Regional aspects of prolonged meteorological droughts over Mexico

Matías Méndez[1] and Víctor Magaña[1]

1: Centro de Ciencias de la Atmósfera, Universidad Nacional Autónoma de México, Mexico City, Mexico.

Revised July 9th, 2009

Corresponding Author Address: Matías Méndez, Centro de Ciencias de la Atmósfera, Universidad Nacional Autónoma de México, Circuito Exterior s/n, Mexico City, Mexico, 04510

email: matias@atmosfera.unam.mx

Abstract

Major prolonged droughts in Mexico during the 20th century are mainly related to anomalous dry summers, such as those observed in the 1930´s, the 1950´s or the 1990's. Droughts in northern Mexico frequently coincide with anomalously wet conditions over Mesoamerica (i.e, southern Mexico and Central America), and vice versa, displaying a dominant "see-saw" structure in persistent precipitation anomalies, mostly in relation to tropical sea surface temperature (SST) anomalies. Changes in the mean flow during the positive phase of the Atlantic Multidecadal Oscillation (AMO) and associated with weaker than normal moisture flux have resulted in some of the most severe droughts in North America. However, drought over northern Mexico may also be related to changes in transient activity in the Caribbean Sea. During the negative phase of the Pacific Decadal Oscillation (PDO), the Caribbean Low Level Jet (CLLJ) weakens and Easterly Wave (EW) activity increases, leading to more tropical convection over Mesoamerica and less moisture flux into northern Mexico. On the other hand, when EW activity is weak over the Intra Americas Seas (IAS) (i.e. the Gulf of Mexico and the Caribbean Sea) due to a stronger than normal CLLJ, precipitation increases over northern Mexico. Therefore, the interaction between easterly waves and the trade winds over the IAS appears to be crucial to explain the spatial patterns of droughts that have affected Mexico. In addition, low frequency modulators, such as AMO or PDO may serve to explain the spatial patterns of severe prolonged droughts in Mexico during the 19th century.

1. Introduction

Persistent or prolonged droughts (duration of more than one year) have had negative consequences on the socioeconomic life of Mexico (Endfield et al. 2004; Endfield and Fernandez-Tejedo 2006; Acuña et al. 2000; Acuña et al. 2002). For example, some studies suggest that the collapse of the Mayan empire is related to a major drought episode in the 8th century (Culbert 1973; Hodell et al. 1995; Hodell et al. 2007). The so-called Mega Drought in the 16th century described through the use of tree-ring data, led to famine and epidemics (Therrell et al. 2004). At the end of the 18th century, the cumulative effect of recurrent droughts (1780, 1782, 1784 and 1785) culminated with the so-called "Year of Hunger", between 1785 and 1786, with devastating consequences for society in the agrarian heartland of central Mexico (Endfield 2007). In the 20th century, prolonged droughts have also affected the socioeconomic life of Mexico, and have led to trans-boundary water conflicts with the US. Knowing the mechanisms that result in prolonged drought, therefore, has and continues to be of importance to the society.

Mexican climate ranges from the hot and dry conditions in the northwestern Sonoran desert, with an annual rainfall of less than 100 mm, to the wet tropical climate in the southern part, where annual rainfall may reach more than 3000 mm. During winter, subsidence from the direct Hadley cell associated with the Inter-Tropical Convergence Zone (ITCZ) in the eastern Pacific maintains stable dry conditions over most of Mexico (Magaña et al. 2003) and little rain. Several studies on persistent droughts are based on tree ring chronologies (Villanueva et al. 2007), but they mainly reflect winter or spring climatic conditions. Over most of Mexico, more than 60% of the annual precipitation occurs during the boreal summer, i.e. from May-June through September-October (Garcia 2003). Because

of this seasonality, an analysis of prolonged drought in Mexico should focus on Northern Hemisphere (NH) summer rains. During this season, trade winds and easterly waves (EW) produce moisture flux from the Americas warm pools into continental Mesoamerica (Mestas-Nuñez et al. 2002; Wu et al. 2009). In the northern part of Mexico subsidence persists most of the year. It is only when easterly waves or tropical cyclones force ascending motions that copious rains occurs in the northeastern states. Over northwestern Mexico, the North America monsoon (Higgins et al. 2006) is associated with numerous Mesoscale Convective Systems and severe rain events from July through September.

Large scale climatic phenomena, such as El Niño Southern Oscillation (ENSO), result in climate variability on inter-annual time scales at locations about the globe (Diaz and Markgraf 2000). In Mexico, El Niño (La Niña) conditions during summer lead to below (above) normal precipitation over most of the territory. More precipitation during La Niña summers is likely due to the northward shift of the eastern Pacific ITCZ, weaker trade winds (Cavazos and Hastenrath 1990) that favor EW activity (Salinas 2006) and more tropical cyclones (TC) in the Intra Americas Seas (IAS) (i.e. the Gulf of Mexico and the Caribbean Sea) (Gray 1993). On the other hand, El Niño summers tend to inhibit precipitation over most of central southern Mexico due to an equatorward shift of the ITCZ (Waliser and Gautier 1993), enhanced subsidence over most of this continental region (Diaz and Bradley 2004) and diminished easterly wave and TC activity over the IAS. Therefore, El Niño is not associated with increased rainfall in Mexico, as it is in the US.

Since water availability in northern Mexico is scarce, the impacts of droughts in this region are considered more severe than in the south and consequently, they have been more deeply analyzed from the socioeconomic perspective (e.g., García-Acosta 2003; Magaña and

Conde 2003). However, dry and wet episodes in central and southern Mexico are also part of the natural climate variability here. Regionally, prolonged dry periods in recent centuries have been analyzed based on documents (e.g. Endfield and Fernandez-Tejedo 2006) and meteorological records (e.g. Mendoza et al. 2005; Mendoza et al. 2006), and *proxy* climatic reconstructions, such as marine and lake core sediments (Metcalfe and Davies 2007), or tree ring data (e.g. Therrel et al. 2002; Villanueva et al. 2006; Villanueva et al. 2007). In this way, severe droughts in northern Mexico during the second half of the 16th century (1545-1600), and during 1752-1768, 1801-1813, 1859-1868, 1910's, 1930´s, 1950´s and 1990´s have been documented. During the 1940´s, 1970´s or mid-1980s, relatively dry conditions were experienced in central-southern Mexico, that contrasted with relatively wet conditions in the north, characterizing a see-saw pattern in precipitation anomalies. Such contrast in persistent climatic anomalies is a characteristic of the decadal climate variability over Mexico and Central America. Even, tree-ring reconstructions of precipitation in North America (Cook et al. 2004) show that a prolonged drought in northern Mexico, as the one in the 16th century, corresponds with wet conditions in the south (Fig. 1). On the other hand, a drought in southern Mexico in 1630's corresponds to wet conditions in the north.

Most explanations of the mechanisms that result in prolonged droughts over North America have been given in terms of persistent sea surface temperature (SST) anomalies, either in the tropical Pacific, in relation to the Pacific Decadal Oscillation (PDO) (Mantua et al. 1997), or in the subtropical North Atlantic ocean, in relation to the Atlantic Multidecadal Oscillation (AMO) (Enfield et al. 2001). This is the case of droughts during the 1930's and 1950's (Schubert et al. 2004a; Schubert et al. 2004b; Cook et al. 2007). Some studies suggest that historical droughts in southern Mexico coincide with the negative phase of the

5

AMO (Mendoza et al. 2007), but may also be affected by the positive phase of the PDO or El Niño conditions (Seager et al. 2009). The positive phase of the PDO is related to more frequent El Niño events and therefore, recurrent droughts are expected over central/southern Mexico. According to Englehart and Douglas (2001), interdecadal shifts in the PDO are associated with significant changes in teleconnectivity between rainfall in Mexico and large-scale climate modes. Brito et al. (2002) determined that in northwestern Mexico, fluctuations of winter rains are characterized by an in phase relationship with the PDO, with the positive phase associated with humid winters and the negative phase with relatively dry winters. During winter, climate across Mexico is influenced by the state of the tropical Pacific Ocean with minor influence from the Atlantic Ocean. During the positive phase of the AMO, the northwest of Mexico is drier and a reduction of extreme daily precipitation events is observed (Curtis 2008). However, there is not much understanding of how the AMO affects the central and southern parts of Mexico.

One of the mechanisms that may modulate precipitation over Mexico is subsidence from the Hadley cell. Its meridional position is modulated by SST anomalies over the eastern Pacific (Diaz and Bradley 2004). Consequently, the latitudinal position of subsidence and negative precipitation anomalies are influenced by ENSO and PDO. This mechanism does not directly incorporate the role of Atlantic SST into summer precipitation. Given that Atlantic SST anomalies are correlated with more or less precipitation, it is necessary to explore how they exert influence. The tropical Atlantic teleconnects with Mexico trough easterly wave activity or moisture fluxes, from the Caribbean Sea, as well as through other mechanisms.

