Regional Aspects of Prolonged Meteorological Droughts over Mexico and Central America

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ABSTRACT

Major prolonged droughts in Mexico during the twentieth century are mainly related to anomalous dry summers, such as those observed in the 1930s, the 1950s, or the 1990s. 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 “seesaw” structure in persistent precipitation anomalies, mostly in relation to tropical sea surface temperature (SST) anomalies. A warmer North Atlantic Ocean, expressed as a positive phase of Atlantic multidecadal oscillation (AMO), is related to the occurrence of major droughts in North America associated with weaker-than-normal moisture flux into northern Mexico. 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) because of 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 nineteenth century.

1. Introduction

Persistent or prolonged droughts (duration of more than one year) have had negative consequences on the socio-economic life of Mexico (Endfield et al. 2004; Endfield and Fernández-Tejedo 2006; Acuña-Soto et al. 2000, 2002). For example, some studies suggest that the collapse of the Mayan empire was related to a major drought episode in the eighth century (Culbert 1973; Hodell et al. 1995, 2007). The so-called mega-drought in the sixteenth century, described through the use of tree-ring data, led to famine and epidemics (Therrell et al. 2004). At the end of the eighteenth 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 twentieth century, prolonged droughts have also affected the socioeconomic life of Mexico and have led to transboundary water conflicts with the United States. Understanding the mechanisms that result in prolonged drought, therefore, has been 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 intertropical 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-Diaz 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, that is, from the May–June period through

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September–October (Garcı´a 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 produce moisture flux from the Americas warm pools into continental Mesoamerica (i.e., the geographical area that extends from central Mexico down through Central America) (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 (EW) or tropical cyclones (TC) force ascending motions that copious rains occur 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 interannual 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-Prieto 2006), and more TCs 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 because of an equatorward shift of the ITCZ (Waliser and Gautier 1993), enhanced subsidence over most of this continental region (Diaz and Bradley 2004), and diminished EW and TC activity over the IAS. Therefore, El Niño is not associated with increased rainfall in Mexico, as it is in the U.S. southwest.

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., Garcia-Acosta et al. 2003; Magaña and Conde 2003). However, dry and wet episodes are also part of the natural climate variability in both central and southern Mexico. Regionally, prolonged dry periods in recent centuries have been analyzed based on documents (e.g., Endfield and Fernández-Tejedo 2006), 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., Therrell et al. 2002; Villanueva-Diaz et al. 2007). In this way, severe droughts in northern Mexico during the second half of the sixteenth century (1545–1600) and during 1752–68, 1801–13, 1859–68, the 1910s, the 1930s, the 1950s, and the 1990s have been documented. During the 1940s, 1970s, or mid-1980s, relatively dry conditions were experienced in central-southern Mexico that contrasted with relatively wet conditions in the north, characterizing a seesaw 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 sixteenth century, corresponds with wet conditions in the south (Fig. 1). On the other hand, a drought in southern Mexico in the 1630s 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 1930s and
The study is divided in four sections. In section 2, the data and methodologies used for the analysis are presented. Section 3 is an analysis of the relationships between PDO, AMO, and persistent regional precipitation anomalies in Mexico. In section 4, 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 5.

2. Data and methodology

This study is based on drought indices derived from precipitation records. Drought indices may also be derived from records of other meteorological variables such as soil moisture, temperature, or hydrologic variables (streamflow, levels of water reservoirs, etc.). These indices reflect the impact of precipitation deficiency on the availability of water and provide reliable information on the intensity, duration, and spatial extent 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), which is considered a highly valuable estimator of drought severity. 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 time scales of a few months, whereas groundwater, streamflow, 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 temporal 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 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 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 through an equal 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. An 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 temporal 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 streamflow 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 streamflow, reservoir levels, and groundwater levels at the longer time scales. The use of SPI-24 reduces the signals of year-to-year climate variability.

