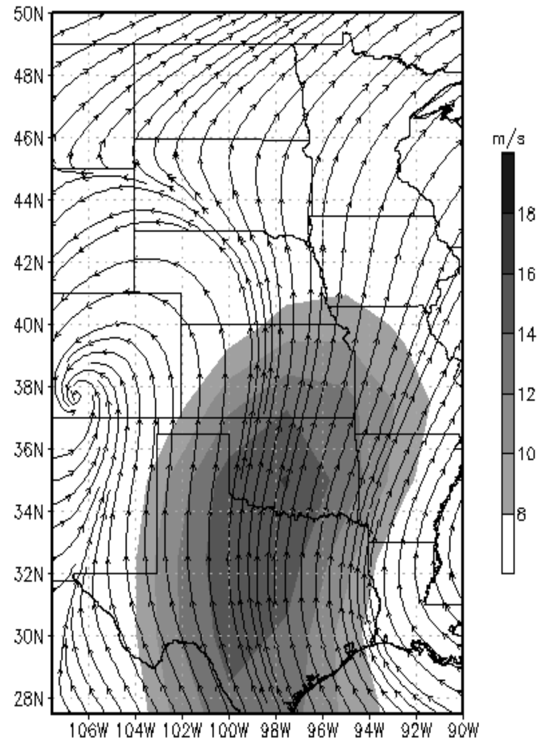
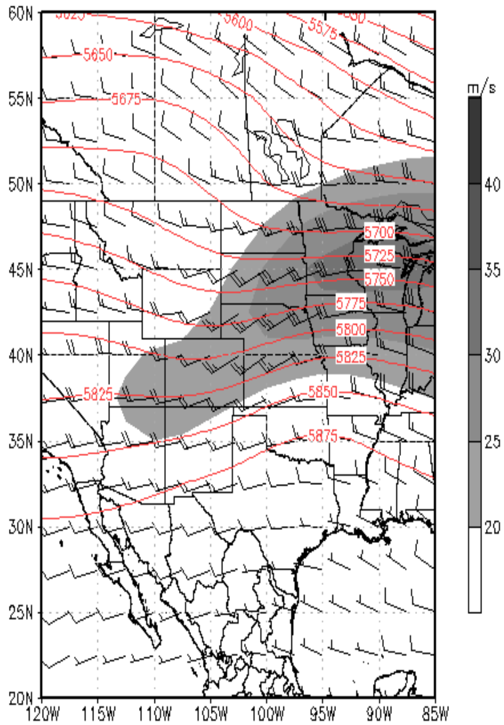


Figure 6.10 and 6.11: Same as Figures 6.08 and 6.09 for the Semi-Robust class.

For the third class (6-8), the upper-level composite shares little to no similarities with the other four patterns associated with each class of winds. Slight troughing is present in the western United States and northern Great Plains, with a ridge axis along the eastern Great Plains near Arkansas, Missouri and Iowa. Though the 200 mb jet weakens from the previous class, it nevertheless shifts slightly southward and westward. Similarly, the wind maxima at 850 mb weakens and shifts westward, with a clearer anticyclonically turning of the winds coming from the western Gulf of Mexico with an increase in latitude (Figures 6.12 and 6.13).



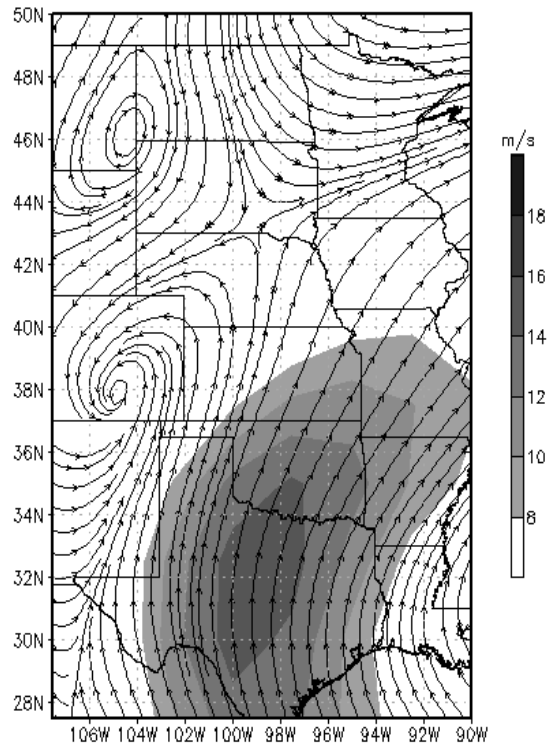
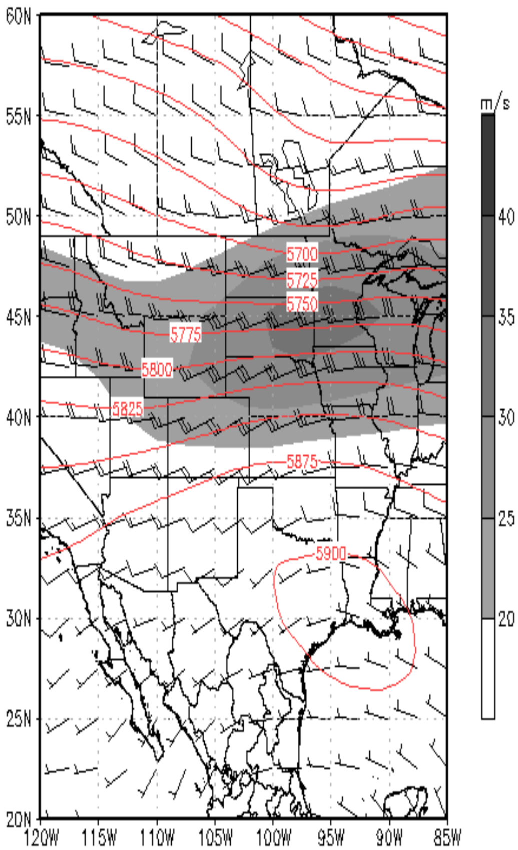


Figure 6.12 and 6.13: Same as Figures 6.08 and 6.09 for Class 3 winds (6-8).

Though not very well defined, a broad ridge of higher heights dominates much of the study area for the fourth class (3-5). A trough is evident across Manitoba. Compared to the previous class, the 200 mb jet shifts to the west, located just north of the ridge axis and just south of the trough axis along the United States/Canadian border. Kansas continues to be of west/southwesterly flow just south of the 200 mb jet. The coinciding 850 mb wind maxima continues to be located south of the jet stream at 200 mb. Additionally, with the less amplified flow present in the upper-levels, the winds weakens and curves to the east with increasing latitude compared to previous classes (Figures 6.14 and 6.15).



