A Multi-Scale Study of U.S. Drought
Risk and Predictability

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Abstract

This thesis proposes a multi-scale study of U.S. drought risk and predictability in order to better understand 1) how land, atmosphere, and oceans interact during the onset of drought, 2) how the processes involved in generating mechanisms of drought can be better identified by a multi-scale study, rather than a single dimensional study, and 3) how reliable known drought mechanisms are in explaining and predicting recent and future drought events.

In this thesis, there are seven chapters. Chapter 1 summarizes the objectives and goals of the thesis research. Chapter 2 investigates changes in the low flow regime over the eastern U.S. region due to climate change and variability. It also assesses the association between low flows and large-scale atmospheric circulations. In Chapter 3, the multi-scale driving mechanisms of droughts and floods over the southeastern U.S. are studied using a recently published regional reanalysis dataset. It establishes favorable conditions of the southeastern U.S. droughts and floods on land, and in the atmosphere and oceans. In Chapter 4, the influence of tropical cyclones (TCs) on the eastern U.S. droughts is assessed as a protective factor by comparing two land surface model simulations that are forced with or without TC-related precipitation. Chapter 5 investigates recent changes in drought risk influenced by the Pacific and Atlantic Oceans. Chapter 6 examines the current predictability of U.S. drought using a multi-model ensemble forecasting system. This study tests whether current climate forecasting models can reproduce the observed associations between global sea surface temperature and U.S. precipitation. Chapter 7 provides a brief summary of this study and discusses the next steps and future work of this study.

By crossing the spatial-temporal scales in studying drought, this thesis finds that there are previously hidden processes from a fixed scale (one dimensional) approach.
These hidden processes are found via a multi-scale study by investigating interactions of different scale processes on land, and in the atmosphere and oceans. This thesis also finds that some of the recent drought events are inconsistent with the known drought mechanisms found from previous studies (e.g. strong associations of global oceans with U.S. drought occurrences). It is the gap between what we know and what we have that causes the limited predictability of recent drought events. To fill the gap, this thesis suggests that the design of a multi-scale study for drought-generating mechanisms is necessary. An update of our understanding of the mechanisms taking account of their historical changes will lead to a more robust improvement in our predictability and provides a more actionable information for water resources management and policy.
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Chapter 1

Introduction

1.1 Background

Drought is a naturally occurring climate phenomenon that may persist from one season through several seasons and even multiple years. It is defined as a period of abnormally dry weather, sufficiently long enough to cause a serious hydrological imbalance (AMS 2003). Drought is classified into four main types: meteorological, hydrological, agricultural, and socio-economic drought, as defined by deficits of precipitation, streamflow, soil moisture, and water supply, respectively.

Drought is one of the most costly natural hazards in the United States (Wilhite et al., 2000). It causes economic, social, and ecological impacts on society. During the drought of the 1930s, most of the United States, especially in the Midwestern region, experienced severe agriculture losses, large-scale migration, and ecological disruption. During the 1950s, another major drought persisted over most of the U.S. for several years, and the state of Texas was declared a federal drought disaster area. The 1988 “flash” drought was relatively short-term, but its severity and spatial extent were comparable to the 1930s drought. As a result, it caused the worst agriculture losses
in US history, about US $39 billion.

Drought is a hydroclimate extreme event induced by interactions between land, atmosphere, and ocean, possibly compounded by human activities as well. For example, the Dust Bowl of the 1930s was initiated by a combination of strong negative (cold) and positive (warm) anomalies of sea surface temperatures (SSTs) over the Pacific and the North Atlantic Oceans, respectively (Schubert et al., 2004b). These persistent SSTs modulated large-scale atmospheric circulation, inducing anti-cyclone anomalies at the mid-troposphere level and strong high pressure anomalies at the upper-troposphere level in extra-tropical regions. These large-scale pressure systems blocked moisture supply from the Gulf of Mexico. The Dust Bowl was exacerbated by positive anomalies of land surface temperature and high evaporative demand, as well as poor agricultural practices, which resulted in soil erosion and the worst dust storm in U.S. history (Worster, 1979). The mechanisms of the Dust Bowl were relatively well studied due to its significant socio-economic impact. Still, our understanding of the generating mechanisms of drought is insufficient due to their complexity.

