Spatio-Temporal Variability of Precipitation and Drought in the State of Arizona, USA

Nyarkoh

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SPATIO-TEMPORAL VARIABILITY OF PRECIPITATION AND DROUGHT IN THE STATE OF ARIZONA, USA

by

Samuel Nimako Nyarkoh

A thesis submitted to the Graduate College in partial fulfilment of the requirements for the degree of Master of Science Geography
Western Michigan University
June 2018

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SPATIO-TEMPORAL VARIABILITY OF PRECIPITATION AND DROUGHT IN THE STATE OF ARIZONA, USA

Samuel Nimako Nyarkoh M.S.
Western Michigan University, 2018

As a result of climate change, several dry regions continue to get drier as precipitation amounts decline. This decline impact water resources and hinder economic development. Understanding the variability of precipitation and drought through climatic, agricultural and hydrological studies is therefore critical to decision makers and stakeholders in developing proactive measures that promotes economic development.

This study therefore uses dataset from 55 Meteorological stations containing 110-year (1900-2010) monthly precipitation data and a series of spatial and temporal tests to investigate the spatial and temporal patterns of precipitation and drought as well as the effects of local factors such as topography and vegetation on Arizona precipitation and drought.

Result shows that the western half of Arizona is mostly dry while the eastern part is mostly wet. Over, 60% of Arizona’s drought is ‘Normal drought’ and 99% of this ‘normal’ drought occur in the southeastern portion of Arizona. Monsoon and Non-monsoon rainfall also affects Arizona differently. Summer precipitation affects mostly the eastern fringes of Arizona, including the South east (Climate Division 7 or CD7) and the south east of the Colorado Plateau, while Winter precipitation or Non- Monsoon precipitation affects mostly the Central highlands (especially Gila County or CD4).
ACKNOWLEDGEMENTS

This project cannot be deemed complete without thanking the wonderful people who contributed to making this possible. First and foremost, I want to say thank you to my committee chair professor Dr. Laiyin Zhu, for his time and patience in guiding me throughout this research. His constant advice and edits to this manuscript was priceless. Another appreciation goes to my committee members Dr. Benjamin Ofori-Amoah and Dr. Chansheng He, for their contribution in the review and organization of this project.

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Finally, I would like to thank my beloved wife Tameca Nyarkoh, for her patience and support by sacrificing a lot of the fun time for me to work on this manuscript. Thank you.

Samuel Nimako Nyarkoh
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CHAPTER I

INTRODUCTION

Water remains the most precious resource on the planet and is very crucial for supporting human civilization, economic development, and ecosystem services. While multiple demands compete for water, only 1% of the earth’s water is freshwater, which is to be shared among over 7 billion people. The distribution of this freshwater is not uniform. Some regions continue to be wetter than others and other places also continue to be more dryer than others. Several studies investigating the precipitation patterns or variability within a region have been made. For example, Krepper et al. (1989) determined the spatial and temporal variability of rainfall in Argentina by analyzing monthly and seasonal rainfall. Rowell et al. (1995) also examined rainfall data in the north of Africa by using spectral analysis and obtained peaks at 3–8 years in Sudan and at 2–3 years in Guinea.

This research is focused on Arizona because Arizona has an arid climate and is characterized by high degree of interannual and decadal variability (Dettinger et al. 1998). Precipitation variability in this area is of concern to society because of its effect on human activities and economic development. Arizona faces constraints on its water supply more than any other state in the U.S. This made them resort to groundwater for their water needs. In 1980, Arizona Department of Water Resources was established to oversee and control water usage. Currently, Arizona relies on the Colorado river and Winter and Summer precipitation as their main source of water supply. Reports indicate that the Colorado river has experienced extensive drought conditions in recent years. This has impacted the level of Lake Mead to a historically low reservoir level. Projections have shown that if no action is taken to address the gap between
water supply and demand, Lake Mead could reach a critical stage within the next few years, and this will trigger a larger mandatory restriction on the use of the Colorado water. If this happens, it could have a devastating impact on Arizona’s agriculture, environment, and economy (Western resources advocate, 2017). Places distant from the Colorado river relies mainly on precipitation for their water needs, but precipitation is not consistent, some regions receive more precipitation than others and other places also continue to be more dryer than others. Thus, investigating the spatial and temporal variations in precipitation and the patterns existing within the climate divisions will open the doors for proactive and appropriate adaptation strategies to be implemented.
Problem Statement

Arizona is generally dry, but some places remain dryer than others and precipitation amounts vary spatially across the region. Previous research has attributed the variations in precipitation over the Southwest of U.S.A to monsoon or airmasses originating from the Pacific and Gulf of California (Houghton. 1970), topography (Despain.1987; Bartlein. 1993; Mock. 1996), and Large scale atmospheric factors such as ENSO (Zhang et al. 1997; Evans et al. 2001). These studies were conducted at larger scales like the Southwest of USA, Arizona and New Mexico etc.

However, the variations existing within the individual Climate Divisions(CD’s) are not yet fully explored. This study delves deeper into investigating the behavior and trends existing within the precipitation dataset as well as the patterns within the Climate Divisions(CD's).

Subsequently, the results are expected to classify which Climate Divisions have the longest drought occurrences and the distribution of normal, moderate, severe and extreme drought within the CD’s by using a series of temporal and spatial analysis tests.
CHAPTER II

REVIEW OF LITERATURE

Precipitation Pattern

Precipitation in the western region of the United States varies spatially because of the numerous small-scale climatic controls embedded within the larger scale controls, mainly due to the topography of the region (Bryson and Hare 1974; Hirschboeck 1991). It is demonstrated that because of the varied topography of the Southwest United States, several topographic variables exert influences on spatial patterns of temperature and precipitation. Similarly, numerous attempts to model precipitation by statistical prediction have been made to model precipitation in many places (e.g. Hevesi et al. 1992, Bigg 1991, Prudhomme and Reed 1999, de Montmollin et al. 1980, Kilsby et al. 1998, Ffolliott et al. 1989).

Several studies have examined the spatial variations of precipitation over the United States and investigated the climatic controls responsible for it (Horn et al. 1957). Tang and Reiter (1984) also analyzed spatial patterns of inter-monthly changes of precipitation. Cayan and Roads. (1984) also investigated monthly precipitation along the West Coast in relation to atmospheric circulation patterns. These researchers demonstrated the effect of large scale atmospheric controls influencing precipitation variations spatially. Two precipitation peaks namely, winter and summer were identified for the states west of the Great Plains region (Mitchel 1976).