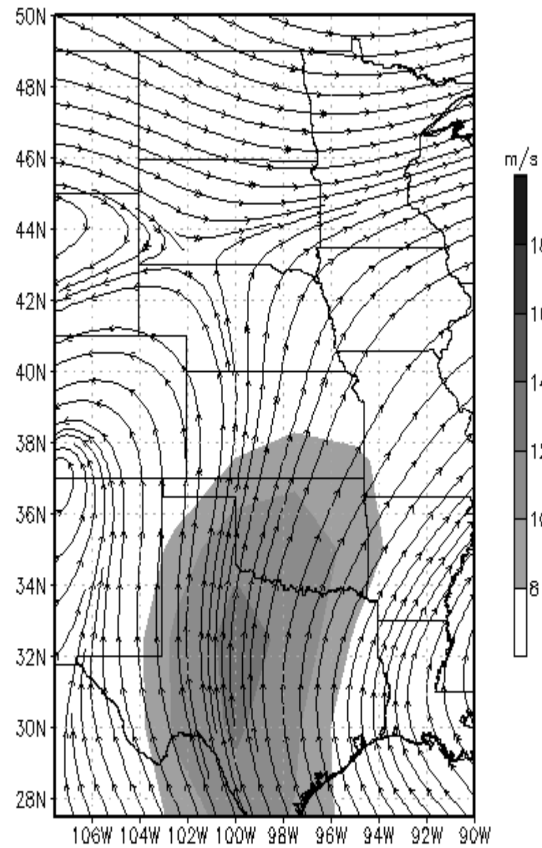


Figure 6.14 and 6.15. Same as Figures 6.08 and 6.09 for Class 4 winds(3-5).

The final class of low-level winds, those exhibiting little to no wind maxima-like qualities, features similar patterns of the previous class in the upper-levels. The flow pattern throughout the United States is much less amplified than the previous classes, with a slight trough-pattern in the western High Plains followed by a downstream ridge in the Midwest. It also appears Kansas resides south of the right-entrance area of the 200 mb jet streak. The zonal-like flow pattern set up across the United States results in a weak low-level wind field in the southern Great Plains. Though, at 850 mb an area of low-level of convergence in northern Kansa can be seen associated with a frontal boundary located in Nebraska. With no amplified trough and little in the way of moisture influx from the Gulf, nighttime precipitation must result from smaller-scale disturbances in the upper-level that can generate uplift through the atmosphere. Keep in mind, these wind classes are based on rainfall events greater than 1.5 inches (38.1mm). Therefore, substantial moisture must be present in the atmosphere. Though no jet-stream like area is present in the lower-levels, moisture nevertheless continues to flow from the Gulf of Mexico up to the central Great Plains, indicating that a fast wind maxima is not the only type of low-level wind that can produce heavy nighttime rainfall events. (Figure 6.16 and 6.17).

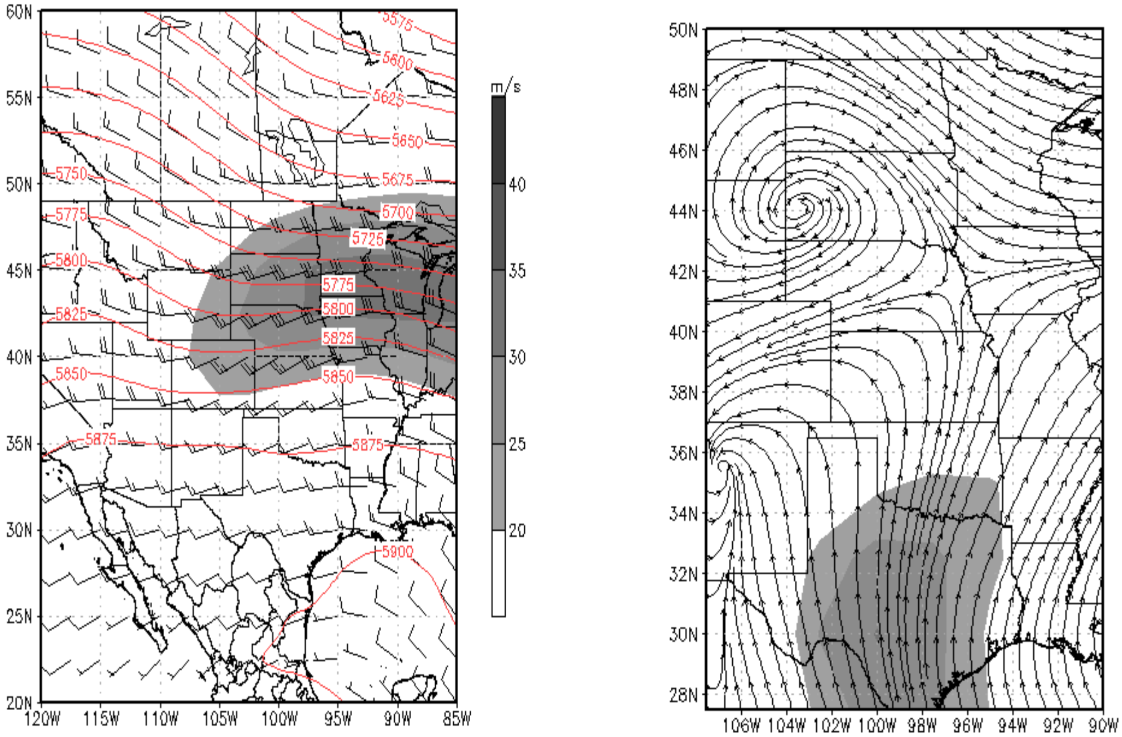


Figure 6.16 and 6.17: Same as Figures 6.08 and 6.09 but for Class 5 LLJs (0-2).

6.4: Discussion:

From the composite analyses, it is apparent that the structure southerly low-level winds are linked with the upper-level patterns. More amplified flow resulted in a stronger low-level winds persisting for much of the nighttime and early daytime hours at lower levels of the atmosphere. However, with less amplified flow and weaker winds at the top of the troposphere, the mechanisms needed to generate a spatially and vertically robust winds diminish in intensity. The resulting low-level winds are weaker and cover less area geographically. Despite this, heavy nighttime rainfall events can occur in absence of both an amplified upper-level pattern and a strong wind maxima in the southern and central Great Plains. Whether analyzing circulation patterns on a bi-weekly basis from May to June, or basing the analysis off of low-level wind characteristics, a number of features at varying levels of the atmosphere contribute to nighttime rainfall in the central Great Plains.

CHAPTER SEVEN:
CONCLUSION AND IMPLICATIONS

"And that's all I have to say about that."