In a changing climate, some regions of the U.S. have become more vulnerable to drought because land surface, atmospheric circulations, and oceans are altered by global warming. In the southeastern U.S. region, annual low flows have decreased due to increasing temperature over the late 20th century (Patterson et al., 2012), which elevates the risk of low flow drought. Also, summertime precipitation over the southeastern U.S. has intensified in recent decades by changes in large-scale circulation driven by global warming (Wang et al., 2010). It leads to more intense precipitation events during wet periods, and also more persistent precipitation deficits during dry periods, which increases the risk of flood and drought over the region.

Since the late 1990s, the phases of the Pacific Decadal Oscillation (PDO) and Atlantic Multi-Decadal Oscillation (AMO) have been altered. Also, global warming
trends have compounded these multi-decadal SST variations since the 1950s. These recent changes of the oceans are associated with climate change and its natural variability at a decadal scale. It is noteworthy that the influences of these recent changes on drought-generating mechanisms have not been fully understood. As a result, several regions of the U.S. failed to mitigate severe droughts in recent years. In 2011, severe drought hit the southern U.S. with the worst severity around the State of Texas, which caused over US $7 billion in agriculture losses (Fannin, 2012). Also, the 2012 Midwestern U.S. drought caused US $20 billion in agriculture losses (Booton, 2012). In 2013, the State of California declared a drought state of emergency with over US $2 billion in agriculture losses (Howitt et al., 2014). The failures of recent drought forecasting underscore the importance of better understanding the changes in drought-generating mechanisms induced by climate change and variability. By investigating these changes, we can protect against the adverse impacts on society.

1.2 Motivation

Over the past half century, U.S. drought studies have focused mainly on 1) identifying and characterizing drought occurrences based on a coarse temporal scale (month, season, year) (Palmer, 1965; McKee et al., 1993; Sheffield et al., 2004; Andreadis and Lettenmaier, 2006; Mo, 2008; Sheffield et al., 2009) and 2) developing statistical models of drought characteristics (Sen, 1980; Matheir et al., 1992). These studies are based on one-dimensional temporal and spatial scales, which limits our understanding of multi-scale drought-generating mechanisms via land-atmosphere-ocean interactions. Therefore, a multi-dimensional spatial scale study of U.S. drought is crucial to understanding the favorable terrestrial, atmospheric, and oceanic conditions for drought occurrence that provide more reliable information for drought forecasting.
Recent changes on land, and in the atmosphere and oceans question the stationarity of the known drought-generating mechanisms for the U.S., which means that major drivers may be changed in the generating processes of recent drought events. The non-stationarity of known drought mechanisms, however, has been little understood and has to be considered, as the previous studies focused on the mechanisms themselves. The limitations of the previous studies are due to the data available and a single climatology. Therefore, a multi-scale temporal study of U.S. drought is necessary in order to examine the stationarity of the known mechanisms.

To conduct a multi-scale study of drought, long-term and recently updated hydrometeorological datasets, that are physically consistent, are necessary. In recent years, the hydroclimate research community has been endeavoring to generate physically consistent and long-term data products for the North American climate system, including terrestrial, atmospheric, and oceanic variables using data assimilation techniques (Mesinger et al., 2006; Mitchell et al., 2004). These products are generated in real time and updated every month, enabling the exploration of drought mechanisms for the U.S. across multi-spatial and multi-temporal scales. Therefore, this thesis proposes a multi-scale study for U.S. drought risk and predictability, with a focus on understanding the generating mechanisms.