There are a number of dynamical elements during the NH summer that should be considered to explain drought over Mexico in relation to persistent SST anomalies. The main objective of the present study is to examine the regional spatial structures of prolonged droughts in Mexico for the instrumental period (1903-2002) and to determine their relationship with the very low frequency variability from the Pacific and Atlantic oceans, represented as the PDO and AMO. This work also explores other regional aspects of drought, such as the meridional extent of the persistent negative precipitation anomalies, and the east-west contrasts in precipitation anomalies, over northern Mexico. The use of PDO and AMO to explain the 19th century is also discussed.

The study is divided in four sections. In section two, the data and methodologies used for the analysis are presented. Section three is an analysis of the relationships between PDO, AMO and persistent regional precipitation anomalies in Mexico. In section four, the main findings are discussed in terms of the dynamical mechanisms that may teleconnect SST anomalies with regional climate in Mexico. Summary and conclusions are given in section five.

2. Data and Methodology

This study is based on drought indices from precipitation records. Drought indices may also be derived from records of other meteorological variables such as soil moisture, temperature or hydrologic variables (stream flow, levels of water reservoirs, etc.). These reflect the impact of precipitation deficiency on the availability of water, and provide reliable information on the intensity, duration and spatial extension of drought (Keyantash and Dracup 2002; Heim 2002). However, it is the dynamical mechanisms that lead to

summer precipitation anomalies that are the main focus of this analysis. In this study, meteorological droughts are analyzed by means of the so-called Standardized Precipitation Index (SPI) (McKee et al. 1993) considered a highly valuable estimator of drought severity. The SPI quantifies the precipitation anomalies on multiple time scales (3, 6, 12, 24 months), depending on the process to be examined. For instance, soil moisture conditions respond to precipitation anomalies on relatively short scales of a few months, whereas groundwater, stream flow, and reservoir storage reflect longer-term precipitation anomalies (more than one year). The SPI has three main advantages: First, its simplicity, since it is based on precipitation records and requires only two parameters to calculate (the shape and scale for the Gamma Probability Distribution), compared with the numerous parameters necessary to calculate the Palmer Drought Severity Index (PDSI) (Palmer 1965). Second, its temporary versatility is useful for the analysis of the dynamics of drought on various time scales. Third, its standardized form allows one to characterize the frequency of extreme (wet or dry) conditions and to define various droughts intensities (Table 1). The SPI also has disadvantages. First, the assumption that a suitable theoretical probability distribution can be found to model the raw precipitation data prior to standardization is not always met. Second, the SPI does not allow identification of regions of extreme wet and dry conditions when it is obtained for a very long time period because droughts occur with the same frequency at all locations over a long time period. Third, at short time scales (about a few months) misleadingly large positive or negative SPI values could be generated by relatively small anomalies when the normal regime for the region analyzed is characterized by low seasonal precipitation. In spite of these issues, the SPI is useful index to characterize prolonged droughts.

The SPI requires building of a frequency distribution from precipitation data at every location in the domain for a given time period (usually 1, 3, 6, 12, 24 months). A Gamma probability density function (PDF) may be fitted to the empirical distribution of precipitation frequency for the selected time scale and the cumulative distribution of precipitation to be determined. The cumulative distribution is transformed trough an equi-probability transformation to a standard Normal distribution (mean of zero and variance of one) to obtain the SPI values. Positive SPI values indicate greater than median precipitation, and negative values indicate less than median precipitation. Because the SPI follows a Normal distribution, wet and dry climates can be represented and monitored using this index. A SPI value equal to zero means that the corresponding monthly precipitation represents 50 % of the cumulative Gamma distribution. SPI values ranging from − 1 to +1 represent a normal amount (68%) of probability and values out of this range represent relevant deviations from the normal regime. Qualitatively, SPI values ranging from -2 to -1 (13.6% of cumulative probability) are associated with moderately dry and very dry episodes. Values exceeding -2 (2.3% of cumulative probability) are representative of extremely dry episodes (Table 1).

Because of its temporary versatility, the SPI can be calculated and analyzed using a specified time scale. For example, SPI-1 reflects the short term conditions and its application can be related to, for example, soil moisture; SPI-3 provides information on short- and medium-term moisture conditions and seasonal estimates of precipitation; SPI-6 or SPI-9 characterizes medium-term trends in precipitation and is considered to be more sensitive to moisture conditions than the Palmer Index at these time scales. It may also be associated with anomalous stream flow and reservoir levels. Droughts usually take a season

or more to develop, so SPI-12 is used to reflect the long-term precipitation patterns, related to volumes of rivers or reservoir levels. For the present study, SPI-24 months will be used to characterize persistent drought, capturing low frequency climate variability associated with stream flow, reservoirs levels, and groundwater levels at the longer time scales. The use of SPI-24 reduces the signals of year to year climate variability.

Analyses of precipitation or SPI for the 20th century are fairly reliable since the Mexican Weather Service has digitized historical records of daily precipitation for more than 3000 stations. The number of reports for the early part of the 20th century is relatively meager with 20 to 30 meteorological weather stations over Mexico. The number of stations increased to hundreds and even thousands after the 1930´s and 1950's, respectively. The precipitation reports have been subjected to quality control, by including a duplicate station check and spatial consistency (or "buddy") check. In addition, the coherence in monthly precipitation among neighboring stations (within a radius of 20 km) was used as a data quality criterion. Reports of anomalously large monthly precipitation (above the 95th percentile) or weak monthly precipitation (below the 5th percentile) in a station should coincide with the occurrence of positive or negative anomalies in precipitation in at least one of the neighboring stations. Using climatological stations from Mexico, the United States, Central America and the Caribbean, a gridded monthly precipitation dataset was developed using an objective analysis scheme of the Cressman type (Cressman 1959). This scheme is an interpolation method that corrects a first guess precipitation field by a linear combination of residuals (corrections) between calculated and observed values. In the current analysis, the scheme begins with a first guess of gridded monthly precipitation data obtained from the Climate Research Unit (CRU) of the University of East Anglia (Mitchell

and Jones 2005) for the 1901-2002 period. The background value at each grid point is then successively adjusted on the basis of nearby observations within the radius of influence (initial value is about 100 km) in a series of successive scans through the data, reducing this radius in order to increase precision. The monthly precipitation data set covers the period 1901-2002 in a 0.5° x 0.5° mesh, and is available at the International Research Institute for Climate and Society (IRI) (http://ingrid.ldeo.columbia.edu). This dataset has been used in other studies of drought for Mexico and North America (Mendez and Magaña 2007; Seager et al. 2009).

In this study, the PDO is defined as the first Empirical Orthogonal Function (EOF) of monthly SST anomalies in the North Pacific Ocean (20°N – 65°N and 120°E –100°W), (Mantua et al. 1997). Its time evolution or Principal Components corresponds to the so-called PDO index and it has been used to characterize the decadal variability of the Pacific Ocean. When the PDO index is positive, the north central Pacific Ocean tend to be cooler, and the west coast of North America tends to be warmer than normal. The opposite is true when the PDO index is negative.

The AMO has been characterized following Enfield et al. (2001) by considering the first EOF of SST anomalies in the North Atlantic region (between 20°N – 65°N and 100°W – 0°E). Its spatial pattern is characterized by SST anomalies in the North Atlantic basin (see Fig. 1 from Enfield et al. 2001). The amplitude of AMO exhibits a decadal timescale that modulates interannual variations. Time series of AMO and PDO for the 20th century (1903-2002) are available at the NOAA Climate Diagnostics Center (CDC). The analysis of dry periods during 19th century in Mexico were determined by means of *proxy* climatic reconstructions (Biondi et al. 2001) based on tree-ring chronologies of Southern and Baja

California and these show decadal-scale variability back to 1661 AD. Additionally, a tree-ring based reconstructions of the AMO index (Gray et al. 2004) were used.

In the present study, the dynamical processes that result in precipitation anomalies in Mexico for the second half of the 20th century are examined through the use of tropospheric wind data from the NCEP/NCAR Reanalysis (Kalnay et al. 1996) for the period 1948-2002. Specifically, EW activity is analyzed in the IAS region through the estimate of high frequency (3-9 days) variance of the meridional wind at 700 hPa from June through September.

3. Persistent droughts in Mexico during the 20th century

An analysis of annual precipitation anomalies, for the last century (1903-2002) shows that frequently, dry (wet) periods in northern Mexico (e.g. Chihuahua) correspond to wet (dry) periods in southern Mexico (e.g. Chiapas) (Fig. 2). This out phase relationship is also observed in previous prolonged drought periods (Fig. 1), where northern Mexico precipitation anomalies are out of phase with those in Central America.