-Forrest Gump

7.1 Summary:

Presented in this research are important aspects of the nighttime rainfall characteristics for Topeka and other stations across Kansas for the late spring- early summer precipitation maximum for a 63 year period from 1950-2012. Supplementing the analysis of nighttime rainfall data in Kansas includes a study of both the sub-synoptic and synoptic scale mechanisms that generate nocturnal convection in the central United States, most notably the Great Plains Low-level Jet. A statistical climatological as well as a synoptic climatological approach was used to generate a more complete understanding of both the long-term patterns of nighttime rainfall and the factors which contribute to the diurnal nature of rainfall in the central Great Plains.

Hopefully, the results and discussion presented in this research strengthens the ideas that a.) nighttime rainfall is extremely important to the Great Plains hydroclimate, and b.) different mechanisms can generate nighttime convection. A total of 463 nocturnal events in May-July were recorded during the 63 year period, equating to around seven nocturnal events per season. Nocturnal rainfall contributes even more to the moisture budget later in the warm season, as evident from the increase in the percentage of nighttime rainfall from May to July. However, the number of nocturnal events did not increase significantly from May to July, as each month recorded a similar number of nocturnal events (154 for May, 156 for June, and 153 for July). Additionally, as evident from the analysis of other stations, the stations furthest east experienced more nighttime rainfall than those westward. No statistically significant long-term trends were evident in the data.

The main objective of the research; however, was to develop a synoptic climatology which delved into the understanding of the atmospheric mechanisms which contribute to nighttime rainfall during the late spring-early summer precipitation maximum with an emphasis on the evolution and structure of the southerly low-level winds in the Great Plains. Indeed, a wind maxima existed in some form or another for almost all nocturnal rainfall events of 1.5 inches (38.1mm) or greater. Whether by analyzing the low-level jet on a monthly basis or based on the wind characteristics within the study area, the low-level winds exhibited considerable variation in terms of its strength, coverage and duration. Though, on average the southerly winds reached their peak at 6z (0000LST) near central Oklahoma at 850 mb. By far, those wind maxima associated with nighttime rainfall events in May were the strongest vertically, geographically, and temporally. As the warm season progressed, the wind maxima became more contracted to the lower levels of the atmosphere and tended to be more a nighttime phenomenon, only developing between 6z (0000LST) and 12z (0600LST). Although southerly low-level winds vary incredibly during nighttime rainfall events, they could nevertheless be categorized based on similar characteristics, resulting in five classes displaying the low-level wind types. Interesting to note, however; is that whereas winds at 700 mb displayed qualities which fit the criteria of a low-level wind maxima in this research, visual analysis indicated that winds at 700 mb should not be considered part of the southerly wind maxima, as the winds rarely originated from the Gulf of Mexico.

Though it has been long established that the upper-level circulation patterns that generate convection change throughout the year, it nevertheless became apparent that with changing upper-level patterns from May-July also came the changing nature of both nighttime precipitation and the configuration of southerly winds. As noted in Chapter Four, from May to July the spatial coverage of storms greatly decreases, as the result of a change from large-scale upper-level disturbances such as migratory mid-tropospheric troughs and closed to lows, to smaller sub-synoptic waves moving through the long wave pattern. Moreover, as the influence of larger disturbances waned from late spring into the summer, so too did the vertical and geographic strength of the low-level winds. As a result, the mechanisms behind the southerly wind formation change in a matter of weeks, typically switching from synoptic mechanisms in late May and early June to the thermal, nocturnally-induced wind maxima seen most commonly seen in the summer.

7.2 Limitations:

Throughout the research, various issues and limitations arose which need to be addressed for the benefit of both this research and future research. Regarding rainfall, the lack of stations with long-term hourly precipitation data hindered the ability to understand the spatial complexities of nighttime rainfall. Moreover, the primary use of data from one station raises questions about its ability to sufficiently represent the rainfall for an area. A 1.5 inch (38.1mm) rainfall in Topeka may not equate to the same amount of rain just 10 miles away. Regardless, some type of mechanism created a heavy rainfall event at that location, and given the main objective of developing a synoptic climatology of nocturnal rainfall, one station provided more than enough data to do so.

The limitations of analyzing low-level winds and their relation to nighttime convection need to be addressed for both this research and future research. As previously mentioned, in this research the temporal resolution between nighttime rainfall and wind observations vary. Though the NCDC provided hourly surface weather data, estimates of the character of the upper levels of the atmosphere derived from the reanalysis datasets are only available for 4 time slots. The 6-hour time gap between observations, particularly during the nighttime hours from 6z (0000LST) and 12z (0600LST), may miss key periods where wind observations would be pertinent towards an improved understanding of the coupling of the low-level jet and nighttime rainfall. Limitations also exist when using reanalysis datasets. Whereas the error associated with low-level jets is relatively small (less than 5%) (Wang and Chen, 2009) , it is important to remember, however; that reanalysis datasets may not adequately represent the atmospheric state for certain regions and time periods and have been known to have significant biases for certain variables in different areas and time periods. Even today, the atmosphere is not fully observed. Certainly, the availability of data in the United States has increased tenfold over the past few decades, thereby strengthening the confidence in using a reanalysis dataset to research subsynoptic to synoptic scale phenomenon such as the low-level jet and other upper-atmospheric features. Nevertheless, reanalysis datasets should not be considered observations of reality.

7.3 Future Research:

With any research activity come new ideas about ways to improve our understanding of the phenomenon examined. This thesis being no exception, a number of new directions could be applicable based on the findings presented here. Nighttime rainfall, for example, needs much further examination. Though studies have examined the diurnal rainfall variability in the Great Plains, few if any have taken station data for the entire region to analyze its contribution to the overall rainfall total and if that has changed over the years. With the importance of precipitation to the region's agriculture and the growing concern of precipitation variability resulting from a changing climate, understanding the diurnal aspects of nighttime precipitation over a multi-decadal period could perhaps pinpoint if and where any changes to the nature of nighttime rainfall have occurred.

In regards to LLJ and southerly wind research, much has been done from both an applied and theoretical sense. Nevertheless, with its importance to precipitation, trying to pinpoint the location, strength, and timing of the low-level jet could greatly aid operational forecasting. Forecast modeling continues to struggle with nighttime rainfall in the Great Plains. From a climatological sense, higher resolution data could further delve into the vertical and geographical aspects of the LLJ. Datasets such as the North American Regional Reanalysis (NARR), a much higher resolution dataset both vertically and geographically, should be utilized for LLJ research. This research chose Reanalysis V1 due to its temporal length as it goes back to 1950. NARR, on the other hand, only goes back to 1980. Regardless, more temporal and spatially resolute climatologies of nocturnal rainfall and the LLJ can be developed using higher resolution datasets. Further analysis of the LLJ should also take into account larger-scale processes beyond the upper-level patterns discussed here. Though research has analyzed the relation to the Bermuda High and the influx of moisture, future research could take into account how both nocturnal precipitation and the LLJ relates to the shifts in the surface high pressure cell.