### 1.3 Goals & Overview

This thesis seeks to better understand 1) how land, atmosphere, and oceans interact during the onset of drought, 2) how the processes involved in generating mechanisms of drought can be better identified by a multi-scale study, rather than a single dimensional study, and 3) how reliable the known drought mechanisms are in explaining and predicting recent and future droughts. The goals of this thesis are motivated by
the following key questions:

“**What terrestrial, atmospheric, and oceanic conditions are favorable for U.S. drought occurrence?**”

“**What are the hidden processes involved in the U.S drought mechanisms that are not seen within a single-dimensional framework?**”

“**Can the known drought-generating mechanisms explain the occurrence of recent and future droughts?**”

This thesis consists of five core chapters. Chapter 2 addresses the first question by investigating the attributes of changes in low flows over the eastern U.S. region. It also examines the association between low flows and large-scale atmospheric circulations. Chapter 3 and 4 are also generated in response to the first question. Chapter 3 investigates the multi-scale driving mechanisms of droughts and pluvials over the southeastern U.S. using a recent regional reanalysis dataset. It focuses on a comprehensive understanding of favorable conditions for droughts and pluvials over the southeastern U.S. in the land, atmosphere, and oceans. In Chapter 4, the influence of tropical cyclones on eastern U.S. drought is assessed as a protective factor of eastern U.S. drought risk by merging two different datasets (precipitation and tropical cyclone track information) and conducting experimental simulations of a land surface hydrology model forced with or without tropical cyclone-related rainfall. Chapter 5 addresses the second question by investigating changes in drought risk influenced by Pacific and Atlantic Oceans. Chapter 6 responds to the last question. It assesses the current predictability of U.S. drought using a multi-model ensemble forecasting system based on the known associations between global sea surface temperature and U.S. precipitation from previous studies. It is motivated by the disparity among seasonal forecasts of the 2012 Midwestern U.S. drought from different climate models.
1.4 Publications

In this dissertation, one core chapter (Chapter 2) has been submitted for publication and four core chapters (Chapter 3 through 6) have been published. Below are the full references:


Chapter 2

Changes in the Low Flow Regime over the eastern United States: Variability, Trends, and Attributions

2.1 Background

Stream flow is a hydrological response of the land surface at the catchment scale to precipitation and is a major water resource available for human use and ecological needs. During the dry season, low precipitation (low supply) and high evapotranspiration (high demand) conspire to reduce stream flow and water availability. This natural seasonal hydrological phenomenon is known as the low flow period (Smakhtin, 2001). Low flows may also occur in the cold season when frozen soils and/or rivers cause a reduction in flow. With variations in climate, decreases in seasonal precipitation and increases in potential evapotranspiration can lead to reductions in low
flows, with adverse effects on human activities and ecosystem function (Bradford and Heinonen, 2008). These low flow season hydrological droughts can be exacerbated by water management (Wada et al., 2013) with a range of impacts from environmental (e.g. migration of wildlife) to economic (e.g. increased costs due to water scarcity). For example, since the 1950s, groundwater withdrawals in the U.S. have increased dramatically and can affect low flow generation at the regional scale (Castle et al., 2014; Hughes et al., 2012; Konikow, 2013). Future projections of changes in hydrological droughts indicate that they will become more intense and longer (Hayhoe et al., 2007; Feyen and Dankers, 2009) even with adaptation to mean changes in climate (Wanders et al., 2014). Given the impacts of low flow droughts and the potential for changes in the future, better understanding of the low flow regime and its generating mechanisms is necessary.

Drivers of low flow variability are complex, and include antecedent precipitation, atmospheric demand, surface water management (urbanization, dams, reservoirs, and irrigation), and groundwater withdrawals. Antecedent precipitation (AP) is one of the major drivers. During the low flow season, when the catchment is relatively dry, streamflow is an accumulated response and the impact of any precipitation events is delayed and muted. This impact depends on the geographical characteristics of the catchment (e.g. regional climate, terrain slope, soil type, basin size, groundwater dominance, hydrological connectivity, etc.). For example, Figure 2.1a shows the delayed response to AP for USGS station ID 01531500 (located at Towanda, PA), for which it takes up to three months from the last major streamflow peak to reach the annual minimum, and the impact of any precipitation events is small compared to the wet season. Better understanding of the variability of low flows and its driving mechanisms can provide invaluable information for improving seasonal forecasting of low flows, quantifying hydrological drought risk, and understanding the potential
future changes in low flows under climate change and direct anthropogenic influences.