Winter precipitation is produced from large frontal systems moving over the Southwest (McPhee et al. 2004). Summer precipitation on the other hand results from thunder storms within
the North American monsoon circulation. In an attempt to understand trends and patterns in Arizona precipitation series, researchers have analyzed the effect and linkage between Arizona precipitation and large scale atmospheric forcing such as El Niño Southern Oscillation (ENSO), Pacific decadal oscillation and the North American monsoon (Mantua and Hare 2002; Mantua et al. 1997). The link between Arizona winter precipitation and ENSO is stronger for the La Niña phase, which is characterized by cool central sea surface temperatures (SST’s) and dry winter in Arizona than the El Niño phase which is characterized by warm central Pacific SST’s and sometimes wetter Arizona winters.

The complex topography and its geographical proximity to the Pacific Ocean, the Gulf of California, and the Gulf of Mexico are known to contribute to this region’s high climatic variability. Houghton (1970) studied the seasonal and spatial characteristics of precipitation in the Great Basin and found that precipitation variations results from airmasses originated from the Pacific and the Gulf of California and also from the Great basin cyclogenesis. He also emphasized the effect of topographic concerns. Mock (1996) found that a number of climatic controls interact with one another to jointly explain spatial precipitation variations. The Southwest of the United States is located between the mid-latitude and the subtropical atmospheric circulation regimes. Location shifts in these atmospheric circulation regimes are found to be the fundamental reason for the region’s climatic variability. Studies of sea surface temperatures (SST) in the North Pacific have highlighted the presence of Pacific decadal oscillation (PDO) as the leading mode of variability, particularly on decadal timescales (Mantua et al. 1997).
Winter and Summer Precipitation

During winter, occasional cyclonic storms that attain very large sizes typically enter North America over the states of Washington and Oregon. These westerly storms are capable of producing intense precipitation over Arizona. Strong Pacific/North American (PNA) patterns are known to be linked to above average precipitation in the Southwest (Redmond and Koch 1991). The southwestern trough is a strong low-pressure trough positioned over the Southwest with concurrent high-pressure ridges found over the Gulf of Alaska and the Great Lakes. This trough is known to be weakly related to PNA (Woodhouse 1997). Over sixty percent of January precipitation totals are attributed to the southwestern trough, though the total percentage of precipitation related to southwestern trough is lower in eastern New Mexico (Burnett 1994).

El Niño–Southern Oscillation (ENSO) is known to be associated with winter precipitation in the southwest (Zhang et al. 1997; Evans et al. 2001). Pacific decadal oscillation (PDO) has been considered the leading mode of variability, particularly on decadal timescales (Mantua et al. 1997). However, PDO is known to have very little multiyear persistence during summer, thus decadal variability of North Pacific SST is largely a winter/spring phenomenon (Newman et al. 2003). Localized orographic factors are also known to affect precipitation (Jorgensen et al. 1967).

During summer, and around June each year, North American monsoon brings moisture to most of Arizona. This moisture originates from a combination of oceanic sources including; the Gulf of California, the eastern tropical Pacific Ocean, and the Gulf of Mexico (Adams and Comrie 1997; Higgins et al. 1997). Convective storms occur when local conditions cause these moistures to ascend. Monsoon is defined as a distinctive seasonal change in wind direction of at least 120 degrees. (Tang and Reiter 1984). Climate variation within the region also results from
overall topographic relief and from proximity to the moisture sources of the Gulf of Mexico, the Gulf of California, and the eastern Pacific Ocean. Additionally, high temperatures, high rates of evapotranspiration, and rain shadow effects of mountain ranges all contribute to regional aridity (Scott 1991).

Commonly Used Precipitation Indexes

Drought has several definitions because there are several forms of drought. However, the most basic definition of drought is that of an episode of unusually low precipitation that causes damage to ecosystems, agriculture, and freshwater supplies. Due to the ununiform pattern of droughts, there are many other drought definitions specific to each sector. The commonly studied droughts include agricultural, meteorological, hydrological, and socioeconomic drought (Heim 2002). Due to different interpretations of drought, several indices have been created in the attempt to quantify drought specific to a given definition. However, no drought index has been able to fully satisfy all definitions of drought, thus several drought indices are used. These include the Palmer Drought Severity Index (PDSI) (Dai 2011), and the Crop Moisture Index (CMI) (Palmer 1968), the Standardized Precipitation Index (SPI) (Vasiliades et al. 2011), the Soil Moisture Drought Index (SMDI) etc.

Palmer Drought Severity Index (PDSI) and the Standardized Precipitation Index (SPI) are the drought indices commonly used in analyzing drought. PDSI was the first drought index developed in the United States and was one of the most widely used and easily available indices. It measures the departure from normal of the moisture supply and mostly used as a meteorological drought index (Palmer 1965). PDSI requires temperature and precipitation data in
its computation and allows for other terms in the basic water balance equation to be determined. The derived parameters from PDSI include soil recharge, loss of water from the surface and subsurface layers, runoff and evapotranspiration. (Refer to Heim (2002) for detailed explanation and derivation of PDSI). However, it should be noted that PDSI has some limitations. These limitations include (i) the mark or level indicating a drought condition or wet spell is arbitrary and have little scientific validity. (ii) It is unstandardized; thus, it is incomparable across regions (Guttman et al 1992). (iii) There is a lag in predicting drought conditions and has no predetermined time-scale, which can be misleading. (iv) and it is unable to predict accurately locations with climatic extremes and mountainous areas (Alley 1984). The SPI was then developed to quantify precipitation deficits over multiple time scales, with time steps ranging from one month to seventy-two months.

The application of these different time scales allows measurement of the impact of drought on various water resources (McKee et al. 1993). Essentially, SPI is an index of the standard deviation of a given precipitation deficiency. While the PDSI measures both input and output side of the water balance equation, the SPI measures only the input and does not consider seasonal differences in output variables such as evapotranspiration. Drought is mostly represented in terms of drought variables (Mishra and Singh 2010). These include drought intensity, drought frequency, and duration.

Spatial Interpolation Methods

Rainfall records are mostly incomplete because of missing rainfall data within the measured period, or in other words, insufficient rainfall stations within the study area (Chen and
Liu. 2012). Probable rainfall data can be estimated through spatial interpolation techniques. Inverse-distance weighting (IDW) is one of the most frequently used deterministic models in spatial interpolation. The general idea for this method is that the attribute value of an unsampled point is the weighted average of known values within the neighborhood, and the weights are inversely related to the distances between the prediction location and the sampled locations. However, several previous studies have revealed that the decline in the spatial relationship between any two locations is not simply proportional to distance (Fotheringham and O’Kelly 1989). As a result, the inverse-distance weight is adjusted by either a constant power or a distance-decay parameter to account for the diminishing strength in the relationship with increasing distance. Chen et al. (2012) conducted a study to evaluate the relationship between interpolation accuracy and two critical parameters of IDW (power and a radius of influence) in Taiwan.