Simply, nighttime rainfall in the central Great Plains contributes greatly to the region's hydroclimate during the warm season. How and where that rainfall forms, however; varies greatly through the workings of multiple mechanisms throughout the atmosphere.

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Appendix A: Topeka Monthly Rainfall Total Climatology (1950-2012)

Year	May	June	July	Total
1950	6.12	5.81	12.02	23.95
1951	4.12	10.59	11.57	26.28
1952	3.78	0.92	0.85	5.55
1953	4.53	1.07	4.17	9.77
1954	4.38	3.78	5.21	13.37
1955	4.62	5.06	2.21	11.89
1956	2.36	5.32	4.44	12.12
1957	4.18	5.1	5.21	14.49
1958	2.29	7.52	8.58	18.39
1959	4.7	2.87	2.92	10.49
1960	1.7	3.43	3.16	8.29
1961	5.29	2.43	7.25	14.97
1962	4.23	4.46	3.68	12.37
1963	4.91	2.71	3.13	10.75
1964	2.09	8.1	1.71	11.9
1965	3.41	10.14	3.33	16.88
1966	0.41	8.83	0.75	9.99
1967	5.07	15.2	3.06	23.33
1968	3.37	3.18	10.17	16.72
1969	3.77	3.46	3.26	10.49
1970	5.46	5.82	1.39	12.67
1971	4.83	3.1	4.07	12
1972	2.9	1.14	4.81	8.85
1973	4.37	2.96	10.16	17.49
1974	3.69	3.72	2.9	10.31

1975	3.88	4.85	0.68	9.41
1976	4.63	1.69	2.04	8.36
1977	7.83	10.91	1.37	20.11
1978	5.75	4.57	2.26	12.58
1979	2.25	5.63	5.84	13.72
1980	4.85	0.56	0.87	6.28
1981	5.93	9.4	7.63	22.96
1982	9.39	5.99	5.08	20.46
1983	4.93	6.08	0.59	11.6
1984	3.45	10.17	1.66	15.28
1985	3.79	5.15	2.9	11.84
1986	7.53	2.51	4.21	14.25
1987	3.89	4.86	2.78	11.53
1988	3.08	3.13	1.74	7.95
1989	4.05	4.76	5.21	14.02
1990	4.45	5.57	3.01	13.03
1991	7.09	1.49	1.47	10.05
1992	1.75	3.35	6.37	11.47
1993	6.95	2.18	10.98	20.11
1994	0.95	4.63	3.16	8.74
1995	11.81	3.44	5.12	20.37
1996	7.72	7.97	2.65	18.34
1997	3.54	1.36	2.59	7.49
1998	2.08	7.22	9.32	18.62
1999	6.38	6.2	0.59	13.17
2000	2.08	7.25	2.77	12.1
2001	3.85	6.49	2.31	12.65

2002	4.87	4.12	0.81	9.8
2003	3.7	3.7	0.7	8.1
2004	4.46	6.39	7.27	18.12
2005	4.58	9.59	2.08	16.25
2006	3.2	1.18	3.41	7.79
2007	10.25	4.39	1.99	16.63
2008	3.55	7.5	3.67	14.72
2009	1.44	6.54	7.8	15.78
2010	6.58	9.54	4.6	20.72
2011	5.58	2.29	1.56	9.43
2012	2.58	2.39	1.64	6.61

Appendix B: Topeka Monthly Nighttime Rainfall Totals (1950-2012)

Year	May	June	July	Seasonal
1950	5.14	2.98	9.92	18.04
1951	2.01	6.88	8.31	17.2
1952	3.15	0	0	3.15
1953	2.5	0.45	0.81	3.76
1954	2.51	2.94	4.54	9.99
1955	3.36	3.4	1.35	8.11
1956	1.52	4.24	3.2	8.96
1957	3.33	3.24	4.32	10.89
1958	0.31	2.41	7.17	9.89
1959	0.36	0	1.79	2.15
1960	0.3	1.56	0.44	2.3
1961	2.43	0.63	4.48	7.54
1962	0.96	2.17	2.22	5.35
1963	1.23	1.45	2.49	5.17
1964	0.26	4.03	1.07	5.36
1965	0.28	6.59	3.13	10
1966	0	5.83	3.8	9.63
1967	3	10.6	1.64	15.24
1968	2	1.57	9.75	13.32
1969	0.31	1.96	2.09	4.36
1970	3.4	1.37	0.43	5.2
1971	1.47	0	3.31	4.78
1972	0	0	3.5	3.5
1973	0.49	1.34	7	8.83

1974	0.73	1.44	1.73	3.9
1975	2.51	0	0.49	3
1976	1.62	0	0.69	2.31
1977	2.6	10.32	1	13.92
1978	3.87	3.37	0.75	7.99
1979	1.09	4.43	1.11	6.63
1980	2.61	0	0.41	3.02
1981	3.15	6.12	4.86	14.13
1982	5.62	3.72	2.3	11.64
1983	3.1	1.97	0	5.07
1984	2.44	6.56	1.09	10.09
1985	2.3	0.72	1.7	4.72
1986	4.55	0.48	2.28	7.31
1987	1.83	0.67	1.68	4.18
1988	0.38	2.99	0	3.37
1989	2.48	1.51	3.42	7.41
1990	2.24	3.3	0.74	6.28
1991	1.95	3.3	0.38	5.63
1992	1.16	0.96	3.33	5.45
1993	2.94	0	5.88	8.82
1994	0.36	2.78	1.41	4.55
1995	4.64	1.27	3.27	9.18
1996	4.88	3.58	1.26	9.72
1997	2.01	0.27	1.29	3.57
1998	1.01	3.28	6.35	10.64
1999	3.13	3.95	0	7.08
2000	1.23	3.19	1.64	6.06

2001	0.38	4.93	1.35	6.66
2002	0	2.95	0	2.95
2003	2.5	3.34	0	5.84
2004	2.97	4.48	5.69	13.14
2005	0	4.91	0.59	5.5
2006	2.7	0.29	2.1	5.09
2007	8.51	1.65	0.34	10.5
2008	2.35	2.85	1.41	6.61
2009	0	4.32	5.49	9.81
2010	2.9	3.8	1.17	7.87
2011	2.14	0.99	0.6	3.73
2012	0.72	2.02	1.32	4.06

Appendix C: Rainfall Events >1.5 inches (38.1)