Figure 2.1: (a) Time series of daily stream flow and precipitation at one station (USGS ID: 01531500) for 1998. (b) Time series of annual 7-day low flows and 30-day total antecedent precipitation at one to three month lead before the date of the annual low flow event (e.g. the antecedent precipitation accumulated during 0-29, 30-59, and 60-89 days, respectively) for 1962-2011.

Low flows are particularly important over the eastern U.S. where the interaction of human activities with natural systems is relatively high. The hydroclimatic regime over the eastern U.S. has changed over the past half century. During the 1950s-1990s, low and median flows (but not high flows) have increased (Lins and Slack, 1999; Douglas et al., 2000) due to increasing fall precipitation (Small and Vogel, 2006). McCabe and Wolock (2002) found that these increases were generally due to abrupt changes around the 1970s. More recent studies (Patterson et al., 2012; Sayemuzzaman and Jha, 2006; Sadri et al., 2014) found that some regions of the eastern U.S., includ-
ing the Mid-Atlantic and Southeastern regions, have decreasing trends in annual and seasonal averaged flow and increasing trends in surface temperature. Coopersmith et al. (2014) found that temporal shifts in daily precipitation and stream flow statistics have been observed after the 1970s and more significantly after 1980s. Future scenarios from climate and hydrological models (Hayhoe et al., 2007) show significant decreasing trends in warm-season low flow statistics (e.g. low flow volumes) due to increasing temperature. The observed non-stationarity in the low flow regime and the projected changes suggest that the risk of hydrological drought in eastern U.S. is increasing and may continue in the future. In this paper we evaluate variability in the low flow regime over the eastern U.S. to 1) understand long-term changes and their attribution, and 2) to identify potential oceanic and atmospheric drivers of inter-annual variability. We use this information to try to understand the potential for seasonal prediction of low flow anomalies and to identify scenarios associated with elevated streamflow drought risk. We do this by examining long-term flow records from USGS stations and exploring the connections with large-scale climate drivers in the context of climate variability and change. In section 2.2, the data and methods are described. Section 2.3 presents the results using different versions of the Mann-Kendall (MK) trend test and explores the connections between inter-annual variability of low flows and large-scale climate drivers, as well as other possible drivers. In section 2.4 we summarize our findings.


2.2 Data and Methods

2.2.1 Stream Flow Data

We downloaded daily stream flow records for 4878 stations over the eastern U.S. from the U. S. Geological Survey (USGS) Water Data for the Nation database. 2811 of these stations have remained active over the last decade. We calculated annual 7-day low flows (Q7) defined as the minimum low flow in each calendar year from the 7-day moving window time series of daily stream flows. Most stations have a fairly narrow season for Q7 during August-October (ASO), while some stations in North and South Carolina have a wider season (July-October). Following Sadri et al. (2014) we evaluated the Q7 time series for evidence of management via statistical testing for step changes and examination of station notes. Step changes may be due to a change of gauge datum, change of gauging instrument, or implementation of water management by a reservoir or withdrawal (Survey, 2014). We used the non-parametric Pettitt test (Pettitt, 1979) to identify step changes. 149 stations were identified with no step change and continuous records over the period 1962-2011 (Figure 2.2). This period was chosen as it is long enough (50 years) to identify long-term changes and the set of stations has good spatial coverage of the domain. These stations are located in the northeastern (NE) and Mid-Atlantic (Mid-ATL) U.S. regions, and the northern part of the southeastern (SE) region (North and South Carolina). The USGS annual water-data reports indicate that 97 out of the 149 stations have been regulated before or after 1962 (dams, power plants, mills, reservoirs, etc.), although any impacts on the low flows were not large enough to be picked up by the Pettitt test. The rest of the selected stations have no record of regulation, and we therefore assume that the low flow data are unaffected by regulation (empty circles in Figure 2.2). This information enables us to qualitatively assess the impact of human impacts on the
low flows regime over the study region.