The 20th century provides clear examples of intense droughts in northern Mexico during the 1930´s, 1950´s, and the late 1990´s (Fig. 3). There was also an intense and prolonged drought in the late 1910´s (not shown) that extended over most of Mexico and Texas, but only a few stations recorded data at that time, making it impossible to reliably document its spatial structure across North America. The signal of such intense drought may be observed in the precipitation anomaly in Chihuahua (Fig. 2) The Dust Bowl drought in the 1930´s (1934-1939) has probably been the most extensively documented dry event due to the

12

severe impacts it had across North America (Schubert et al. 2004a; Cook et al. 2008). During this period, several more precipitation reports are available for Mexico and the SPI-24 spatial pattern is more reliable. It is observed that during the summer months, the Dust Bowl drought affected parts of northwestern Mexico, near the US border (Fig. 3a). During the same period, significant positive precipitation anomalies (SPI-24 > 1.5) occurred over most of Mesoamerica and parts of the Caribbean, showing the characteristic north-south "see-saw" pattern of drought in this region.

The 1950´s drought (1953-1957) corresponds to the most severe event of this type in the recent history of the US (Seager et al. 2005) and Mexico. A broad area of precipitation deficit (SPI-24 < -2), extending from the US Midwest through northern Mexico, including the Baja peninsula, is observed (Fig. 3b). During this period a positive precipitation anomaly (SPI-24 > 1) is observed over southern Mexico, the Greater Antilles and Central America, except over a small part of Honduras. This drought pattern extends well across a large portion of North America, including parts of southeastern US. There seems to be a well defined dry-wet transition region around 20° N.

The most recent persistent drought in North America occurred during the second half of the 1990´s (1996-2002), extending over most of northern Mexico and parts of Texas, affecting some of the most important river basins shared by Mexico and the US (Fig. 3c). The SPI-24 negative anomaly (SPI-24 < -1) affected only some parts of the southern US but most of northern Mexico. As in previous cases, during this episode, positive SPI-24 anomalies are observed in the Caribbean and Mesoamerica regions, except over parts of the Yucatan peninsula and Guatemala. The impact of this drought episode was concentrated mainly

13

between the northern states of Mexico, and parts of Texas and Arizona, where water transfers from the Conchos River from Mexico are expected in the US. The transition from anomalously wet to dry occurs around 20° N.

As part of climate variability a dry episode is followed by a wet period (SPI-24 > 1). The relatively wet episodes in northern Mexico (Fig. 4), contrast with negative precipitation anomalies over Mesoamerica and the Caribbean region as during the 1940´s (1941-1943) (Fig. 4a), the 1970´s (1972-1979) (Fig. 4b) and the 1980´s (1985-1988) (Fig. 4c), reflecting the characteristic "see-saw" pattern associated with very low frequency (decadal) precipitation variability in North America. At times, these episodes also show a negative precipitation anomaly in the eastern part of the US (1940´s and 1980´s), exhibiting a zonal contrast in the SPI-24 pattern. As in the previous case of droughts in northern Mexico, the transition zone between wet and dry is roughly around 20° N.

The north-south dipole of low frequency variability of summer precipitation is more clearly visible in the first two Empirical Orthogonal Functions (EOFs) of the seasonal (June-September) SPI-24 for North America, Mesoamerica and the Caribbean (8.25°N-45.25°N and 66.25°W-125.75°W). The leading mode (first EOF) of SPI-24 has opposite signs loading patterns over US-northern Mexico and Mesoamerica-Caribbean regions, displaying a well defined "see-saw" spatial structure (Fig. 5a), with a transition zone between negative and positive anomalies around 20° N. This mode explains approximately 15% of the total variance. The spatial pattern of the second EOF (Fig. 5b) corresponds to a contrasting SPI-24 structure between northern Mexico-southern Mexico and the eastern part of the US, as well as between the Caribbean and western Mesoamerica. This mode explains around 8%

of the total variance, and provides a contrasting zonal structure that may be considered a quadripole. Characteristic spatial structures of drought, either in the north or south, vary from one episode to another. The broad spatial structure of drought may be reconstructed by a linear combination of EOF1 and EOF2 which shows the relative importance of each mode. The amplitude of EOF1 or EOF2 serves to characterize the magnitude of droughts at the regional level.

By looking at the Principal Components (PCs) for the period 1903 – 2002 (Fig. 6a and 6b), it is possible to reconstruct the dry or wet regional signal in SPI-24, in either northern or southern Mexico. The 1930´s drought may be examined by looking at EOF1 in its positive phase, while EOF2 is in a transition from its negative phase to a slightly positive phase. Such condition corresponds to a positive precipitation anomaly along the US-Mexico border that limits the extent of drought in the central and northeastern US. A closer look at the 1930's drought evolution indicates that EOF2 is negative but going to zero from 1933 through 1934. During this period, the SPI-24 anomaly extends across much of the upper half of the US. After 1935, the drought concentrates in the central US, extending down to Mexico. This corresponds to the period when the PC for EOF2 is around zero. The average pattern for SPI-24 mostly reflects the effect of EOF1. During this period, the Mesoamerica and Caribbean regions show a positive precipitation anomaly.

The 1950´s severe drought pattern in Mexico corresponds to the combined effect of positive EOF1 and positive EOF2, whose signals maximize during the 1953-1957 period. This results in an intense negative precipitation anomaly that extends all across the southern and US Midwest, as well as northern Mexico. The positive sign of EOF2 leads to positive

SPI-24 anomalies over the northeastern US and this tends to weaken the intensity of the 1950´s drought over this region. The Mesoamerican region shows positive SPI-24 pattern that comes from a large loading of PC1.

During the 1990´s drought, EOF2 is in its negative phase, which induces a positive drought anomaly in SPI-24 in northern Mexico. PC1 is in a transition from negative to positive phase along the US-Mexico border by the end of the 20th century, that reinforces the intensity of drought. This may be observed in the large negative precipitation anomaly in Chihuahua around the year 2000.

Under a similar reasoning, the wet periods for northern Mexico may be examined. The 1940´s period corresponds to an intense negative phase of EOF2 and a relatively weak EOF1. Therefore, the spatial pattern of drought in this period resembles EOF2. The SPI-24 pattern during the 1970´s shows a large negative PC1, and a large positive PC2, that combine to result in a positive SPI-24 over the eastern US and northeastern Mexico. Over Mesoamerica, PC1 and PC2 combine to result in large negative precipitation anomalies. The 1980´s SPI-24 pattern is a combination of a large negative PC1 and a large negative PC2 that clearly shows up as a negative precipitation anomaly over the eastern US, Central America and the Caribbean.

The combination of the first two EOFs of SPI-24 captures most of the regional characteristics of persistent droughts over the US, Mexico, Central America and the Caribbean. EOFs for SPI-24 are related to the low frequency modes of the Atlantic and Pacific Ocean. For instance, PC1 of SPI-24 is positively correlated with the AMO, mainly after the 1920´s, when more reliable SPI-24 data are available (Fig. 6a). On the other hand,

the PC2 of SPI-24 is negatively correlated with the PDO (Fig. 6b). This characteristic

spatial structure in SPI-24, a result of persistent SST anomalies, may serve to explore the

forcing mechanisms that teleconnect SST anomalies with regional precipitation over North

America, Central America and the Caribbean. The dynamical elements that teleconnect

such persistent SST anomalies, such as those during AMO, have been mainly related to

changes in quasi-stationary circulation anomalies (e.g. Hadley cell, low-level jet in the Gulf

of Mexico). To examine precipitation anomalies over Mexico, it is necessary to also

consider easterly waves activity over IAS and their interaction with the Caribbean Low

Level Jets (CLLJ) and even tropical cyclones. Since information on winds necessary to

examine EW activity is only available from the 1948, the subsequent analysis will be

developed for droughts after this year.

4. Mechanisms that produce prolonged droughts in Mexico

Several studies of the physical mechanisms that result in prolonged droughts relate

persistent climatic anomalies to anomalous SST. For the analysis of drought, it is necessary

to consider teleconnections originating in the Pacific and the Atlantic oceans North

America and producing persistent precipitation anomalies in Mexico. Quasi stationary

circulations have been proposed to relate distant regions (Horel and Wallace 1981) to local

effects. However, changes in mean circulations may be the result of transient activity

(Trenberth and Mo 1985), whose accumulated impact for a prolonged time period results in

climatic anomalies. The contrasting precipitation anomalies between northern and southern

Mexico could be related to a displacement of the meridional circulations such as the local

Hadley cell that modulates subsidence and moisture convergence or divergence (Seager et

al. 2009). This structure is generally associated with the mean position and extension of the

eastern Pacific ITCZ (Waliser and Gautier 1993). Descending motion reduces precipitation and causes drought over northern or central Mexico and the southern US (Mendez and Magaña 2007). However, there are regional elements, such as the zonal contrast in SPI-24 over northern Mexico that characterize droughts in North America and these are related to the conditions in the Atlantic Ocean in general, and the IAS in particular.