Day	Month	Year	Amount (inches)	Time
6-7th	May	2007	4.1	10pm-8am
6	May	2007	3.62	6am-2pm
31	May	1978	3.19	1am-2am, 7-9am
12-13th	May	2010	2.9	6pm-3am
25-26	May	1955	2.69	11pm-9am
11-12th	May	1982	2.64	11pm-5am
27	May	1984	2.44	8am-5pm
31-1	May-June	1996	2.39	7pm-10am
18-19	May	1950	2.36	10pm-4am
5-6th	May	1982	2.25	3pm-9am
26	May	1950	2.03	2am-8am
26	May	1980	1.91	8am-10am
8	May	1989	1.84	7am-9am
16-17	May	1986	1.75	8pm-9am
10	May	2003	1.67	2am-4am
16	May	1957	1.64	3am-2pm
22-23	May	1995	1.58	11pm-10am
3	May	1999	1.52	9am-1pm
26	June	1951	2.07	2am-9am
2	June	1954	1.58	2am-12pm
25	June	1955	1.52	3am-5am
17	June	1956	1.69	7am-8am
30	June	1956	1.6	2am-5am

12	June	1964	1.8	1am-6am
28	June	1965	2.1	12am-6am
12-13th	June	1966	2.4	9pm-3am
26	June	1966	1.65	2am-11am
20-21	June	1967	5.28	6pm-6am
11-12th	June	1967	2.67	9pm-3am
24	June	1967	1.51	1am-9am
18-19	June	1977	2.56	10pm-7am
12	June	1977	2.07	1am-3am
17-18	June	1977	2.06	11pm-8am
21	June	1977	1.72	1am-6am
19-20	June	1978	3.37	10pm-6am
28	June	1979	2.53	2am-9am
7-8th	June	1979	1.54	9pm-8am
27	June	1981	1.84	1am-6am
21	June	1981	1.6	4am-8am
21	June	1984	4.15	3am-11am
29-30	June	1988	2.99	9pm-9am
22	June	1989	1.51	2am-5am
15	June	1990	1.66	9am-12pm
8	June	1994	2.17	1am-4am
5-6th	June	1996	2.65	10pm-10am
8	June	1998	1.56	1am-10am
28	June	1999	1.67	1am, 7am-8am
20	June	2000	2.38	3am-7am
19-20	June	2001	2.37	5pm-4am
4	June	2002	2.95	9am-1pm

23	June	2003	1.61	4am-8am
27	June	2004	1.51	4am-12pm
3	June	2005	2.95	7am-12pm
10-11th	June	2005	1.68	10pm-11am
15-16	June	2009	1.61	9pm-5am
13-14th	June	2010	1.56	11pm-2am
20-21	June	2012	2.02	11pm-2am
12	July	1951	4.18	12am-12pm
18-19	July	1973	3.88	10pm-8am
19-20	July	1995	3.27	10pm-3am
26-27	July	1981	3.19	10pm-7am
28	July	2009	2.99	6am-2pm
20	July	1973	2.87	12am-4am
10	July	1957	2.83	1am-10am
15-16	July	1968	2.82	11pm-6am
3	July	1950	2.8	12am-5am
29-30	July	1998	2.72	11pm-6am
24	July	2004	2.46	1am-1pm
20-21	July	1961	2.32	9pm-10am
2	July	1954	2.25	12am-5am
25-26	July	1968	2.14	11pm-7am
9	July	1950	2.05	3am-10pm
11	July	1951	2	12am-11am
10	July	1982	1.89	12am-4am
26	July	1961	1.84	7am-11am
26	July	1998	1.8	4am-9am
25	July	1974	1.73	8am-11am

31	July	1989	1.72	12am-7am
5	July	1979	1.71	12am-6am
15	July	1989	1.7	9am-12pm
30	July	1968	1.69	5am-6am
31	July	1968	1.67	2am-9am
9-10th	July	1993	1.66	10pm-8am
17	July	1950	1.62	5am-8am
10-11th	July	1958	1.62	9pm-3am
16	July	1950	1.5	12am-7am

Appendix D: Wind Characteristics for May at 37.5˚N-97.5˚E for 0z (1800LST)

925mb		850mb		700mb	
Direction	Speed (m/s)	Direction	Speed (m/s)	Direction	Speed (m/s)
159	11.7	180	13.5	213	9.9
22	7.3	13	5	251	10.3
140	14.8	161	18.7	197	19.1
92	8.9	116	9	200	11.1
174	8.2	193	10.5	244	7.9
151	8.2	169	7.9	232	9
25	3.1	191	1.5	220	13.1
146	10.5	165	13.5	192	14
108	9.5	152	8.4	259	11.9
76	2.5	152	1.7	212	9.3
193	5.9	225	6.1	306	11.5
151	9.4	167	11.5	228	13.6
158	10.3	193	9.3	235	5.9
177	10.5	195	13.5	223	16.2
130	6	170	7.8	218	9.7
162	13.7	182	20.1	191	24.2
141	11.9	166	15.3	188	16.4
166	6.8	192	10.4	212	18.5

Appendix E: Wind Characteristics for May at 37.5˚N-97.5˚E for 6z (0000LST)

925 mb		850 mb		700 mb	
Direction	**Speed (m/s)**	**Direction**	**Speed (m/s)**	**Direction**	**Speed (m/s)**
185	12.6	212	15.3	241	10.2
17	7.2	51	3.8	194	8.4
151	15.3	173	23.2	196	20.4
112	10.6	146	13.5	188	17.3
177	13.3	196	14	196	7.38
176	14.1	193	15.4	227	11.9
310	0.9	208	4	216	12.6
166	13.9	180	18.9	185	13.3
140	11.6	182	13	254	10.9
1	4	15	2	208	3.2
197	11	233	10.5	291	11.5
152	12.3	180	15	223	11.2
170	12.1	192	11.4	199	5.4
189	14.4	213	17.3	249	14.4
151	9	190	14.2	216	18.7
171	10.4	187	24.4	200	24.3
168	13.5	182	17.7	191	18.2
230	4	226	10.9	226	20.3

Appendix F: Wind Characteristics for May at 37.5˚N-97.5˚E for 12z (0600LST)

925 mb		850 mb		700 mb	
Direction	Speed (m/s)	Direction	Speed (m/s)	Direction	Speed (m/s)
254	5.8	301	7.94	266	8.4
360	7.5	23	3.9	199	11.6
161	13.2	189	19.3	210	16.9
126	12	155	17.6	185	20.2
201	11.5	216	14.5	211	10.6
188	11	204	14.2	219	10.9
334	7.9	344	6.4	247	5.4
147	6	180	12.3	205	13.5
165	13.1	207	17.4	261	13.5
11	11.6	18	10.1	288	4.7
237	11.1	264	12	285	10.7
205	10.8	178	6.6	57	4
345	2.8	334	3.9	284	4
196	12.4	221	15.7	246	12.9
183	6.1	227	12.4	232	21.5
183	8.9	204	15.6	211	19.6
234	4.1	235	9	222	16.7
325	8.7	287	8.2	247	20.2
214.2	9.1	210.4	11.5	226.4	12.5

Appendix G: Wind Characteristics for May at 37.5˚N-97.5˚E for 18z (1200LST).