2.2.2 Precipitation Data

For precipitation, we use a recently published daily dataset from Livneh et al. (2013; ftp://ftp.hydro.washington.edu/pub/blivneh/CONUS/) to compute antecedent rainfall over the catchments upstream of the 149 stations. The data covers the continental U.S. (CONUS) at 1/16th degree (6km) spatial resolution for 1915-2011. We use the latest version (version 1.2), which corrects a previous wet bias in daily precipitation before the 1950s. Antecedent precipitation (AP) is calculated over the basin corresponding to each station for each low flow date, and for a range of window sizes (30, 60, ..., 300, 330-day) to understand the lagged relationship with low flows. The catchment masks are derived from 30 arc second digital elevation model data and scaled up to 1/16th degree. Some of the stations share a few grid cells in their catchment masks (Figure 2.2).

2.2.3 Potential Evapotranspiration Data

We also use monthly average potential evapotranspiration (PET) from the Climate Research Unit (CRU) TS v3.21 dataset. The data have 0.5 degree (50 km) spatial resolution for 1901 to 2012. PET is defined as the maximum possible evapotranspiration given no restrictions on available water and is used as an indicator for atmospheric water demand. The PET data are calculated using the Food and Agricultural Organization (FAO) grass reference evapotranspiration equation (Ekstrom et al. 2007) using temperature, water vapor pressure, and cloud cover from the CRU dataset.
Figure 2.2: Map of the locations of the 149 USGS stations and the masks of their corresponding upstream basins. The sizes of blue circles represent the sizes of the basins.

2.2.4 Climate Indices

To identify teleconnections, we use seven monthly climate indices from the NOAA Earth System Research Laboratory and compute 3-month averages over 1962-2011 for different periods from January-March (JFM: -8 month lead to the peak season (ASO) of low flow events) to September-November (SON: +1 month lag). The strongest teleconnections (based on correlation) with Q7 are found with the North Atlantic Oscillation (NAO) and Pacific-North American pattern (PNA). The analysis excludes decadal SST variations over the Pacific and Atlantic Oceans due to the relative short study period (50 years).
2.2.5 Trend Test

To quantify trends in the Q7 time series and associated precipitation and PET, we use the non-parametric Mann-Kendall (MK) trend test. The MK trend test requires that the data be serially independent input because otherwise this will lead to overestimation (underestimation) of the significance of trends with positive (negative) serial correlation in the data, which often exist in long-term hydroclimate records. Therefore, we use three different versions of the MK test (Kumar et al., 2009): 1) MK test without autocorrelation (MK-0), 2) MK test with one-lag autocorrelation and trend-free pre-whitening (MK-1), and 3) MK test with complete autocorrelation structure (MK-2). The difference among the results from the different versions of the MK tests show the impact of autocorrelation structure in the hydrological data on the significance of trends. We compute the Theil-Sen (TS) slope of the local trends which is the median of all sub-trends in the series. We use a significance level of alpha=0.05. We also test the field significance of the local trends to account for spatial correlation among stations using the False Discovery Rate (FDR) approach (Wilks, 2006). The local trend is defined as field significant when the local significance (alpha) is less than or equal to the global significance level (here, alpha=0.05) weighted by its rank over the total station number (i/N; i is the rank of the local station and N is the number of stations (here, N=149)).

2.3 Results

2.3.1 Correlation Between Q7 and Antecedent Precipitation

To identify the time period of antecedent precipitation that is most associated with Q7, we compute temporal correlations between Q7 and monthly (30-day) total an-
Antecedent precipitation at different lagged times from -1 month (-29 to 0 days from the date of occurrence of the low flow event) to -11 months (-329 to -300 days). For most of the stations the inter-annual variability of Q7 has a maximum correlation with monthly total antecedent precipitation at 1 to 3 months (Figure 2.3). We therefore focus on the 90-day total antecedent precipitation (AP-90) when examining the attribution of trends in Q7.

Figure 2.3: (a) Lag/lead time with maximum correlation between Q7 and EAP-30 from +1 lag to -11 month lead from the dates of Q7 occurrence for 1962-2011.