As shown by Schubert et al. (2009), the positive phase of AMO explains most of the 1930's drought, by an intensified high pressure system over the north Atlantic that leads to a weakening of the low level jet over the Gulf of Mexico and diminished moisture flux into the US Midwest. A warm IAS, during the positive phase of AMO, also favors tropical convective activity over the Caribbean that manifests as more easterly waves that travel across the Caribbean Sea, bringing much of the rain received during NH summer wet season (Ashby et al. 2005), and produce numerous storms over Mesoamerica. Furthermore, under favorable conditions, EWs may become tropical cyclones whose activity, modulated by AMO (Goldenberg et al. 2001), produces intense rains.

On the other hand, persistent El Niño (La Niña) condition, during the positive (negative) phase of the PDO, results in prolonged dry (wet) conditions over Mesoamerica. El Niño is related to a stronger CLLJ (Amador 1999) and diminished EW activity (Salinas 2006). EWs favor the occurrence of intense precipitation events over Mesoamerica. A decrease in EW activity results in negative precipitation anomalies over most of southern Mexico (Salinas 2006). Therefore, EW activity may also be affected by PDO. EW activity may be estimated by averaging the 700 hPa meridional wind 3-9 days variance in the central Caribbean sea. EW activity over the IAS is related to the intensity of the CLLJ. A strong

18

CLLJ tends to inhibit EWs development (Fig. 7). On the other hand, a relatively weak (~10 m/s) CLLJ tends to favor more EWs over the IAS and consequently, more precipitation over the Caribbean and Mesoamerica. This is consistent with changes in barotropic instability of the CLLJ (Molinari et al. 1997). Enhanced tropical convection over Central America may in turn reinforce subsidence over northern Mexico and consequently favor drier conditions (MacDonald et al. 2008).

There is also an EW guide around 20°N (Patricola et al. 2004), that may affect northern Mexico. The mean flow over the Gulf of Mexico may influence the passage of these systems at these latitudes. Ladwig and Stensrud (2009) suggest that tropical easterly waves north of 20°N produce precipitation over northwestern Mexico that extends to the west-central United States. The positive phase of AMO results in weaker trade winds over the IAS (Fig. 8). However, weaker trade winds over the Gulf of Mexico may reduce the number of northern tropical EWs that reach northern Mexico, leading to drier conditions as well. Contrary to what happens over Caribbean Sea, a weaker easterly flow does not enhance EW activity, presumably because the flow at these latitudes is not barotropically unstable.

The CLLJ is a key dynamic feature of the IAS climate (Wang 2007). As a barotropically unstable circulation, it may trigger and/or amplify EWs and even TCs that may produce above normal precipitation during the NH summer. The intensity of the CLLJ is related to ENSO and the PDO, in such a way that a warm eastern Pacific strengthens the CLLJ (Amador 1999) and reduces EW formation (Salinas 2006), as well as tropical convective activity over the Caribbean (south of 20°N) (Magaña et al. 2003). Diminished precipitation

over the Caribbean and Mesoamerica may result in a weaker local Hadley cell and less stable conditions over northern Mexico. A linear correlation between the CLLJ intensity and SPI-24 (Fig. 9) shows that an intense CLLJ results in negative precipitation anomalies over Mesoamerica and positive anomalies over northern Mexico and south-central US, with a spatial structure that resembles EOF2. Therefore, intensity of the CLLJ is responsible for the relationship between the PDO and PC2 of SPI-24.

The conditions in the tropical eastern Pacific and the tropical Atlantic together serve to explain the dynamics of the IAS circulations and the processes that result in dry and wet periods over the US, Mexico, Central America and the Caribbean. Moisture flux into the US Midwest appears to be controlled to a large extent by the conditions in the Atlantic, as reflected by the first EOF of SPI-24 and AMO. On the other hand, the PDO relates to the EOF2 of SPI-24. If the PDO is in its positive phase and the Atlantic in its negative phase, as during the 1980´s, the conditions for a rainy northern Mexico exists, but with a dry Mesoamerica and Caribbean. Combinations of a positive or negative AMO and PDO may explain the spatial structure of very low (decadal) regional climate variability over Mexico, depending on the intensity of the anomaly in each ocean. As proposed by McCabe et al. (2004), the drought events in North America may be explained in terms of PDO and AMO. In the case of Mexico, this is summarized as (Table 2):

Drought variability during 19th century over Mexico

The relationships between AMO, PDO and regional prolonged droughts in Mexico may be used to examine droughts during the 19th century, i.e. with this period serving as an independent sample. Since direct measurements of precipitation do not exist for most of this period, the reconstruction of drought episodes in various regions of Mexico has been done based on historical documents. According to Contreras (2005), the most severe droughts in northern central Mexico occurred during 1808-1811, 1868, 1877, 1884-1885 and 1892-1896, with the latter being considered the most severe. On the other hand, the first two decades of the 20th century correspond to dry conditions over southern Mexico (Mendoza et al. 2007). Reconstructions of the AMO and the PDO show that during the early part of the 19th century, an intense negative phase of the AMO occurred in conjunction with frequent episodes of positive PDO. This combination of AMO and PDO should correspond to dry conditions in Mesoamerica, and most probably, the Caribbean. Most droughts for the second half of the 19th century may be associated with positive PDO periods and relatively weak AMO anomalies. For instance, the 1890´s corresponded to a major drought over most of Mexico under the positive phase of the PDO (Fig. 10). The spatial pattern of the late 1910´s drought in Mexico corresponds with large negative precipitation anomalies over central and southern Mexico, reflecting a positive PDO phase that combines with negative AMO conditions.

Given the relationship between the tropical oceans, low frequency variability and precipitation anomalies in Mexico, it is possible to extrapolate AMO and PDO for the next coming decades and have an estimate of the future conditions of persistent drought at the regional level. NASA's Earth Observatory News announced the PDO has entered a

21

negative phase, which, under persistent positive AMO conditions may result once again in a significant drought over northern Mexico.

5. Summary and conclusions

The severe negative impact of prolonged droughts in Mexico motivates us to examine the elements that modulate regional climate variability on decadal time scales. An analysis of persistent precipitation anomalies and prolonged droughts over Mexico during the last century shows a dominant and contrasting spatial pattern between north and south. An EOF analysis of SPI-24 showed that EOF1 corresponds to this dominant "see-saw" structure, while the SPI-24 EOF2 shows there are contrasting zonal structures in precipitation anomalies like the ones between Midwestern and eastern US or between northern Mesoamerica and the Yucatan peninsula. This spatial characterization of persistent negative SPI-24 values provides insight into the regional dynamics of drought. EOF1 and EOF2 for SPI-24 are modulated by very low frequency variability in the tropical oceans. EOF1 of SPI-24 relates to AMO while EOF2 is associated with PDO. The combined effect of EOF1 and EOF2 of SPI-24 or PDO and AMO serve to reconstruct some of the regional patterns of droughts over Mexico, the southern US, Central America and the Caribbean. The phase and amplitude of EOF1 and EOF2 of SPI-24 or AMO and PDO for the 20th century, has proven to be useful in examining the characteristics of drought even during the 19th century.

Low (internannual) and very low (decadal) frequency climate variability over the tropics is largely determined by high frequency transient activity. In the case of Mexico, EWs interacting with the mean flow is a crucial to understanding the cause of years of intense

22

precipitation or drought at the regional level. From a preliminary analysis, it is shown that the intensity of the CLLJ and EW formation determine climate variability over the IAS and Mexico. Several additional analyses are necessary to determine threshold values for the CLLJ that inhibit or enhance EW activity and precipitation in Mexico. At present, we have explored the dynamics of droughts through the modulation of high frequency transients by stationary circulations and vice versa. Attempts to project future intense droughts in Mexico will depend on projections of PDO and AMO. The behavior of these modes of climate variability will serve to construct regional climate change scenarios in the coming decades.

References

Acuña-Soto R., L. Calderon Romero, J.H. Maguire, 2000: Large epidemics of haemorrhagic fevers in Mexico 1545-1815. *Am. J. Trop. Med. Hyg* **62(6):**733–739.

Acuña-Soto R, D. W. Stahle, M. K. Cleaveland, M. D. Therrell, 2002: Megadrought and megadeath in 16th century Mexico. *Emerg. Infect. Dis*. **8(4)**:360–362.

Amador J. A. 1998: A climatic feature of the tropical Americas: The trade wind easterly jet. *Top Meteor Oceanogr* **5(2)**:1–13.

Ashby, S. A., Taylor, M. A., and A. A. Chen, 2005: Statistical models for predicting Caribbean rainfall. *Theor. and Appl. Climatology*, **82**, 65-80.

Biondi, F., A. Gershunov, and D.R. Cayan, 2001: North Pacific decadal climate variability since AD 1661*, J. Climate*, **14(1)**.

Brito L., A. Leyva C., A.V. Douglas and D. Lluch B., 2002: Pacific Decadal Oscillation and the filled capacity of dams on the rivers of the Gulf of California continental watershed. *Atmósfera,* **15(2)**, 121-136.