925 mb		850 mb		700 mb	
Direction	Speed (m/s)	Direction	Speed (m/s)	Direction	Speed (m/s)
352	8.3	358	10.7	275	8.7
346	7.5	328	2.8	248	9.2
185	12.9	204	17.2	218	18.4
148	10	181	15.3	211	18.4
184	5.9	202	8.6	217	11.8
176	9.6	200	10.4	213	10.8
330	8.2	335	6.9	242	7.8
139	10.3	172	14.7	185	12.7
179	9.1	222	12.3	249	13.1
337	12.8	352	11.5	277	4.7
227	3.4	283	6.4	274	10.5
234	0.9	235	5.1	240	15.4
325	4.5	330	5.2	271	9.5
196	12.4	212	19.3	217	16.8
258	5.3	261	8.9	238	19
134	9.3	177	12.2	193	17.9
216	5.1	228	6.7	243	15.5
321	4.5	285	4.6	244	14.8

Appendix H: Wind Characteristics for June at 37.5˚N-97.5˚E for 0z (1800LST)

925 mb		850 mb		700 mb	
Direction	**Speed (m/s)**	**Direction**	**Speed (m/s)**	**Direction**	**Speed (m/s)**
183	13.3	208	16.1	251	14.4
138	9.6	179	10.4	228	13.7
160	7.6	194	8.9	259	9.2
153	11.2	179	12.4	244	9.0
165	10.0	186	9.9	261	6.1
184	9.7	201	11.6	236	13.7
177	11.2	191	12.9	225	12.8
132	4.3	142	5.5	188	4.9
173	10.1	186	10.3	234	4.9
174	9.8	195	14.4	216	20.5
165	8.8	190	10.0	200	6.2
149	7.0	179	8.3	244	11.9
135	4.2	147	4.6	305	1.2
127	3.0	178	3.3	261	6.3
149	4.7	176	5.7	244	10.6
144	5.6	160	8.2	218	5.6
176	11.3	196	13.4	261	11.8
144	3.6	170	3.4	222	6.1
186	6.4	206	7.5	291	5.5
155	7.3	183	8.7	254	10.4
158	8.2	179	8.7	262	6.4
161	7.2	195	8.0	229	10.0
176	8.3	203	11.0	234	9.8
120	2.2	156	3.9	192	7.3
150	11.8	166	13.0	229	7.5

103	6.4	171	2.5	282	10.4
158	8.0	182	7.9	270	10.4
161	13.6	175	18.0	221	19.8
156	5.5	213	6.9	279	11.2
177	15.7	194	18.9	231	15.4
97	5.6	145	6.1	217	10.6
210	10.7	218	12.3	217	9.0
161	11.8	185	14.7	237	12.1
139	6.5	145	8.2	171	2.6
141	8.7	169	11.3	227	10.4
116	3.1	157	7.0	214	12.2
154	8.2	181	9.3	232	11.1
172	5.9	197	9.6	222	14.1
173	6.5	192	8.7	219	7.9
155	8.0	181	9.5	237	9.8

Appendix I: Wind Characteristics for June at 37.5°N-97.5°W for 6z (0000LST)

925 mb		850 mb		700 mb	
Direction	Speed (m/s)	Direction	Speed (m/s)	Direction	Speed (m/s)
199	16.4	218	21.2	256	11.4
157	12.7	190	16.8	235	16.9
167	7.1	212	8.8	271	8.1
166	13.4	191	16.1	236	8.7
184	14.9	209	17.5	258	9.5
179	10.9	201	14.6	239	6.9
179	18.1	196	20.3	219	11.5
150	6.4	161	5.6	176	4.0
176	10.5	195	11.1	202	3.2
169	14.0	194	20.4	223	16.5
181	12.8	202	14.7	237	10.0
170	11.6	192	16.6	226	12.9
155	5.6	177	6.0	131	2.1
167	4.8	200	4.8	257	5.9
158	6.5	194	8.4	249	12.4
157	10.1	185	12.0	211	8.7
180	12.2	210	13.9	268	8.4
120	5.3	158	5.3	188	9.5
191	9.3	222	8.6	301	4.3
184	12.9	213	15.3	253	7.7
173	14.6	195	15.4	266	6.0
176	9.3	204	11.2	248	6.3
178	10.0	208	11.1	219	9.2
122	2.6	148	2.6	181	6.8
172	13.7	195	17.8	220	6.7

183	8.5	212	7.3	289	8.0
189	9.2	217	11.5	282	9.3
165	18.7	185	25.2	219	22.1
205	9.6	231	12.9	267	10.6
184	15.5	210	21.2	240	11.8
159	8.5	177	8.3	246	6.0
210	14.4	217	13.5	221	8.3
177	15.7	201	19.1	233	12.8
149	7.7	169	7.4	202	2.9
163	14.4	185	16.3	231	10.7
173	13.7	184	17.6	204	12.0
193	12.5	214	15.6	256	12.2
182	6.6	195	10.5	214	13.2
198	8.6	214	6.5	285	6.5
173	11	197	13	235	9

Appendix J: Wind Characteristics for June at 37.5˚N-97.5˚E for 12z (0600LST)

925 mb		850 mb		700 mb	
Direction	**Speed (m/s)**	**Direction**	**Speed (m/s)**	**Direction**	**Speed (m/s)**
220	14.4	242	19.1	259	10.1
199	8.4	234	13.5	254	17.3
153	2.5	255	3.5	293	9.2
179	12.5	208	15.0	236	7.7
194	13.6	224	18.0	256	9.8
201	9.5	224	12.7	239	6.9
197	11.1	218	14.8	233	8.7
59	6.3	39	7.0	314	3.5
195	6.9	221	6.7	239	2.9
190	9.7	218	15.0	234	15.0
240	6.8	256	10.1	262	11.9
225	5.5	240	12.8	255	14.6
218	1.1	232	4.8	223	7.6
76	2.8	68	2.4	243	4.9
192	5.2	226	6.4	265	11.1
183	9	207	10.9	216	11.7
225	1.4	269	4.1	290	7.1
76	1.6	156	2.0	206	9.0
222	5.8	270	6.6	309	8.5
200	10.2	233	13.9	249	10.0
192	10.7	222	13.8	244	8.5
205	10.1	225	12.2	237	5.3
222	6.3	248	9.0	250	9.6
303	4.0	278	3.6	205	9.5
190	10.9	216	16.1	223	7.4

175	5.8	227	6.0	269	11.5
188	2.7	239	6.6	263	7.9
161	16.5	196	24.0	230	23.9
230	7.0	261	9.9	271	11.1
204	11.8	242	13.7	244	12.0
176	3.9	214	5.4	281	6.3
221	14.1	223	14.4	218	12.0
184	13.3	207	18.9	227	13.8
177	4.3	211	4.8	203	3.3
182	11.4	207	16.0	225	10.8
220	7.0	228	9.4	220	11.1
221	5.5	255	9.7	276	11.5
217	1.0	246	6.0	239	11.1
11	2.1	14	2.9	333	7.5
188	8	215	10	250	10