The value of both ‘radius of influence’, and the control parameter were determined by ’root mean squared error’. The results revealed that the optimal distance for IDW in the interpolation of rainfall data have a radius of influence up to 10-30 km in most cases and the optimal alpha value was between zero and five. It was also identified that rainfall data obtained through interpolation using IDW, displayed more accurate results during the dry season than in the flood season. The high correlation coefficient values of over 0.95 confirmed IDW as a suitable method of spatial interpolation to predict the probable rainfall data. Most studies use a power of two (2) to interpolate precipitation (Lloyd 2005). Lu et al. (2008) developed an algorithm to search for “optimal” adaptive distance-decay parameters, using cross-validation to evaluate the results. IDW is a very simple method and does not require a theoretical distribution
for the data. The method assumes the principle of Tobler’s first law of geography (Tobler 1970) that gives more weight to near locations than distant locations.

Mann-Kendall’s Trend Test

In finding the presence of trends within a dataset, several methods could be employed but one of the methods is the Mann-Kendall’s Trend Test. It is a correlation test that measures the strength of monotonic association between the ranks of a time series and their time order. For example, DaSilva (2004) used M-K method to test for climate variability in Northeast of Brazil. The null hypothesis in the Mann-Kendall test states that data is independent and randomly ordered. However, it is worthwhile to note that the existence of serial correlation in time series can adversely affect the power of the trend tests (Yue and Wang 2002; Yue et al. 2002; Yue and Hashino 2003; Khaliq et al. 2009). Thus, it is necessary to investigate the effect of serial correlation on trend detection tests. In order to eliminate the effect of serial correlation, Prewhitening is mostly used to eliminate the influence of serial correlation on the Mann-Kendall (MK) test in trend-detection studies of hydrological time series. However, its ability to accomplish such a task was never really taken into consideration.

Yue and Wang (2002) conducted a study to investigate this issue by the application of Monte Carlo simulation. The simulation results demonstrated that when trend exists in a time series, the effect of positive or negative serial correlation on the Mann-Kendall (MK) test is dependent upon sample size, the magnitude of serial correlation, and magnitude of the trend. Thus, when the sample size and magnitude of the trend are large enough (say n ≥ 50), serial correlation no longer significantly affects the MK test statistics. Hence, removal of positive lag 1
autoregressive (AR (1)) from time series by prewhitening will remove a portion of trend and hence reduces the possibility of rejecting the null hypothesis while it might be false. In other words, removal of negative AR (1) by prewhitening will inflate trend and lead to an increase in the possibility of rejecting the null hypothesis while it might be true. They concluded that, prewhitening is not suitable for eliminating the effect of serial correlation on the Mann-Kendall’s (MK) test when trend exists within a time series.
CHAPTER III
DATA AND METHODOLOGY

Study Area

Arizona is located between latitudes 31.2°- 37.0° N. It is a southwestern U.S. state with mountains and deserts. It is the sixth largest and the 14th most populous of the 50 U.S states. Its capital and largest city is Phoenix. It is bordered by New Mexico, Utah, Nevada, California, and Mexico. Southern Arizona is known for its desert climate, with very hot summers and mild winters while the Northern Arizona remain forested. Of the state's land area of 113,998 square miles (295,000 km²), approximately 15% is privately owned. The remaining area is public forest and park land, state trust land and Native American reservations. Due to its large area and variations in elevation, the state has a wide variety of localized climate conditions and divided into seven (7) climate zones (see figure 1). The wettest climate division (CD4, Gila County) receives 477.5mm of precipitation per water year, whereas the driest division (CD5, Yuma and La Paz Counties) receives 116.8mm per year (McPhee et al.2004).

The lower elevations have a desert climate with mild winters and extremely hot summers. The weather is mild from late fall to early spring, averaging a minimum of 60 °F (16 °C). The coldest months ranges from November through February, with temperatures ranging from 40 to 75 °F (4 to 24 °C), with occasional frosts. Most of the state has a bimodal precipitation peak. El Niño Southern Oscillation is also known to influence inter-annual variability (Sheppard et al. 2002). The temperatures start to rise about midway through February, with warm days, and cool, breezy nights. The summer months of June through September
(Monsoon season) bring a dry heat from 90-120 °F (32-49 °C), with occasional high temperatures exceeding 125 °F (52 °C) in the desert area. As a result of long-term variations in Pacific Ocean circulation and Atlantic Ocean circulation (McCabe et al. 2004), Arizona is prone to multi-year and multi-decade drought (Schneider and Cornuelle 2005). It has an average annual rainfall of 12.7 in (323 mm), which comes during two rainy seasons, with cold fronts coming from the Pacific Ocean during the winter and a monsoon in the summer.
Figure 1: Map of study area with climatic zones. CD stands for climate division.
Acquisition and Processing of Datasets

Research was conducted for trends in Arizona’s precipitation. A number of statistical analysis were performed on the precipitation data to establish spatial and temporal patterns within the dataset. Monthly precipitation data from fifty-five (55) meteorological stations were obtained from the National Climate Data Center (NCDC) Climate Divisional Database (https://gis.ncdc.noaa.gov/maps/clim/summaries/monthly) for this study (see sample data in Table 1). Roughly 40% of the dataset were missing and data cleaning and preprocessing have been done. Out of the 55 meteorological stations, only four stations were used for the temporal analysis and the entire 55 stations were further processed for the spatial analysis.

The first part of this study was to analyze the precipitation timeseries. The Mann-Kendall’s trend test was then used to find trends within the dataset. One most complete dataset falling within each climate division was selected to represent that climatic division. This was done to avoid excessive interpolation which may skew the results. The following values represent the percentages of missing data for the seven meteorological stations selected (CD1(65%), CD2(9.4%), CD3(2.5%), CD4(17.8%), CD5(34%), CD6(2.6%) and CD7(4%)) respectively. All stations with missing data more than 10% were again eliminated to prevent excessive interpolation. This left only four meteorological stations to be studied. Inverse Distance Weight (IDW) method was used to estimate all missing data within the remaining dataset.