<table>
<thead>
<tr>
<th>SPI</th>
<th>Category</th>
<th>Probability (%)</th>
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<tbody>
<tr>
<td>&gt;2.0</td>
<td>Extremely humid</td>
<td>2.3</td>
</tr>
<tr>
<td>1.5 to 1.99</td>
<td>Severely humid</td>
<td>4.4</td>
</tr>
<tr>
<td>1.0 to 1.49</td>
<td>Moderately humid</td>
<td>9.2</td>
</tr>
<tr>
<td>0.5 to 0.99</td>
<td>Humid</td>
<td>15.0</td>
</tr>
<tr>
<td>0.0 to 0.49</td>
<td>Normal to slightly humid</td>
<td>19.1</td>
</tr>
<tr>
<td>−0.49 to 0.0</td>
<td>Slightly dry to normal</td>
<td>19.1</td>
</tr>
<tr>
<td>−0.99 to −0.5</td>
<td>Dry</td>
<td>15.0</td>
</tr>
<tr>
<td>−1.49 to −1.0</td>
<td>Moderately dry</td>
<td>9.2</td>
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<tr>
<td>−1.99 to −1.5</td>
<td>Severely dry</td>
<td>4.4</td>
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<tr>
<td>&lt;−2.0</td>
<td>Extremely dry</td>
<td>2.3</td>
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North America (Mendez-Perez and Magana 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°–65°N and 120°E–100°W) (Mantua et al. 1997). Its time evolution or principal components correspond 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 tends 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° and 65°N and 100°W and 0°). 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 phase exhibits a decadal time scale that modulates interannual variations of climate in the United States and other regions of the world (Sutton and Hodson 2005). Time series of AMO and PDO for the twentieth century (1903–2002) are available at the National Oceanic and Atmospheric Administration (NOAA) Climate Diagnostics Center (CDC) Web site. The analysis of dry periods during nineteenth century in Mexico were determined by means of proxy climatic reconstructions of PDO (Biondi et al. 2001) based on tree-ring chronologies of Southern and Baja California, and these show decadal-scale variability back to A.D. 1661. Additionally, 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 twentieth century are examined through the use of tropospheric wind data from the National Centers for Environmental Prediction–National Center for Atmospheric Research (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 (Diedhiou et al. 1999) from June through September in the central Caribbean Sea (17.5°N and 70°W). Finally, the Caribbean low-level jet (CLLJ) intensity is defined by averaging the 925-hPa zonal wind anomalies multiplied by −1 over the region 12.5°–17.5°N and 80°–70°W (Wang 2007).

3. Persistent droughts in Mexico during the twentieth 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-of-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 twentieth century provides clear examples of intense droughts in northern Mexico during the 1930s, 1950s, and the late 1990s (Fig. 3). There was also an intense and prolonged drought in the late 1910s (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 1930s (1934–39) has probably been the most extensively documented dry event because of the 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 U.S. 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 seesaw pattern of drought in this region.

The 1950s drought (1953–57) corresponds to the most severe event of this type in the recent history of the United States (Seager et al. 2005) and Mexico (Seager et al. 2009). A broad area of precipitation deficit
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 United States.

The most recent persistent drought in North America occurred during the second half of the 1990s (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 United States (Fig. 3c). The negative precipitation anomaly (SPI-24 < -1) affected only some parts of the southern United States but most of northern Mexico. As in previous cases, during this episode, positive SPI-24 values 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 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 United States. There seems to be a well-defined dry–wet transition region 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 1940s (1941–43) (Fig. 4a), the 1970s (1972–79) (Fig. 4b), and the 1980s (1985–88) (Fig. 4c), reflecting the characteristic seesaw 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 United States (1940s and 1980s), 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 EOFs of the seasonal (June–September) SPI-24 for North America, Mesoamerica, and the Caribbean region (8.25°–45.25°N and 66.25°–125.75°W). The leading mode (first EOF) of SPI-24 has opposite signs loading patterns over the United States–northern Mexico and Mesoamerica–Caribbean regions, displaying a well-defined seesaw 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 United States and the eastern part of the United States, 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 quadrupole. Characteristic spatial structures of drought, either in the north or south, vary from one episode to another. The broad spatial structure of each individual observed 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 (Figs. 6a,b), it is possible to reconstruct the dry or wet regional signal in SPI-24, in either northern or southern Mexico. The 1930s 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 U.S.–Mexico border that limits the extent of drought in the central and northeastern United States. A closer look at the 1930s 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 United States. After 1935, the drought concentrates in the central United States, 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 1950s severe drought pattern in Mexico corresponds to the combined effect of positive EOF1 and positive EOF2, whose signals maximize during the 1953–57 period. This results in an intense negative precipitation anomaly that extends all across the southern and Midwest United States as well as northern Mexico. The positive sign of EOF2 leads to positive SPI-24 anomalies over the northeastern United States and this tends to weaken the intensity of the 1950s drought over this region. The Mesoamerican region shows positive SPI-24 pattern that comes from a large loading of PC1.

During the 1990s 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 U.S.–Mexico border by the end of the twentieth 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 1940s 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 1970s shows a large negative PC1 and a large positive PC2 that combine to result in a positive SPI-24 over the eastern United States and northeastern Mexico. Over Mesoamerica, PC1 and PC2 combine to result in large negative precipitation anomalies. The 1980s 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 United States, 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 United States, Mexico, Central America, and the Caribbean. EOFs for SPI-24 are related to the low-frequency modes of the Atlantic and Pacific Oceans. For instance, PC1 of SPI-24 is positively correlated with the AMO, mainly after the 1920s 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 also necessary to consider easterly wave activity over the IAS. Easterly waves interact with the Caribbean low-level jet and tropical cyclones—both of which have significant impacts on precipitation over Mexico. Since information on winds necessary to examine EW activity is only available from 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 prolonged drought in Mexico, the teleconnection between the Pacific and Atlantic Oceans should be considered. 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 United States (Mendez-Perez 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.