Appendix K: Wind Characteristics for June at 37.5˚N-97.5˚E for 18z (1200LST)

925 mb		850 mb		700 mb	
Direction	Speed (m/s)	Direction	Speed (m/s)	Direction	Speed (m/s)
198	9.0	225	11.7	242	13.6
279	6.3	285	10.9	261	17.9
101	4.7	121	4.3	282	5.3
180	9.1	207	9.0	247	6.0
204	11.4	227	15.4	227	10.0
193	6.1	212	8.4	206	10.8
187	9.8	206	11.6	220	9.7
31	5.8	52	6.0	297	4.6
165	5.6	177	6.0	193	4.1
186	7.7	212	11.5	212	14.6
332	6.9	320	7.0	272	10.8
299	3.3	285	6.3	258	10.9
134	4.7	175	5.0	246	6.1
172	2.8	188	5.2	232	11.4
185	4.4	212	4.5	260	10.0
182	9.8	201	13.3	219	11.8
16	2.5	16	1.9	253	5.4
27	1.8	176	1.3	224	12.0
243	4.5	291	6.1	318	11.0
189	9.1	227	10.6	248	10.3
193	12.9	219	14.6	228	10.8
204	9.7	230	12.1	236	10.7
155	4.1	217	3.6	236	5.3
337	3.8	292	1.8	225	11.5
183	10.5	210	12.2	215	11.3

122	5.1	203	3.4	260	12.7
270	3.6	289	7.1	261	11.7
178	14.4	212	18.6	247	19.8
297	5.7	322	10.8	296	14.6
204	7.9	241	8.8	236	11.8
157	3.4	202	5.5	233	6.4
245	6.7	238	10.9	228	13.2
180	12.0	204	16.2	221	13.3
103	3.2	69	1.9	5	1.2
200	5.8	219	10.3	206	11.8
201	7.2	230	6.4	225	10.4
215	5.8	244	7.0	262	11.5
241	3.8	261	4.6	249	11.4
360	2.3	345	2.4	328	2.1
194	6	217	8	239	10

Appendix L: Wind Characteristics for July at 37.5˚N-97.5˚E for 0z (1800LST)

925 mb		850 mb		700 mb	
Direction	Speed (m/s)	Direction	Speed m/s)	Direction	Speed (m/s)
161	6.4	172	6.5	288	3.9
158	10.3	174	10.3	247	6.5
168	11.6	189	11.4	242	5.4
173	11.4	203	12.3	253	8.6
145	7.9	173	10.1	226	11.4
176	9.6	187	10.2	219	10.1
136	6.5	155	6.6	238	3.4
197	9.4	214	11.9	257	10.0
172	7.6	179	9.2	223	7.1
188	8.9	204	7.6	231	1.3
180	15.4	187	19.0	222	16.2
158	3.2	187	3.8	255	3.8
158	7.7	179	8.1	265	6.7
185	11.0	199	12.6	234	6.9
185	8.9	192	9.3	230	7.4
172	9.1	186	11.6	231	13.6
121	5.7	135	4.9	291	4.4
161	2.4	187	4.1	210	3.2
135	4.9	152	5.8	225	8.3
182	8.4	201	9.7	260	8.8
140	6.2	160	5.1	312	3.1
196	4.6	227	5.0	243	3.3
163	5.9	182	8.2	223	10.3
176	8.1	200	7.2	257	7.0
142	7.8	172	7.6	240	10.7

155	6.8	185	8.7	225	11.0
60	5.4	88	4.9	233	3.6
162	2.5	197	1.0	342	0.6
161	8	181	8	247	7

Appendix M: Wind Characteristics for July at 37.5˚N-˚E for 6z (0000LST)

925 mb		850 mb		700 mb	
Direction	Speed (m/s)	Direction	Speed (m/s)	Direction	Speed (m/s)
191	9.9	210	11.2	255	4.6
175	12.1	203	14.5	254	8.2
181	12.2	210	16.2	219	7.7
179	12.2	204	15.4	215	7.3
156	9.3	187	12.7	215	11.0
183	12.3	201	15.2	233	9.7
150	10.7	176	9.8	213	4.0
181	11.8	209	14.0	247	10.2
175	6.3	201	8.1	237	5.0
194	10.1	216	8.8	184	4.1
178	19.2	193	23.7	193	13.7
180	11.5	203	10.0	203	3.1
170	14.7	198	15.9	198	6.7
179	11.4	201	13.1	201	5.2
189	12.5	202	13.7	202	7.6
176	15.4	195	19.6	195	13.4
152	10.8	186	8.3	186	6.1
141	8.7	168	8.5	168	4.1
153	6.8	176	10.2	176	9.1
193	8.3	220	9.9	220	7.1
171	8.3	195	6.0	304	2.5
188	8.0	202	5.3	231	2.7
190	11.2	205	15.0	228	9.7

197	7.7	227	6.9	278	7.7
161	10.3	194	12.5	235	13.5
149	11.4	186	11.3	234	9.2
82	6.6	118	6.4	217	2.6
197	3.4	196	2.2	311	2.1
172	10	196	12	223	7

Appendix N: Wind Characteristics for July at 37.5˚N-97.5˚W for 12z (0600LST)

925 mb		850 mb		700 mb	
Direction	**Speed (m/s)**	**Direction**	**Speed (m/s)**	**Direction**	**Speed (m/s)**
234	8.1	258	9.5	278	6.2
183	11.4	218	13.3	269	7.6
193	7.7	243	10.3	234	5.9
209	8.5	245	10.7	240	7
179	8.9	208	13	234	10.7
198	12.3	218	16.2	225	9.2
195	6	216	8.6	229	5.5
217	7.2	252	10.3	283	11.4
209	2.9	230	4.5	230	5.5
225	7.2	239	5.1	223	5
195	17.1	210	22.1	222	12
222	10.7	238	10.7	240	5.9
202	13.7	233	17.5	279	8.5
231	4.1	237	7.4	236	7.3
206	7.4	226	10.9	210	9.7
203	7.2	235	13.4	239	12.7
217	4.5	256	6.9	282	7
85	3.5	174	1.9	278	2.1
204	5.4	222	10.8	227	11.7
288	5.7	295	8.6	279	10.6
212	4.7	255	5.1	288	5.5
262	4.8	260	5.2	250	3.7
205	11.5	224	14.6	227	10.2
289	5	288	6.6	285	9.3

192	4	240	8.8	247	14.4
227	6	248	9.8	262	9.5
64	5.5	108	2.2	234	4.6
325	4.4	318	4.8	303	3
210	7.3	235.5	9.6	251.2	7.9