Cavazos, T. and S. Hastenrath, 1990: Convection and rainfall over and their modulation by the Southern Oscillation. *Int. J. Climatol.*, **10**, 377-386.

Contreras Servín, Carlos 2005: Las sequías en México durante el siglo XIX. Investigaciones Geográficas, Boletín del Instituto de Geografía, UNAM ISSN 0188-4611, Núm. 56, 2005, pp. 118-133.

Cook, E. R., C. A. Woodhouse, C. M. Eakin, D. M. Meko, and D. W. Stahle, 2004: Long-Term Aridity Changes in the Western United States. *Science*, **306(5698)**, pp. 1015-1018, 5 November 2004.

Cook E. R., R. Seager, M. A. Cane and D. W. Stable, 2007: North American droughts: Recons-tructions, causes and consequences. *Earth Sci. Rev.* **81**, 93-134.

Cook, B.I., R.L. Miller and R. Seager, 2008: Dust and sea surface temperature forcing of the 1930's 'Dust Bowl' drought. *Geophys. Res. Lett.,* **35**.

Cressman G. P., 1959: An operational objective analysis system. Mon. Wea. Rev., 87, 367–374.

Culbert, T.P. 1973: The Classic Maya collapse. Albuquerque: University of New Mexico Press.

Curtis Scott, 2008: The Atlantic multidecadal oscillation and extreme daily precipitation over the US and Mexico during the hurricane season. *Climate Dyn.* **30(4)**: 343.

Diaz H. F. and V. Markgraf (Eds.), 2000: El Niño and the Southern Oscillation: Multi-scale Variability and Global and Regional Impacts, Cambridge University Press, 496 pp.

Diaz H. F. and R.S. Bradley (eds.), 2004: The Hadley Circulation: Present, Past and Future. Dordrecht: Kluwer Academic Publishers, 511 pp.

Englehart, P.J., and A.V. Douglas, 2002. Mexico's summer rainfall patterns: an analysis of regional modes and changes in their teleconnectivity. *Atmosfera* 15:147-164.

Endfield G. H., I. Fernández-Tejedo and S. L. O'Hara, 2004: Drought, and disputes, deluge and dearth: climatic variability and human response in colonial Oaxaca, Mexico. *J. Hist. Geogr.* **30**:249–276

Endfield, G. H. and I. Fernández-Tejedo, 2006 Decades of drought, years of hunger: archival investigations of multiple year droughts in late colonial Chihuahua. *Climatic Change* **75(4)**, 391-419.

Endfield, G.H., 2007: Climate and crisis in eighteenth century Mexico. *The Medieval History Journal* 10 (1 and 2): 99-125.

Enfield, D.B., A.M. Mestas-Nunez, and P.J. Trimble, 2001: The Atlantic Multidecadal Oscillation and its relationship to rainfall and river flows in the continental U.S., *Geophys. Res. Lett.*, **28**: 2077-2080.

García, Enriqueta, 2003: Distribución de la precipitación en la República Mexicana. Investigaciones Geográficas, Boletín del Instituto de Geografía. UNAM No. 50, 2003, pp. 67-76. (in spanish).

García-Acosta V., J. M. Pérez-Zevallos, A. Molina del Villar, 2003: *Desastres agrícolas en México. Catálogo histórico*. Tomo I: Épocas prehispánica y colonial (958–1822). Fondo de Cultura Económica and CIESAS, México, 506 pp.

Goldenberg S.B., C. W. Landsea, A. M. Mestas-Nunez, and W. M. Gray, 2001: The recent increase in Atlantic hurricane activity: Causes and implications: *Science*, **293**, 474–479.

Gray, C., 1993: Regional meteorology and Hurricanes. In: Climate Change in the Intra-Americas Sea. G. Maul, Ed., Eduard Arnold, 87-99.

Gray, S.T., L.J. Graumlich, J.L. Betancourt, and G.T. Pederson, 2004: A tree-ring based reconstruction of the Atlantic Multidecadal Oscillation since 1567 A.D. *Geophys. Res. Lett.*

Heim, R. R., Jr., 2002: A review of Twentieth-Century drought indices used in the United States. Bull. Amer. Meteor. Soc., 83, 1149-1165

Higgins, W., D. Ahijevych, J. Amador, A. Barros, E.H. Berbery, E. Caetano, R. Carbone, P. Ciesielski, R. Cifelli, M. Cortez-Vazquez, A. Douglas, M. Douglas, G. Emmanuel, C. Fairall, D. Gochis, D. Gutzler, T. Jackson, R. Johnson, C. King, T. Lang, M.I. Lee, D. Lettenmaier, R. Lobato, V. Magaña, J. Meiten, K. Mo, S. Nesbitt, F. Ocampo-Torres, E. Pytlak, P. Rogers, S. Rutledge, J. Schemm, S. Schubert, A. White, C. Williams,

A. Wood, R. Zamora, and C. Zhang, 2006: The NAME 2004 Field Campaign and Modeling Strategy. *Bull. Amer. Meteor. Soc.*, **87**, 79–94.

Hodell, D.A., J. H. Curtis, and M. Brenner, 1995: Possible role of climate in the collapse of the Classic Maya Civilization. *Nature* **375**: 391-394.

Hodell, DA, M. Brenner, and J. H. Curtis, 2007: Climate and cultural history of the Northeastern Yucatan Peninsula, Quintana Roo, *Mexico Climatic Change*, **83**, 215-240.

Horel, J. D. and J. M. Wallace, 1981: Planetary scale atmospheric phenomena associated with the Southern Oscillation. Mon. Wea. Rev., 109, 813-829.

Kalnay, E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, M. Iredell, S. Saha, G. White, J. Woollen, Y. Zhu, A. Leetmaa, R. Reynolds, M. Chelliah, W. Ebisuzaki, W. Higgins, J. Janowiak, K. Mo, C. Ropelewski, J. Wang, R. Jenne, and D. Joseph, 1996: The NCEP/NCAR 40-year reanalysis project. *Bull. Am. Met. Soc.*, **77**, 437-472.

Keyantash, J. and J. A. Dracup, 2002. The Quantification of Drought: An Analysis of Drought Indices. *Bull. Amer. Meteor. Soc.,* **83 (8)**, 1167-1180.

Ladwig, W.C., and D.J. Stensrud, 2009: Relationship between Tropical Easterly Waves and Precipitation during the North American Monsoon. *J. Climate*, **22**, 258–271.

MacDonald, Glen, D. W. Stahle, J. Villanueva-Diaz, N. Beer, S. J. Busby, J. Cerano-Paredes, J. E. Cole, E. R. Cook, G. Endfield, G. Gutierrez-Garcia, B. Hall, V. Magana, D. M. Meko, M. Méndez-Pérez, J. Sauchyn, E. Watson, A. Woodhouse, 2008: Climate Warming and 21st-Century Drought in Southwestern North America *EOS, Transactions American Geophysical Union*, **89(9)**.

Magaña Víctor, J.L. Vázquez, J.L. Pérez and J. B. Pérez, 2003: Impact of el Niño on precipitation in Mexico. *Geofis Int* **42**:313–330.

Magaña V. and C. Conde, 2003: Climate variability and climate change and their impacts on the freshwater resources in the border region: A case study for Sonora, México. In: *Climate and water-transboundary challenges in the Americas* (H. F. Díaz and B. S. Morehouse, Eds.) Kluwer Academic Publishers, Dordrecht, 373-393.

Mantua, N. J., S. R. Hare, Y. Zhang, J. M. Wallace, and R. C. Francis, 1997: A Pacific interdecadal climate oscillation with impacts on salmon production. *Bull. Amer. Meteor. Soc*., **78**, 1069-1079.

McCabe, G. J., M. A. Palecki, J. L. Betancourt, 2004: Pacific and Atlantic Ocean influences on multidecadal drought frequency in the United States. *Proceedings of the National Academy of Sciences*. **101**: 4136–4141.

McKee, T. B., N. J. Doesken and J. Kleist, 1993: The relationship of drought frequency and duration to time scales. *Proceedings of the Eighth Conference on Applied Climatology*. Boston, MA: American Meteorological Society. 179–184.

Mendez-Perez, J.M. and V. Magaña, 2007: Meteorological drought episodes over Mexico region. *American Geophysical Union. 2007 Joint Assembly. Fall Meeting*. Acapulco, Mexico. May 22-25, 2007.

Mendoza B., E. Jáuregui, R. Díaz-Sandoval, V. García-Acosta, V. Velasco, G. Cordero 2005: Historical droughts in central Mexico and their relation with El Niño. *J. Appl. Meteorol.* **44**:709–716.

Mendoza, B., V. Velasco, and E. Jáuregui, 2006: A Study of Historical Droughts in Southeastern Mexico. J. Climate, 19, 2916–2934.