Below are some of the constants and methods employed in the IDW interpolation. An exponent (power) of 2 was applied in the interpolation because Lloyd (2005) concluded from his
study that the optimal power varies between zero and five and most studies use a power of 2 to interpolate precipitation. The study used “number of neighbors” for the neighborhood size.

Though the focus of this research was in Arizona, the network of meteorological stations used in the analysis extended to neighboring portions of New Mexico, Utah, Colorado, and California. This was done to minimize possible edge effects arising from interpolation. The second, third and fourth part of the study also used the same dataset obtained for the first test. The dataset for the spatial analysis was prepared differently. The monsoon season months (June, July, August and September) were selected for the period. The dataset was organized by finding the sum of the four monsoon months in each year. Out of the four monsoon months, if more than one month was missing, the entire year was eliminated. After that process, the dataset was grouped into decades. Again, in finding the decadal sum’s, if more than one year was missing, the decade was eliminated.
Table 1: Format of Precipitation Data used in this Study

<table>
<thead>
<tr>
<th>Stn No</th>
<th>Station</th>
<th>State Code</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation</th>
<th>Period</th>
<th>%_Missing</th>
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<td>Nogales</td>
<td>AZ</td>
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<td>1161.90</td>
<td>1899-09-01-1983-06</td>
<td>36.80</td>
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<td>2</td>
<td>Ruby 4 nw</td>
<td>AZ</td>
<td>31.50</td>
<td>-111.28</td>
<td>1214.00</td>
<td>1895-04-01-1955-12</td>
<td>75.40</td>
</tr>
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<td>3</td>
<td>Arivaca</td>
<td>AZ</td>
<td>31.57</td>
<td>-111.31</td>
<td>1112.20</td>
<td>1899-10-01-2016-10</td>
<td>50.50</td>
</tr>
<tr>
<td>4</td>
<td>Bisbee</td>
<td>AZ</td>
<td>31.43</td>
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<td>1617.60</td>
<td>1895-01-01-1985-02</td>
<td>47.90</td>
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<td>Fail huachuca sierra vista municipal airport</td>
<td>AZ</td>
<td>31.58</td>
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<td>1438.40</td>
<td>1900-02-01-2016-04</td>
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<td>Tombstone</td>
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<td>1898-06-01-1975-05</td>
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<td>AZ</td>
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<td>1899-02-01-1954-12</td>
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<td>742.20</td>
<td>1895-01-01-2008-04</td>
<td>8.10</td>
</tr>
</tbody>
</table>
Temporal Analysis

Trend Analysis Methods

To determine if trends exist in the precipitation data, the annual precipitation data for this study was grouped into four periods (1900-1930 1931-1961 1962-2010 and 1900-2010) and investigated for statistical trends using Mann–Kendall trend test. The grouping of the data was arbitrary and done to investigate 30-year trend within the data. Two main tests, namely parametric and non-parametric tests were used in the detection of trends in this study. The parametric trend tests require data to be independent and normally distributed, while the non-parametric trend tests require that the data be independent. The Mann–Kendall trend test was used to find the trends in each period and the direction of the slopes tested by the Sen’s slope test. This was done by the application of a Mann-Kendall’s and Sen’s slope test in MATLAB, using an alpha or (p) value of 5% or (0.05).

The Mann–Kendall Trend Test

The Mann–Kendall trend test (Mann 1945; Kendall 1975) mentioned above is a rank correlation that measures the strength of monotonic association between the ranks of a time series and their time order. It was designed to quantify the precipitation deficit for multiple time scales. This study used timescales of 12 months which correspond to the past 12 months of observed precipitation totals respectively. This study did not delve into the derivation of the
Mann–Kendall and Sen’s slope test but rather applied an existing MATLAB code developed by (Burkey 2006).

Below is the reasoning behind the Mann–Kendall trend test used in this study.

For time series \( X = \{x_1, x_2, x_3, \ldots x_n\} \) and \( Y = \{y_1, y_2, y_3, \ldots y_n\} \) the test statistic \( (S) \) is calculated as in Eq. (1):

\[
S = \sum_{i<j} a_{ij} b_{ij}
\]  

(1)

Where

\[
a_{ij} = \text{sign}(x_i - x_j) = \text{sign}(R_i - R_j) = \begin{cases} 1 & x_i < x_j \\ 0 & x_i = x_j \\ -1 & x_i > x_j \end{cases}
\]  

(2)

and \( b_{ij} \) is similarly defined for the observations in \( Y \). The time range in \( X \) in this study was 1900-2010 in months and \( (n = 1332) \) data points.

\( R_i \) and \( R_j \) are the ranks of the observations \( x_i \) and \( x_j \) of the time series respectively. The S statistic tends to normality for large \( n \), considering the null hypothesis that \( X \) and \( Y \) are independent and randomly ordered with mean and variance given by:

\[
E(S) = 0
\]  

(3)

\[
\text{var}(S) = n(n - 1)(2n + 5)/18
\]  

(4)

where \( n \) is the number of observation. In the case where the values in \( Y \) are replaced with the time order of the time series \( X \), i.e. 1,2,3 \ldots \( n \), the test statistic reduces to:

\[
S = \sum_{i<j} a_{ij} = \sum_{i<j} \text{sgn}[x_j - x_i]
\]  

(5)
With the same mean and variance in Equation’s (3) and (4) Kendall (1955) gives proof of the asymptotic normality of the statistic $S$. The existence of tied ranks in the data results in a reduction of the variance of $S$ to become:

$$V_0(S) = n(n - 1)(2n + 5)/18 - \sum_{j=1}^{m} t_1 (t_1 - 1)(2t_1 + 5)/18$$  \(6\)

Where $m$ is the number of groups of tied ranks. In order to test for the significance of trends, the standardized test statistic $Z = S/[\text{var}(S)]^{0.5}$ is compared with the standardized normal variate at the significance level $\alpha$, where the subtraction or addition of unity in Eqn. (6) is a continuity correction (Kendall 1975).

$$u = \begin{cases} 
\frac{(S - 1)}{\sqrt{V_0(S)}} & S > 0 \\
0 & S = 0 \\
\frac{S + 1}{\sqrt{V_0(S)}} & S < 0 
\end{cases}$$  \(7\)

A detailed discussion of the derivation of mean and variance of $S$ is given by (Kendall 1955).