Appendix O: Wind Characteristics for July at 37.5˚N-97.5˚E for 18z (1200LST)

925 mb		850 mb		700 mb	
Direction	Speed (m/s)	Direction	Speed (m/s)	Direction	Speed (m/s)
270	0.4	330	2.4	294	7.0
176	7.9	224	9.1	268	5.5
182	3.3	259	6.5	276	7.4
217	4.1	252	4.1	277	6.4
174	3.9	222	6.4	242	7.2
212	8.8	230	11.9	225	9.4
169	6.7	204	6.8	236	8.0
248	4.6	269	7.3	280	11.0
214	1.8	248	4.3	255	10.1
214	6.8	226	7.8	240	3.8
183	13.2	203	15.6	219	11.5
186	4.9	223	5.7	227	7.3
183	10.8	216	13.0	239	4.8
356	2.7	322	3.7	246	8.0
187	7.2	207	8.8	218	13.5
204	6.0	237	7.7	236	11.5
93	2.2	12	2.5	313	7.1
107	4.7	166	3.2	246	4.2
210	7.3	239	10.2	252	11.7
289	5.4	306	7.9	286	15.2
150	1.6	302	0.9	300	5.9
169	1.5	258	1.9	269	4.7
198	10.5	218	12.9	233	8.5
319	3.0	316	5.9	301	10.0
170	4.8	227	5.1	250	8.6
86	2.6	329	0.6	273	7.8

33	6.7	45	4.1	288	3.2
51	1.3	29	1.8	292	7.3
187	5	226	6	260	8

Appendix P: Characteristics of May LLJs for Events >1.5 inches (38.1mm)

Date	Speed (m/s)	Level	Latitude	Longitude	Time
May 18-19 1950	18.3	850	40	265	6z
May 26 1950	13.4	700	35	267.5	12z
May 26 1955	27.3	850	35	265	12z
May 16 1957	20.4	850	32.5	97.5	12z
May 31 1978	16.7	850	35	260	12z
May 26 1980	15.8	700	37.5	260	12z
May 5-6 1982	20.5	850	35	267.5	12z
May 12 1982	24	850	32.5	260	6z
May 27 1984	17.4	850	37.5	267.5	12z
May 16-17 1986	19.7	850	32.5	265	6z
May 8 1989	16.2	850	32.5	260	6z
May 22-23 1995	21.5	850	35	265	12z
May-June 31-1 1996	17.8	850	32.5	97.5	6z
May 3 1999	18.5	850	40	265	6z
May 10 2003	25.7	850	35	265	12z
May 6 2007	26.3	700	40	97.5	6z
May 6-7 2007	19.3	850	37.5	265	6z
May 12-13 2010	26.1	700	37.5	267.5	12z

Appendix Q: Characteristics of June low-level winds for events >1.5 inches (38.1mm)

	Speed	Level	Latitude	Longitude	Time
June 26 1951	23.2	850	35	97.5	12z
June 2 1954	20	700	37.5	267.5	18z
June 25 1955	15.5	925	32.5	97.5	6z
June 17 1956	18	850	35	97.5	6z
June 30 1956	18	850	37.5	97.5	12z
June 12 1964	17.1	850	32.5	97.5	12z
June 28 1965	20.7	850	35	260	12z
June 13 1966	13	925	35	267.5	6z
June 26 1966	14.7	850	35	260	6z
June 12 1967	22.7	850	37.5	265	12z
June 20-21 1967	15.6	925	35	265	6z
June 24 1967	21	850	32.5	260	6z
June 12 1977	12.3	925	32.5	97.5	6z
June 17-18 1977	17.7	925	32.5	97.5	6z
June 18-19 1977	18.2	850	32.5	260	6z
June 21 1977	18.8	850	32.5	260	6z
June 19-20 1978	17.7	850	32.5	260	6z
June 7-8 1979	17.3	925	32.5	265	6z

June 28 1979	11.9	850	35	260	12z
June 21 1981	20.6	850	35	97.5	12z
June 27 1981	16.5	850	37.5	260	6z
June 21 1984	14.5	850	35	260	6z, 12z
June 29-30 1988	13.5	925	32.5	97.5	6z
June 22 1989	16.1	850	32.5	260	12z
June 15 1990	18.9	850	35	97.5	6z
June 8 1994	17	925	32.5	265	12z
June 5-6 1996	21.4	925	35	267.5	6z
June 8 1998	27.2	850	35	97.5	12z
June 28 1999	18.4	925	35	267.5	6z
June 20 2000	22.3	850	37.5	265	6z
June 19-20 2001	12.9	850	32.5	97.5	6z
June 4 2002	16.2	925	32.5	97.5	6z
June 23 2003	19.5	850	35	260	6z
June 27 2004	9.7	850	37.5	260	0z
June 3 2005	17.4	850	32.5	260	6z
June 10-11 2005	18.8	850	35	97.5	6z
June 15-16 2009	20.8	850	32.5	97.5	6z
June 13-14 2010	19.4	850	32.5	97.5	6z

Appendix R: Characteristics of July Low-level Winds for Events >1.5 inches (38.1mm)

Date	Speed (m/s)	Level	Latitude	Longitude	Time
July 3 1950	12.8	850	32.5	260	6z
July 9 1950	14.8	850	35	260	6z
July 16 1950	22.9	850	35	97.5	6z
July 17 1950	22	850	32.5	97.5	6z
July 11 1951	17.9	850	32.5	260	6z
July 12 1951	19.2	850	32.5	97.5	6z
July 2 1954	18	850	40	97.5	12z
July 10 1957	11.9	925	40	97.5	6z
July 10-11 1958	16.8	850	35	260	6z
July 20-21 1961	15.7	925	32.5	265	12z
July 26 1961	12.6	850	40	260	0z and 6z
July 15-16 1968	25	850	35	97.5	6z
July 25-26 1968	13.1	925	35	97.5	6z
July 30 1968	17.4	850	37.5	260	6z
July 31 1968	18.1	925	32.5	260	12z
July 18-19 1973	16	925	32.5	97.5	6z
July 20 1973	21.2	850	35	97.5	6z
July 25 1974	14.8	850	32.5	260	6z
July 5 1979	12.5	925	32.5	97.5	6z
July 26-27 1981	16.8	850	32.5	97.5	6z
July 10 1982	16.9	850	32.5	260	6z
July 15 1989	10	925	32.5	97.5	6z
July 31 1989	11.4	925	35	97.5	6z
July 9-10 1993	16.3	850	35	260	6z
July 19-20 1995	12.8	850	32.5	260	6z

July 26 1998	14.9	700	37.5	265	12z
July 29-30 1998	17.2	925	32.5	267.5	12z
July 24 2004	11.9	925	40	267.5	12z
July 28 2009	9.9	925	32.5	97.5	6z