Mendoza, B., V. García-Acosta, V. Velasco, E. Jáuregui and R. Díaz-Sandoval, 2007: Frequency and duration of historical droughts from the 16th to the 19th centuries in the Mexican Maya lands, Yucatan Peninsula. *Climatic Change,* **83**, 151-168

Mestas-Nuñez, A. M., C. Zhang, B. A. Albrecht, and D. B. Enfield, 2002: Warm Season Water Vapor Fluxes in the Intra-Americas Seas: *27th Annual Climate Diagnostics and Prediction Workshop*, Fairfax, VA, 21-25 October

Metcalfe S. and S. Davies, 2007: Deciphering recent climate change in central Mexican lake records. *Climatic Change*, **83**, 169-186.

Mitchell T. D. and P. D. Jones, 2005: An improved method of constructing a database of monthly climate observations and associated high resolution grids. Int. J. Climatol. 25, 693-712.

Molinari J., D. Knight, M. Dickinson, D. Vollaro, and S. Skubis, 1997: Potential vorticity, easterly waves, and eastern Pacific tropical cyclogenesis. *Mon. Wea. Rev*, **125**, 2699–2708

Palmer, W.C., 1965: Meteorological drought. Research Paper No. 45, U.S. Department of Commerce Weather Bureau, Washington, D.C.

Patricola, C. M., E. K. Vizy, and K. H. Cook, 2004: Easterly waves in the tropical Atlantic: Climatology and variability. Proceedings of the 1st International CLIVAR Science Conference, Baltimore, MD.

Salinas Prieto, José Antonio, 2006: Dinámica de ondas del este y su interacción con el flujo medio en el Caribe (in spanish). PhD dissertation (Doctorado en Ciencias de la Tierra)-UNAM, Centro de Ciencias de la Atmósfera. 109pp.

Schubert S. D., M. J. Suarez, P. J. Region, R. D. Koster and J. T. Bacmeiter, 2004a: Causes of long-term drought in the United States Great Plains. *J. Climate* **17**, 485-503.

Schubert S. D., M. J. Suarez, P. J. Region, R. D. Koster and J. T. Bacmeiter, 2004b: On the cause of the 1930s Dust Bowl. *Science* **303**, 1855-1859.

Schubert Siegfried, D. Gutzler, H. Wang, A. Dai, T. Delworth, C. Deser, K. Findell, R. Fu, W. Higgins, M. Hoerling, B. Kirtman, R. Koster, A. Kumar, D. Legler, D. Lettenmaier, B. Lyon, V. Magana, K. Mo, S. Nigam, P. Pegion, A. Phillips, R. Pulwarty, D. Rind, A. Ruiz-Barradas, J. Schemm, R. Seager, R. Stewart, M. Suarez, J. Syktus, M. Ting, C. Wang, S. Weaver, N. Zeng, 2009: A USCLIVAR Project to Assess and Compare the Responses of Global Climate Models to Drought-Related SST Forcing Patterns: Overview and Results. *J. Climate* (accepted).

Seager, R., Y. Kushnir, C. Herweijer, N. Naik, and J. Velez, 2005: Modeling of tropical forcing of persistent droughts and pluvials over western North America: 1856-2000. *J. Climate*, **18**, 4065-4088.

Seager, R., M. Ting, M. Davis, M.A. Cane, N. Naik, J. Nakamura, C. Li, E. Cook and D.W. Stahle, 2009: Mexican drought: An observational, modeling and tree ring study of variability and climate change, *Atmósfera*, **22, (1)**, 1-31.

Therrell, M. D., D. W. Stahle, M. K. Cleaveland, and J. Villanueva-Diaz, 2002: Warm season tree growth and precipitation over Mexico, J. Geophys. Res., 107(D14), 4205.

Therrell, M.D., D.W. Stahle, R. Acuña-Soto, 2004: Aztec drought and the curse of One Rabbit. *Bull. Amer. Meteor. Soc.* **85**: 1263-1272.

Trenberth, K.F., and K. Mo, 1985: Blocking in the Southern Hemisphere. *Mon. Wea. Rev.*, **113**, 3–21.

Villanueva-Diaz, J, D. W. Stahle, B. H. Luckman, J. Cerano-Paredes, M. D. Therrell, M. K. Cleaveland, and E. Cornejo-Oviedo, 2007: Winter-spring precipitation reconstructions from tree rings for northeast Mexico. *Climatic Change*, **83**, 117-131.

Waliser, D.E. and C. Gautier, 1993: A Satellite-Derived Climatology of the ITCZ. *J. Climate*, **6**, 2162-2174.

Wang, C. 2007: Variability of the Caribbean Low-level Jet and its relations to climate. *Climate Dyn.* **29**, 411–22

Wu, M. L. C., S. D. Schubert, M. J. Suarez, N. E. Huang, 2009: An Analysis of Moisture Fluxes into the Gulf of California. *J. Climate*, **22**, 2216–2239.
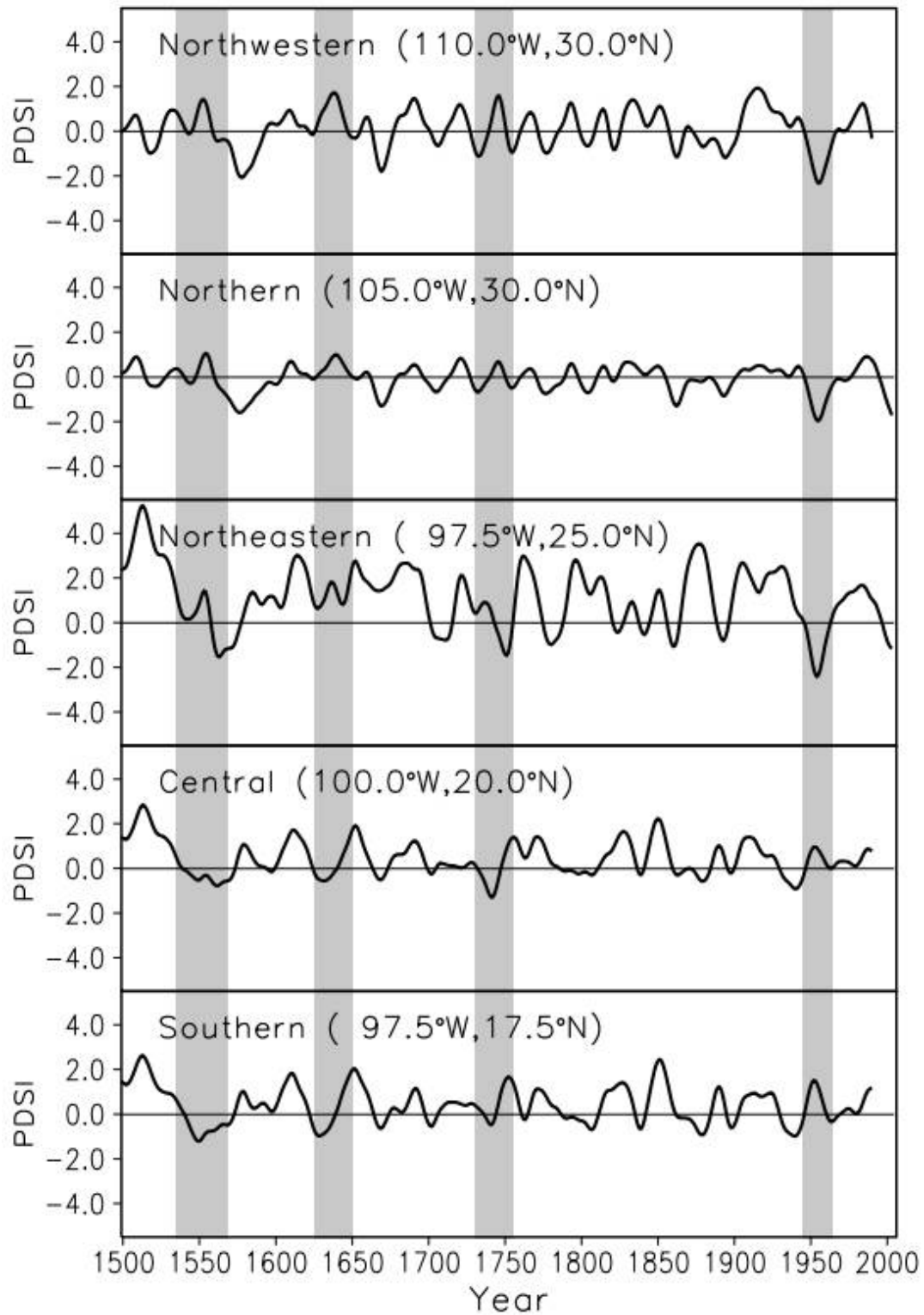
List of figures