Sen’s Slope Estimator

Another non-parametric procedure used in this study for trend determination is the Sen’s slope estimator. It is used for estimating the slope of trend in the sample of N pairs of data (Sen 1968)

$$Q_i = \frac{x_j - x_k}{j - k} \text{ for } i$$

$$= 1, \ldots, N,$$  \(6\)

Where $x_j$ and $x_k$ are the data values at times $j$ and $k$ ($j > k$) and number of data points $(n = 1332 \text{ observations})$. N is all data pairs for which $j$ is greater than $k$. $(Q_i)$ is the slope
estimate. For a single datum within each time period, \( N = \frac{n(n-1)}{2} \), where \( n \) is the number of time periods. For multiple observations in one or more-time periods, \( N < \frac{n(n-1)}{2} \), where \( n \) is the total number of observations. The \( N \) values of \( Q_i \) are ranked from the smallest to largest and the median of slope or Sen’s slope estimator is computed as

\[
Q_{med} = \begin{cases} 
Q_{[\frac{N+1}{2}]} & \text{if } N \text{ is odd} \\
\frac{Q_{[\frac{N}{2}]} + Q_{[\frac{N+2}{2}]} }{2} & \text{if } N \text{ is even}
\end{cases}
\] (7)

The sign of the \( Q_{med} \) indicates the data trend, while its value indicates the steepness of the trend. In order to have a detailed information about whether the median slope is statistically different than zero and the derivation of confidence intervals for Sen’s slope see (Hollander and Wolfe 1973; Gilbert 1987)

Coefficient of Variation

Following the trend detection test, the next interesting research question was to determine the variation in precipitation within each climate division. In other words, the idea was to determine which climate division had both extreme wetness and dryness. In this test standard deviation and coefficient of variation (CV) for each climate division were used to interpret precipitation variability over the period of record. The mean and standard deviation of the annual precipitation was first determined, and the coefficient of variation computed for the precipitation series. The coefficient of variation denoted by \( Cv \) (or occasionally V) is defined as the ratio of
the standard deviation to the mean. It shows the extent of variability in relation to mean of the population.

Often the coefficient of variation is expressed as a percentage which is expressed as;

\[ CV = \frac{S \times 100}{M} \]  

(1)

Where the standard deviation and the mean are denoted by S and M. The coefficient of variation is used as a statistical tool because it apparently permits the comparison of variates free from scale effects (i.e., it is dimensionless).

Multitaper Spectral Analysis

To study the periodicities within the monthly dataset for the selected climate divisions, Multitaper spectrum analysis method was employed. Red-Noise confidence levels were also computed and plotted to ascertain if any of the signal peaks were as a result of noise within the dataset. This test was also an existing MATLAB function computed on the following assumptions. If timeseries is denoted by \(x(n)\), the computation for a conventional direct spectral estimate of the power spectrum \(P_x(f)\) is found by multiplying the data \(x(n)\) by a sequence \(w(n)\) known as the taper. A Discrete Fourier Transform (DFT) is then applied then the square modulus of the resulting function is determined. The expression below shows the relationship between the power spectrum of the taper \(\hat{P}_x(f)\) and the original \(P_x(f)\) series.

\[ \hat{P}_x(f) = \int_{-\frac{T}{2}}^{\frac{T}{2}} |W(f) - (f')|^2 P_x(f') df' \]  

(1)
Where $|W(f) - \langle f \rangle|^2$ is the power spectrum of the taper and the number of observation $n = 1332$. The taper is selected by maximizing the fraction of energy of $|W(f) - \langle f \rangle|^2$ inside the interval $\Omega$ in the neighborhood of $f$.

Choosing the width of $\Omega$ involves a tradeoff between resolution and variance. After computing the tapers, the power spectrum is estimated by the expression.

$$\hat{P}_X(f) = \frac{\sum_{k=0}^{K} |d_k(f) Y_k(f)|^2}{\sum_{k=0}^{K} |d_k(f)|^2} \quad (2)$$

Where $Y_k(f)$ is the DFT of the tapered series, $K$ is the eigen taper, $d_k(f)$ is the frequency-dependent weight. Refer to the works by Thomson (1982) and Lindberg (1986) for more detailed treatments of this topic.

Red-Noise and Red-Noise Confidence Level

A MATLAB function was used for testing the amount of red noise in the signal and calculating the confidence levels using chi square test performed on the mean spectrum of the number of red-noise simulations performed within the Monte Carlo loop (1500 simulations were used in this works for more accuracy). This MATLAB code was developed by (Husson,2012). The Red-Noise and Red-Noise Confidence Level was computed by determining the discrete AR1 process $r$ for times $t_i \ (i = 1,2, ..., N)$ using the expression;

$$r(t_i) = \rho_i \ r(t_i - 1) + \epsilon(t_i) \quad (1)$$

Where $\rho_i = \text{average autocorrelation coefficient} = \exp(-\frac{t_i-t_{i-1}}{\tau})$, $\tau$ is the timescale of the AR1 process and $\epsilon$ indicates ‘white Gaussian noise’ with zero mean and variance $\sigma^2 \equiv 1 -$
$\exp\left(-\frac{2(t_i-t_{i-1})}{\tau}\right)$. Refer to the works by Schulz and Mudelsee (2002) and Thomson (1982) for more detailed treatments of this topic.

Standardized Precipitation Index (SPI)

Following the study of the precipitation timeseries above. The standardized precipitation Index (SPI) (McKee et al. 1993) was used to quantify precipitation deficit for multiple time scales. In this study, SPI was calculated on 12 months’ timescale. A standardized precipitation series is calculated using the arithmetic average and the standard deviation of precipitation series. The standardized precipitation series, $SPI_i$, is calculated as follows:

$$SPI_i = \frac{X_i - \bar{X}}{S_X} \quad (1)$$

Where $X_i$ is the monthly rainfall amount and $\bar{X}$ is the average and $S_X$ is the standard deviation of the precipitation series. A negative value obtained from this equation indicate drought events, while positive values stand for precipitation wet events. The frequencies of Normal, Moderate, Severe and Extreme drought for the four climate divisions were then obtained. Four different drought categories are defined by McKee et al. (1993) as given in Table 2 below.
Table 2: Classification and Evaluation of Different Standardized Precipitation Indices per McKee et al. (1993)

<table>
<thead>
<tr>
<th>SPI Drought Category</th>
<th>SPI Value Category</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0 to -0.99</td>
<td>Normal dry</td>
</tr>
<tr>
<td>-1.0 to -1.49</td>
<td>Moderately dry</td>
</tr>
<tr>
<td>-1.5 to -1.99</td>
<td>Severely dry</td>
</tr>
<tr>
<td>≤-2</td>
<td>Extremely dry</td>
</tr>
</tbody>
</table>

Spatial Analysis

To determine the monsoon precipitation distribution over the region, the decadal monsoon data was plotted in ArcGIS as points. In finding the decadal monsoon data, the sum of the monthly monsoon data was computed, representing the yearly sum of monsoon precipitation. The yearly monsoon data was then grouped into ten-year periods, representing the decadal monsoon data. This was done for the 55 meteorological stations and ranging from 1900-2010.