2.3.2 Trends in Q7 Low Flows for 1962-2011

The MK-0 trend test for Q7 indicates that 12 stations (out of 149) show significant increasing trends and 11 stations show significant decreasing trends for 1962-2011 (Figure 2.4 (a) and Table 2.1). Five and three stations in the NE region have signif-
icant increasing and decreasing trends, respectively. Over the northern part of the Mid-ATL region, seven stations show a significant increasing trend while there are no stations with significant decreasing trends. In the southern part of the domain, including North and South Carolina, there are no stations with an increasing trend and eight stations with a significant decreasing trend. The decreasing trends over the southern part of the domain are consistent with the results from the MK-1 and MK-2 tests. Among the stations with significant trends, the slope (expressed in mm/day by dividing by the basin area) is 0.0026 mm/day/year and -0.0054 mm/day/year, respectively, on a regional basis. For comparison, the mean and standard deviation of Q7 is 0.008 mm/day and 0.009 mm/day, respectively. Based on the FDR approach, four stations are field significant among eleven stations with significant decreasing trends, and 1 out of 12 stations with a significant increasing trend is field significant. This indicates that the results from the MK tests are overestimated by the strong spatial correlation in hydro-climate variables over the study region (Andreadis and Lettenmaier, 2006).

2.3.3 Attribution of Trends in Q7 Low Flows

The MK-0 test for AP-90 shows that 42 stations have significant increasing trends over the NE and Mid-ATL regions and 2 stations have significant decreasing trends over the SE region (Figure 2.4 (d); Table 2.1). The results from the MK-1 and MK-2 tests are very similar due to the weak serial correlation structure in AP-90. A field significance test shows that 14 stations and one station with increasing trend and decreasing trend, respectively, are field significant. The increasing and decreasing trends in Q7 over the NE and South Carolina are consistent with the corresponding trends in AP-90, suggesting that changes in antecedent precipitation are driving the
changes in low flows. Interestingly, the Mid-ATL region shows weak consensus of the trends in Q7 and AP-90, with the consensus weaker during 1982-2011 than during 1962-1991 (Figure 2.5). Five out of eleven stations in this region with significant decreasing trends in Q7 are flagged in the USGS notes as having experienced some form of regulation, indicating that regulation may have overwhelmed any increases in low flows from increasing precipitation.

Changes in PET may have also contributed to changes in low flows. Figure 2.6 shows the results from the MK-0 test for PET over three periods (1962-2011, 1962-1991, and 1982-2011). Here, we present the results for averaged PET over the 6-month warm season (May-October). These results are consistent with the results from using a 3-month moving window of the warm seasons (May-July through August-October;
Table 2.1: Number of stations with significant trends at significance level alpha=0.05 and their averaged Theil-Sen slope using the standard Mann-Kendall test. The parenthesis shows the number of stations with field significant trends and their averaged slope.

<table>
<thead>
<tr>
<th>7-day Annual Low Flow (Q7)</th>
<th>Positive</th>
<th>Negative</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Number of Stations</td>
<td>TSA Slope [mm/day/year]</td>
</tr>
<tr>
<td></td>
<td>1962-2011</td>
<td>(1)</td>
</tr>
<tr>
<td></td>
<td>1962-1991</td>
<td>(1)</td>
</tr>
<tr>
<td></td>
<td>1982-2011</td>
<td>(-)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>90-day Total Antecedent Prec. (AP-90)</th>
<th>Positive</th>
<th>Negative</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Number of Stations</td>
<td>TSA Slope [mm/day/year]</td>
</tr>
<tr>
<td>1962-2011</td>
<td>44 (16)</td>
<td>0.0187 (0.0218)</td>
</tr>
<tr>
<td>1962-1991</td>
<td>30 (-)</td>
<td>0.0364 (-)</td>
</tr>
<tr>
<td>1982-2011</td>
<td>9 (-)</td>
<td>0.0360 (-)</td>
</tr>
</tbody>
</table>

not shown). Parts of the NE and Mid-ATL regions along the Atlantic coast show significant increasing trends with a slope of 0.006 and 0.013 mm/day/year for 1962-2011 and 1982-2011, respectively, on a regional basis. Due to the coarse resolution of the PET data, the field significance of the PET trends are not tested. Even though the significance of these PET trends might be overestimated, their magnitude is comparable to those of antecedent precipitation, which suggests that the regional climate regime over the Carolinas has become drier.