Figure 1. Time series of reconstructed tree ring Palmer Drought Severity Index (PDSI) (20-yr low-pass) for various regions of Mexico (data from Cook et al 2004).
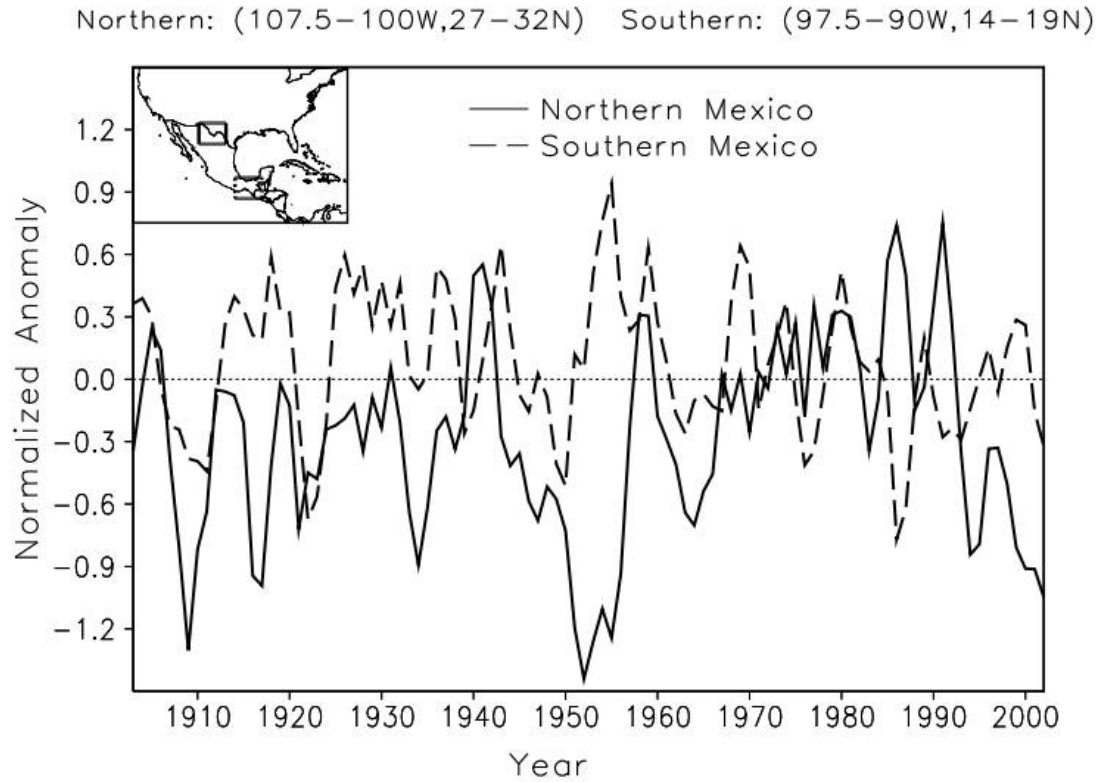
Figure 2. Normalized annual precipitation anomalies (zero line is the 1961–90 mean) in northern Mexico, Chihuahua (solid line) and southern Mexico, Chiapas (dashed line). The anomalies are derived from the developed gridded dataset (see text), using area-weighted averaging.
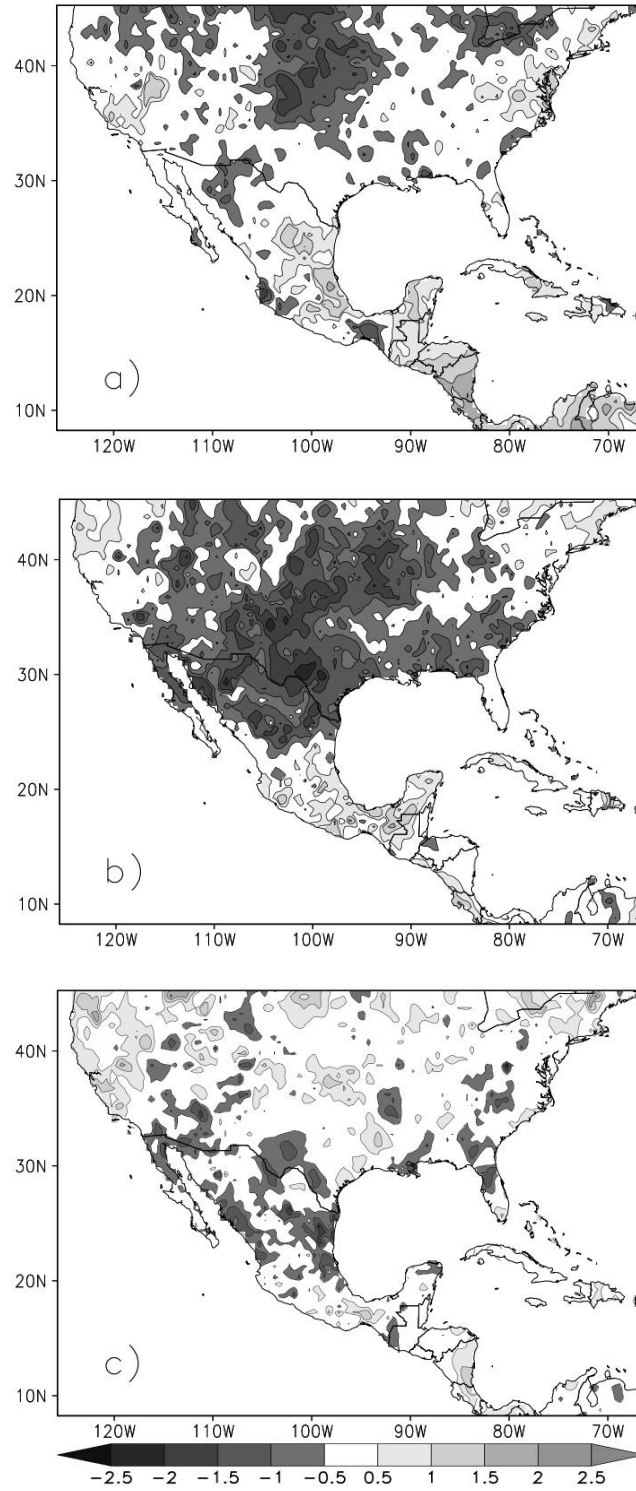
Figure 3. SPI-24 for NH summer during the a) 1930´s (1934-1939), b) 1950´s (1953-1957), and c) 1990´s (1996-2002).

Figure 4. As in Fig. 3 but for the a) 1940´s (1941-1944), b) 1970´s (1972-1979), and c) 1980´s (1985-1988).

Figure 5. a) EOF1 of SPI-24, and b) EOF2 of SPI-24 during NH summer. The number in parenthesis corresponds to the explained variance.

Figure 6. a) Principal Components (solid lines) of EOF1 of SPI-24 and AMO (gray bars), and b) EOF2 SPI-24 (solid lines) and PDO indices (gray bars).

Figure 7. Seasonal anomaly of variance of 3-9-day filtered meridional wind (in m^2s^{-2}) at 700hPa during the NH summer moths at 70°W, 17.5°N (solid line) and anomaly of the

mean seasonal magnitude of the Caribbean Low Level Jet (in ms^{-1}) (averaged over the region 12.5°-17.5°N, 80°-70°W).

Figure 8. Difference between the composites of 925hPa winds during dry (1953-1957, 1996-2002) minus wet (1972-79, 1985-1988) summers. Shading corresponds to the magnitude (ms^{-1}) of the difference.

Figure 9. One point correlation between CLLJ intensity (averaged over the region 12.5°-17.5°N, 80°-70°W) and SPI-24 during NH summer.

Figure 10. AMO (solid line) and PDO (dashed line) indices (reconstructed for the 1800-1900 period and observed for the 1901-2008 period).

Figure 1. Time series of reconstructed tree ring Palmer Drought Severity Index (PDSI) (20-yr low-pass) for various regions of Mexico (data from Cook et al 2004).

Figure 2. Normalized annual precipitation anomalies (zero line is the 1961–90 mean) in northern Mexico, Chihuahua (solid line) and southern Mexico, Chiapas (dashed line). The anomalies are derived from the developed gridded dataset (see text), using area-weighted averaging.

Figure 3. SPI-24 for NH summer during the a) 1930´s (1934-1939), b) 1950´s (1953-1957), and c) 1990´s (1996-2002).

Figure 4. As in Fig. 3 but for the a) 1940´s (1941-1944), b) 1970´s (1972-1979), and c) 1980´s (1985-1988).

Figure 5. a) EOF1 of SPI-24, and b) EOF2 of SPI-24 during NH summer. The number in parenthesis corresponds to the explained variance

Figure 6. a) Principal Components (solid lines) of EOF1 of SPI-24 and AMO (gray bars), and b) EOF2 SPI-24 (solid lines) and PDO indices (gray bars).