As a result of missing data, the number of meteorological stations used for each year was different. Similarly, the remaining dataset was summed up and referred to as non-monsoon season dataset. Kriging interpolation method was selected from Spatial Analyst toolbox. Default parameters were maintained for the interpolation. The symbology of the output spatial map was edited by stretching the plot along a color ramp and using maximum-minimum to set an equal data range for all the maps. The data range was determined from the maximum and minimum precipitation values within the dataset. Thus, the range was set to 0-320, where lower values represents dryness or draught.
CHAPTER IV

RESULTS AND DISCUSSION

This chapter has two main parts. The first part presents the results of temporal analysis methods which include a descriptive analysis of precipitation time-series, trend analysis, coefficient of variability, Periodicities and Red-Noise confidence level and Standard precipitation index. The second part investigates the spatial distribution of decadal monsoon and non-monsoon precipitation.

Temporal Analysis

Precipitation Time series

Precipitation time-series were compared among the selected four climate divisions. Even though precipitation variations differ among the climate divisions, some precipitation events (e.g. 1905) occurred across three climate divisions (CD2, CD3, and CD6). The most noticeable pattern is the extensive peak in 1905 (Figure 2). This was a statewide flood event that led to an increase in flow volume within river basins in Arizona. Webb et al. (2004) showed that the period from 1905 to 1922, had the highest long-term annual flow volume in the 20th century, under the Colorado River Compact. The flood in 1905 was noted as one of the most destructive floods in the history of Arizona (Murphy 1904). The most remarkable in Arizona’s history were those in the Gila River basin (CD4). From a previous research, a series of seven floods, remarkable for the total volume of flow occurred from January 15 to April 30. In November there was another flood, which was remarkable for its magnitude, being the largest on record on
Salt River (Murphy 1904). This flood event created the very high peak depicted on the chart. The mean precipitation amount for CD7 was very low as compared to the mean values of the other climate divisions. This is because CD7 is constantly in normal drought with occasional high precipitations events. The precipitation timeseries in Figure 2 shows that, though each climate division is different, there are some precipitation events that are statewide because they appear in the timeseries of all the climate divisions. For example, 1905, 1940 and 1970 precipitation were experienced statewide but with varying intensities.
Figure 2: Precipitation time series of CD2, CD3, CD6 and CD7 for 1900-2010: (CD-Climate Division)
Trends in the Precipitation Data

Precipitation is seen to have some similarities across the climate divisions, but the next question was to inquire whether there was any statistically significant trend within the precipitation series. Therefore, the dataset was divided into three parts of approximately 30-year interval and tested with the Mann-Kendall’s trend test and the slopes determined by the Sen’s slope test. This division was done based on a study by McPhee et al. (2004), that suggest 20–30-year periods characterized by relatively dry or wet conditions in Arizona.

Climate division 2 (CD2) had a statistically significant downward trend in 1900-1930 and 1962-2010 while 1931-1961 was statistically insignificant. Again, CD’s 3, 6 and 7 were statistically insignificant over the three periods (See Table 2 below). The overall trend (1900-2010) however showed a rather interesting pattern where CD2 and CD7 had a statistically significant downward trend suggesting a decline in precipitation amounts in the two climate divisions. Meko et al. (2007) highlighted in his research that annual precipitation over the southwestern United States has begun to decrease over the last decade as part of a multidecadal cycle but a study by Anderson et al. (2010) found that there does not appear to be any systematic summer time trends over the core (southwestern United States) monsoon region (centered in Arizona and western New Mexico). Although this study did not analyze the trend in summer precipitation, the trends from 1900-1930, 1931-1961 and 1962-2010 similarly were random. However, the trends for 1900-2010 within climate divisions 2 and 7 support the results from the study of Meko et al (2007).
Table 3: Results of the Mann-Kendall’s and Sen’s Slope Test for the Periods 1900-1930, 1931-1961, 1962-2010 and 1900-2010

<table>
<thead>
<tr>
<th>Climate Division</th>
<th>p- level</th>
<th>Sig. Level</th>
<th>Sen's slope</th>
</tr>
</thead>
<tbody>
<tr>
<td>1900-1930</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CD2</td>
<td>0.05</td>
<td>0.02</td>
<td>-3.76</td>
</tr>
<tr>
<td>CD3</td>
<td>0.05</td>
<td>0.06</td>
<td>6.65</td>
</tr>
<tr>
<td>CD6</td>
<td>0.05</td>
<td>0.29</td>
<td>-1.38</td>
</tr>
<tr>
<td>CD7</td>
<td>0.05</td>
<td>0.18</td>
<td>-0.15</td>
</tr>
<tr>
<td>1931-1961</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CD2</td>
<td>0.05</td>
<td>0.81</td>
<td>0.33</td>
</tr>
<tr>
<td>CD3</td>
<td>0.05</td>
<td>0.25</td>
<td>-3.15</td>
</tr>
<tr>
<td>CD6</td>
<td>0.05</td>
<td>0.44</td>
<td>-1.25</td>
</tr>
<tr>
<td>CD7</td>
<td>0.05</td>
<td>0.41</td>
<td>-0.13</td>
</tr>
<tr>
<td>1962-2010</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CD2</td>
<td>0.05</td>
<td>0.00</td>
<td>-1.70</td>
</tr>
<tr>
<td>CD3</td>
<td>0.05</td>
<td>0.06</td>
<td>-2.82</td>
</tr>
<tr>
<td>CD6</td>
<td>0.05</td>
<td>0.27</td>
<td>0.94</td>
</tr>
<tr>
<td>CD7</td>
<td>0.05</td>
<td>0.21</td>
<td>-0.06</td>
</tr>
<tr>
<td>1900-2010</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CD2</td>
<td>0.05</td>
<td>5.09E-07</td>
<td>-0.95</td>
</tr>
<tr>
<td>CD3</td>
<td>0.05</td>
<td>7.49E-02</td>
<td>-0.74</td>
</tr>
<tr>
<td>CD6</td>
<td>0.05</td>
<td>6.16E-01</td>
<td>0.11</td>
</tr>
<tr>
<td>CD7</td>
<td>0.05</td>
<td>1.36E-02</td>
<td>-0.04</td>
</tr>
</tbody>
</table>
Coefficient of Variation (CV)

To determine if the precipitation variability has changed over time, coefficient of variability was computed by finding the mean and standard deviation. Standard deviation and coefficient of variability are linear statistical measures of the temporal variation existing within the dataset. The standard deviation is calculated by obtaining the arithmetic mean of the dataset and measuring how much each value differs from the mean. The coefficient of variability was then computed by dividing the standard deviation by the mean. From Table 3 below, the coefficient of variability is high for both CD2 and CD7. The standard deviation for precipitation was greater than the mean precipitation in CD7, making the CV greater than a hundred percent. However, the mean precipitation was greatest in CD3, followed by CD6. This is likely as a result of their location being on the windward side of the central highlands.