The positive phase of AMO explains most of the 1930s 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 U.S. Midwest (Schubert et al. 2009). 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 producing 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) conditions, 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 1998) and diminished EW activity (Salinas-Prieto 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-Prieto 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 (Diedhiou et al. 1999) during summer in the central Caribbean Sea (17.5°N and 70°W). EW activity over the IAS is related to the intensity of the CLLJ. A strong CLLJ tends to inhibit EW development (Fig. 7). On the other hand, a relatively weak (≤10 m s⁻¹) CLLJ tends to favor more EWs over the IAS and, consequently, more precipitation over the Caribbean and Mesoamerica. As shown in Fig. 7, during 1970s and 1980s drought in Central America, CLLJ was intense (positive CLLJ index) and this resulted in a reduction of EW activity (negative variance anomaly). In general, a strong (weak) CLLJ inhibits (allows) EW activity and this corresponds more (less) precipitation in southern Mexico and Central America. 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).

**Fig. 7.** Seasonal anomaly of variance of 3–9-day filtered meridional wind (m² s⁻²) at 700 hPa during the NH summer months at 17.5°N, 70°W (open circle) and anomaly of the mean seasonal magnitude of the Caribbean low-level jet (in m s⁻¹), averaged over the region 12.5°–17.5°N, 80°–70°W (closed circle).
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. An analysis of 925-hPa wind field difference between drought (1950s and 1990s) and wet periods (1970s and 1980s) in northern Mexico results in a reversed circulation in northeastern Mexico and the Gulf of Mexico (Fig. 8), less moisture flux in those regions, and a reduction of precipitation. In the Caribbean Sea, trade winds decelerate and this allows greater EW activity, resulting in more precipitation in southern Mexico. 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 1998) and reduces EW formation (Salinas-Prieto 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 CLLJ intensity and SPI-24 for summer months (Fig. 9) shows that a strong (weak) CLLJ corresponds to positive (negative) precipitation anomalies in the northern Mexico and negative (positive) anomalies in the southern. In general, a strong (weak) CLLJ inhibits (allows) EW activity and consequently less (more) precipitation in southern Mexico. This shows that an intense CLLJ results in negative precipitation anomalies over Mesoamerica and positive anomalies over northern Mexico and the south-central United States, 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. Resuming, a strong (weak) CLLJ corresponds to negative (positive) precipitation anomalies in southern Mexico associated to less (more) EW activity.

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 United States, Mexico, Central America, and the Caribbean. Moisture flux into the U.S. 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 is in its negative phase, as during the 1980s, the conditions for a rainy northern Mexico exist, but with a dry Mesoamerica and

![Fig. 8. The composites of 925-hPa winds during dry (1953–57, 1996–2002) minus wet (1972–79, 1985–88) summers. Shading corresponds to the magnitude (m s⁻¹) of the difference. The arrows correspond to 1 m s⁻¹.](image-url)
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 in Table 2.

**Drought variability during the nineteenth century over Mexico**

The relationships between AMO, PDO, and regional prolonged droughts in Mexico may be used to examine droughts during the nineteenth century, that is, 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-Servín (2005), the most severe droughts in northern central Mexico occurred during 1808–11, 1868, 1877, 1884–85, and 1892–96, with the latter being considered the most severe. On the other hand, the first two decades of the twentieth 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 nineteenth 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 nineteenth century may be associated with positive PDO periods and relatively weak AMO anomalies. For instance, the 1890s corresponded to a major drought over most of Mexico under the positive phase of the PDO (Fig. 10). The spatial pattern of the late 1910s 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 coming decades and have an estimate of the future conditions of persistent drought at the regional level. The National Aeronautics and Space Administration’s (NASA) Earth Observatory News announced the PDO has entered a 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

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

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<th>PDO</th>
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**FIG. 10. AMO (solid line) and PDO (dashed line) indices (reconstructed for the 1800–1900 period and observed for the 1901–2008 period).**
shows a dominant and contrasting spatial pattern between north and south. An EOF analysis of SPI-24 showed that EOF1 corresponds to this dominant seesaw structure, while the SPI-24 EOF2 shows there are contrasting zonal structures in precipitation anomalies like the ones between the midwestern and eastern United States 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 United States, Central America and the Caribbean. The phase and amplitude of EOF1 and EOF2 of SPI-24 or AMO and PDO for the twentieth century has proven to be useful in examining the characteristics of drought even during the nineteenth century.

Low- (interannual) 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 crucial to understanding the cause of years of intense 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.

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