Figure 7. Seasonal anomaly of variance of 3-9-day filtered meridional wind (in m^2s^{-2}) at 700hPa during the NH summer moths at 70°W, 17.5°N (solid line) and anomaly of the mean seasonal magnitude of the Caribbean Low Level Jet (in ms^{-1}) (averaged over the region 12.5°-17.5°N, 80°-70°W).
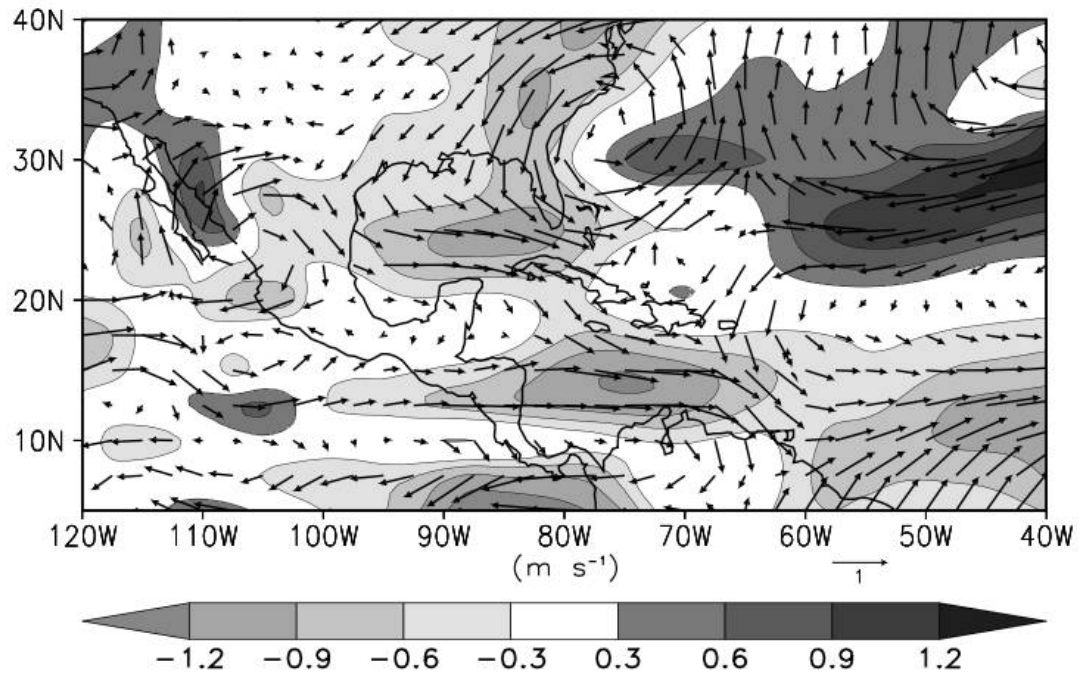
Figure 8. Difference between the composites of 925hPa winds during dry (1953-1957, 1996-2002) minus wet (1972-79, 1985-1988) summers. Shading corresponds to the magnitude (ms^{-1}) of the difference.
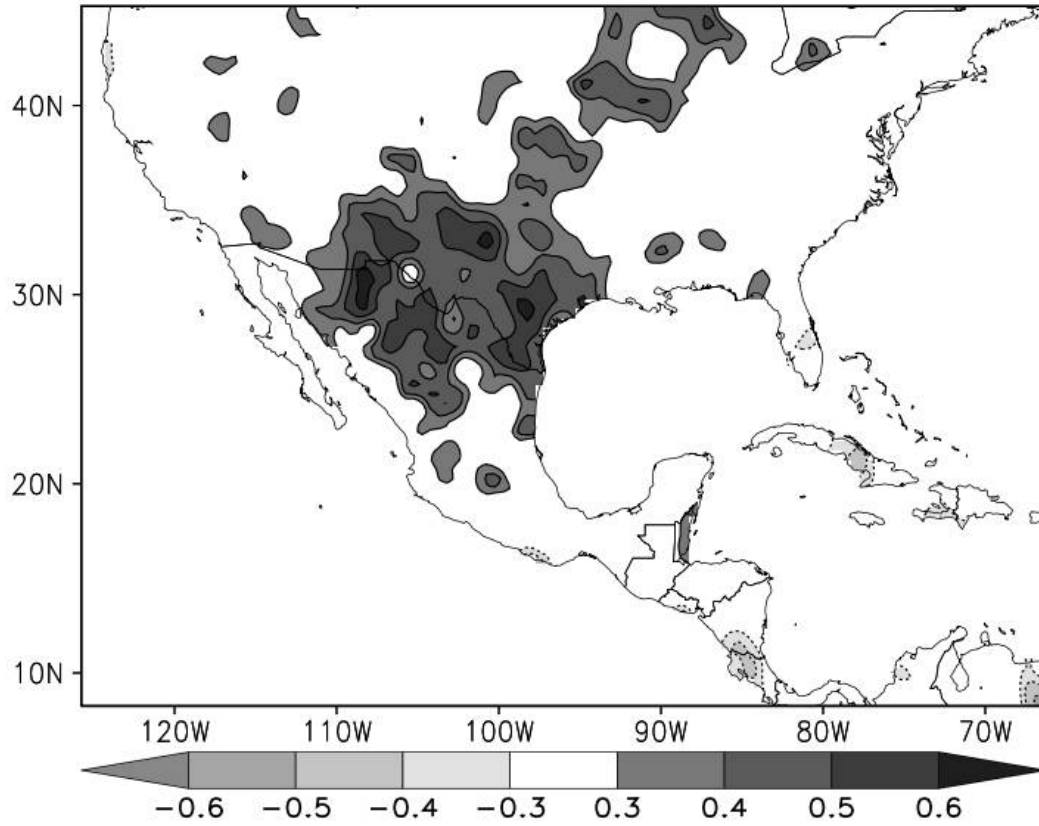
Figure 9. One point correlation between CLLJ intensity (averaged over the region 12.5°-17.5°N, 80°-70°W) and SPI-24 during NH summer.
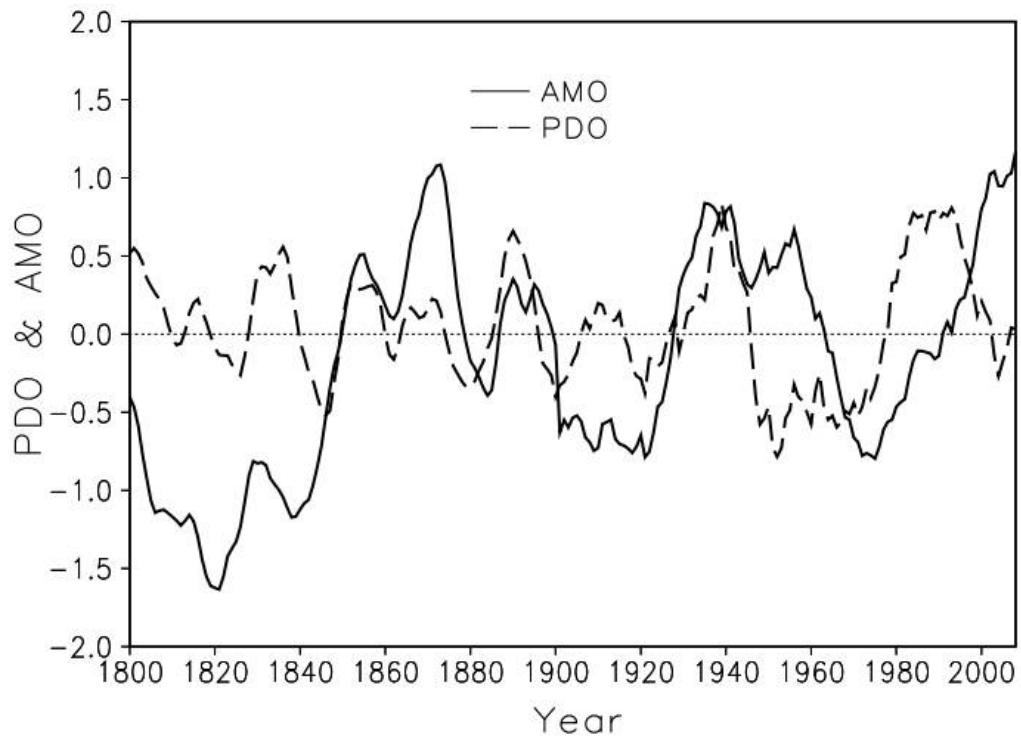
Figure 10. AMO (solid line) and PDO (dashed line) indices (reconstructed for the 1800-1900 period and observed for the 1901-2008 period).

Table 1. Classification of drought severity in terms of SPI and its probability

SPI	Category	Probability (%)
> 2.00	Extremely humid	2.3
1.5 to 1.99	Severely humid	4.4
1.0 to 1.49	Moderately humid	9.2
0.5 to 0.99	Humid	15.0
0.0 to 0.49	Normal to slightly humid	19.1
-0.49 to 0.0	Slightly dry to normal	19.1
-0.99 to −0.5	Dry	15.0
-1.49 to −1.0	Moderately dry	9.2
-1.99 to -1.5	Severely dry	4.4
< -2.0	Extremely dry	2.3

Table 2. Sign of the PDO and AMO indices that lead to regional droughts in Mexico

	Drought northern Mexico	Drought central southern Mexico
PDO	-	+
AMO	+	-