Table 4: Coefficient of Variation (CV) for the Climate Division over the Period 1900-2010.
Multitaper Spectral Analysis and Red-Noise Confidence Level

Multitaper method was employed to estimate the periodicity within the precipitation time series. Red-Noise confidence level was computed to determine the statistically significant trends. The results showed an interesting pattern where three statistically significant signals or peaks were present in all four climate divisions. These statistically significant spectral power (peaks) occurred in the periods of 59 months (approx. 5 years), 30 months (approx. 2.5 years), and 20 months (approx. 1.5 years) across the climate divisions. There were other minor statistically significant peaks that were not consistent across the climate divisions. These minor peaks occur every 15, 12, and 10 months in CD2, while they occur every 14, 12, 11, and 10 months in CD3. In CD6, these minor peaks occur every 19, 15, 14 and 10 months, while in CD7, they occur every 16, 12 and 11 months. Although topography and other local factors are known to affect precipitation in Arizona, this research showed that the major precipitation periodicity peaks were statewide, and they were within the range of recurrence of large scale atmospheric controls that affect Arizona. Previous researchers have highlighted several atmospheric factors within the return periods indicated by the Multitaper spectral analysis. These include, North American monsoon which occurs yearly, combined El Niño-Southern Oscillation phenomenon (ENSO) which occurs between 2-10 years, and the Pacific Decadal Oscillation (PDO) which occurs every 10-30 years.
Figure 3: Multitaper spectral analysis, red-noise spectrum and 95% confidence level for CD2, CD3, CD6 and CD7. (1900-2010)
Standard Precipitation Index (SPI)

Drought events are very closely related to the temporal distribution of precipitation. To understand future droughts, it is necessary to first understand the historical pattern of drought. Using the monthly precipitation data. The temporal and spatial variation of drought was analyzed across the Arizona climate divisions. The analysis showed five major statewide droughts, through the early to mid-1900s 1940s, 1960s 1970s and 1990s to 2010. Figure 4 clearly shows these droughts as well as several other short duration droughts across the individual climate divisions. The chart in figure 4 below shows that there have been multiple extended periods of precipitation with occasional one to two years drought. Some of the longest drought events in CD2 occurred in 1932-1940, 1942-1951 and 1970-1977. Some drought years in CD3 occurred in 1901-1905, 1942-1957 and 2001-2010. CD6 also had some of its longest drought years in 1953-1964, 1969-1977 and 1995-2010. Finally, CD7 also had its longest drought years occurring in 1942-1965, 1968-1977 and 1994-2010.
Figure 4: Monthly timescale SPI for CD2, CD3, CD6, and CD7 over the period 1900-2010.
Figure 5 shows that normal drought is more frequent across the climate divisions and constitute more than 60% of the drought occurrences. One of the interesting patterns observed in Figure 5 is that 99% of the drought in Southeast (CD7) is Normal drought with only 1% Moderate drought at the 12 months’ timescale. Even though CD7 experiences longer drought periods, there are no extremes drought occurrences present. Normal drought is observed to persist longer than any other drought type in Arizona. It persists for longer durations, sometimes stretching over consecutive years and are rampant in Southeast (CD7). There are more normal drought occurrences in CD7 and CD2 than there are in CD3 and CD6. However, CD3 and CD6 has more occurrence of Extreme and Moderate droughts. Analysis of the dataset shows that Moderate, Severe and Extreme droughts are short lived. They do not persist up to a year.

Arizona’s drought is generally known to have long-term effects on farms and forests, including dwindling crops and devastating wildfires and infestation of tree-killing beetles (Negron et al. 2009).
Figure 5: Drought frequencies in the climate divisions using 12-months timescale (1900-2010)
Spatial Analysis

Monsoon Season (June-Sept)

In analyzing the spatial trends in precipitation, monthly monsoon dataset was aggregated annually into decadal monsoon dataset and their decadal sums plotted in ArcGIS. The spatial maps presented in Figure 6 below displays the spatial patterns in the decadal monsoon dataset. The number of gauge stations vary for each year because of missing data. However, in the maps below, the western part shows scarcity of precipitation while the eastern part shows signs of precipitation. This pattern is largely due to the fact that Arizona lies on the leeward side of the mountain ranges in California, example, the White Mountain and the Sierra Nevada which blocks oncoming winds from the Pacific Ocean. Thus, there is a general dryness in the western half of Arizona. On the other hand, there is some amount of precipitation on the south eastern part of the Colorado Plateau and the southeast (CD7) of Arizona.

Comparing the decadal monsoon maps with satellite image of Arizona, it was found that decadal monsoon precipitation events were seen in forest areas or reservation areas like the Navajo, Hopi, Fort Apache, San Carlos, Colorado national forest and around the mountains in the southeast Arizona, suggesting the effect of evapotranspiration and topography as a contributing cause of precipitation in the area. The central highlands did not however, show signs of monsoon precipitation as expected. However, the southern Rocky Mountains and the southern edge of the Colorado Plateau both are known to intercept and promote convectional precipitation from the summer monsoon (Whitlock and Bartlein 1993).
The monsoon system described here is known to be formed when the seasonally warm land surfaces in both the lowlands and the elevated areas, interacts with atmospheric moisture supplied by the nearby maritime sources like the Gulf of Mexico and California, and the Eastern Pacific Ocean (Mitchell 1976; Trewartha 1981; Carleton 1985; Adams and Comrie 1997). Higgins et al. (1997), concluded that the effect of the monsoon extends over much of the western US and northwestern New Mexico. This casts more understanding into the precipitation footprints in the southeast Colorado Plateau area (near northwestern New Mexico) and the southeastern Arizona area. On the contrary, other studies have also found that the monsoon is strongest in northwestern New Mexico, and that Arizona only receives the northernmost fringes of precipitation. Scott 1991 found that Climate variation within the south eastern (CD7) region of Arizona results from overall physiography and topographic relief as well as proximity to the moisture sources of the Gulf of Mexico, the Gulf of California, and the eastern Pacific Ocean. In addition to that, high temperatures, high rates of evapotranspiration, and rain shadow effects of mountain ranges all contribute to regional variability.

ENSO is also known to contribute to the summer precipitation, however, there have been contradictory results from research on the relationship between ENSO and total summer precipitation. Harrington et al. (1992) found that Arizona and New Mexico receive significantly higher monsoon precipitation in July during El Niño years than during La Niña years. Webb and Betancourt (1992) also suggested that El Niño reduces the number of monsoonal storms that affect Arizona. Another study shows that, La Niña events are associated with below-average monsoon rainfall in Arizona and New Mexico, but El Niño events results in normal rainfall (Higgins et al. 1999). Thus, at this point there is no clear relationship between ENSO and total summer precipitation in Arizona (Andrade and Sellers 1988; Adams and Comrie 1997).
Figure 6: Map of Decadal Monsoon Precipitation from 1900-2010: (Red-Low precipitation, Yellow-Medium precipitation and Green-High precipitation)
Non-Monsoon Season (Nov-May)

Winter precipitation in Arizona, comes during the passage of frontal storm systems moving west-to-east guided by the jet stream, typically located in north of Arizona, but occasionally traversing the state. As this moist air masses encounter the Mogollon Rim they are lifted and cooled, which then condenses into water vapor and fall as rain.

In this research, the Non-monsoon decadal dataset was plotted in ArcGIS as shown in Figure 7 below. It is seen that the western half and the south east are mostly dry during the winter season. Precipitation footprints are only seen around the central highland areas (especially, CD4 or Gila) and part of the Colorado Plateau (CD2). This is because, from November through March, storm systems (such as PDO, ENSO, PNA, Southwest trough) from the Pacific Ocean cross the state (Woodhouse 1997, Burnett 1994, Barry and Chorley 1998, Mantua et al. 1997, Zhang et al. 1997). These winter storms occur frequently in the higher mountains of the central and northern parts of the state. This creates precipitation in the central highland area. CD2 is mostly dry because it lies on the leeward side of the central highlands and in a region of generally high altitude with diminished precipitation due to the “rain shadow effect.” As moisture-laden air flows over topographic features such as the central highland ranges, the air is lifted and cooled, much of the moisture is precipitated over the inland mountains. This underscores the influence of the Central Highlands on the Non-monsoon (Winter) precipitation. Whitlock & Bartlein (1993) highlighted that, at the regional scale, orography influences the amount of precipitation received. For example, the Sierra Nevada intercept winter rainfall.
Figure 7: Map of Decadal Non-Monsoon Precipitation from 1900-2010: (Red-Low precipitation, Yellow-Medium precipitation and Green-High precipitation)
CHAPTER V

CONCLUSION

While previous studies were conducted at larger scales like the southwest of USA, this research was conducted on a much smaller scale in order to identify the patterns existing within the climate divisions of Arizona. The result of this research was successful in displaying the trends existing in Arizona precipitation.

It was found that precipitation in Arizona is not uniform, the western half is mostly dry while the eastern half is wet. This is because the western half lies within the leeward side of the mountain ranges in California shielding the rain bearing winds from reaching Arizona. However, some of the winds from the Ocean reaching Arizona have tremendous effect on Arizona’s precipitation. This is realized because the precipitation peaks fall within the return periods displayed by the Multitaper spectral analysis. Precipitation seem to decrease within two climate divisions (CD2 and CD7) while the pattern is random within CD3 and CD6. It was found that summer precipitation affects the southeastern (CD7) Arizona and the south east of the Colorado Plateau while winter precipitation affects the Central highland areas. CD2 and CD7 have the most occurrence of “Normal drought” while “Severe and Extreme” droughts were prevalent in CD3 and CD6. Comparison of satellite image and the spatial maps (Figures 6 and 7) display the presence of precipitation in the forested areas thus it is recommended that further studies be conducted to verify the effect of vegetation on Arizona precipitation.

Finally, missing meteorological data was a limitation to this research. Lack of data and incomplete dataset caused a lot of areas not to be studied and the number of gauge stations in the spatial maps were not uniform making yearly comparison difficult. However, it is hoped that the
patterns displayed in this study will create more understanding in the trends in Arizona precipitation and drought as well as lay the foundation for further work in investigating precipitation in Arizona.
APPENDIX A

Precipitation Timeseries for Climate Divisions 2,3,6 and 7: 1900-2010
APPENDIX B

Multitaper Spectral Analysis and Red-Noise Mean Spectrum Plots with 95% Confidence Level for Climate Division 2,3,6 and 7 over the Period 1900-2010
APPENDIX C

Standard Precipitation Index (SPI) Charts for Climate Divisions 2, 3, 6 and 7 over the Period 1900-2010
APPENDIX D

Spatial Maps of Decadal Precipitation of Arizona for the Monsoon Season (June-September):
1900-2010
Monsoon decadal precipitation for 1900-1909
Monsoon decadal precipitation for 1910-1919
Monsoon decadal precipitation for 1920-1929
Monsoon decadal precipitation for 1930-1939
Monsoon decadal precipitation for 1940-1949
Monsoon decadal precipitation for 1950-1959
Monsoon decadal precipitation for 1960-1969
Monsoon decadal precipitation for 1970-1979
Monsoon decadal precipitation for 1980-1989
Monsoon decadal precipitation for 1990-1999
Monsoon decadal precipitation for 2000-2009
APPENDIX E

Spatial Maps of Decadal Precipitation of Arizona for the Non-Monsoon Season (October-May): 1900-2010
Non-monsoon decadal precipitation for 1900-1909
Non-monsoon decadal precipitation for 1910-1919
Non-monsoon decadal precipitation for 1920-1929
Non-monsoon decadal precipitation for 1930-1939
Non-monsoon decadal precipitation for 1940-1949
Non-monsoon decadal precipitation for 1950-1959
Non-monsoon decadal precipitation for 1960-1969
Non-monsoon decadal precipitation for 1970-1979
Non-monsoon decadal precipitation for 1980-1989
Non-monsoon decadal precipitation for 1990-1999
Non-monsoon decadal precipitation for 2000-2009
REFERENCES