Over South and North Carolina, a combination of decreasing trends in AP-90 (less water supply) and increasing trends in PET (higher water demand), has resulted in stronger decreasing trends in Q7 than any other states in the region. The trends are stronger in recent decades indicating that this region is becoming more vulnerable to
drought, possibly as a result of surface warming, and likely exacerbated by regulation.

### 2.3.4 Longer Term Trends for Selected Stations

The climate of the eastern U.S. is subject to decadal variations, which may impact the robustness of the trends. In other words, a trend may be a part of natural decadal variation. We examine a subset of 14 stations with no regulation and data for 1932-2011 in terms of the trends in Q7 and AP-90 (Figure 2.7). The spatial distributions of the trends in Q7 and AP-90 for 1932-2011 is similar to that for 1962-1991 rather than 1982-2011. This suggests that the low flow regime over the eastern U.S. has changed more significantly since the 1980s, with weak consensus of the trends in antecedent
Figure 2.6: Same as Figure 2.4 except for the period 1932-2011 for stations with available data.

precipitation since other drivers have played a role in more recent decades.

Figure 2.7: Spatial distribution of the trends in potential evapotranspiration (PET) from the MK-0 test for (a) 1962-2011, (b) 1962-1991, and (c) 1982-2011.

2.3.5 Teleconnections with NAO and PNA

In Figure 2.8, we show the 3-month time period with the maximum temporal correlation coefficients between the NAO and PNA, and detrended Q7 (dQ7) over 1962-2011.
for different 3-month lead times (JFM (-8 month lead) through SON (+1 month lag)). The peak season for dQ7, ASO, is defined as 0 month lag and a significance level of 0.05 is used to determine the significance of the correlation coefficients. The NAO is correlated positively and significantly with dQ7 over the Mid-Atlantic and SE regions at one month lead (July-September), which has previously been found to be a favorable condition for summer droughts over the SE (Chapter 3). The NAO is correlated positively but insignificantly with dQ7 over most of the NE region during early spring (March-May). The PNA shows negative and significant correlations with dQ7 over most of the study region. The Mid-ATL region and the southern part of the NE region have significant correlations at two month lead (June - August), while other regions show significant correlations in late winter (January-March). The results for stations without recorded management show similar spatial patterns, indicating that the impact of regulation on inter-annual variability in Q7 is small for these sites.

Figure 2.8: Map of the 3-month time period (colors) for each station which has the maximum temporal correlation between Q7 and the (a) North Atlantic Oscillation and (b) Pacific-North American pattern, for 1962-2011.
2.4 Summary

Our findings can be summarized as follows: 1) A dipole pattern of increasing and decreasing trends in Q7 low flows exists in the northern and southern part of the domain, respectively; 2) The increasing trends in the northern part are associated with increasing antecedent precipitation. The Mid-ATL region and the southern part of the SE region (North and South Carolina and Virginia) have significant decreasing trends in Q7 values that have increased the risk of hydrological droughts, and are possibly linked to increasing trends in PET driven by warming temperature; 3) water management has likely played a role in modulating the impact of climate change on low flows; 4) the NAO and PNA are potential predictors for the occurrence of extreme low flow events, such that a negative NAO and positive PNA form a favorable condition for drought, and vice versa.

These findings suggest that the risk of low flow droughts over the eastern U.S. are higher during persistent negative NAO and positive PNA conditions, and with elevated evaporative demand, and when impacted negatively by human impacts. This suggests that continued climate change and increasing human pressures on water resources may exacerbate hydrological droughts in the future. The direct anthropogenic impacts on low flows, however, remain uncertain and are highly dependent on local hydrological connectivity, although several studies have identified (Boschi et al., 2003; Brutsaert, 2010) linkages between groundwater abstractions and low flows across the eastern coastal plain aquifer. Inter-annual variability in low flows appears to be linked to the PNA and NAO, and recent improvements in the predictability of the NAO (Scaife et al., 2014), at least for the wintertime, provide prospects for low flow prediction. Further work is needed to understand the interplay with larger scale Rossby waves and decadal variations in climate, which may modulate the teleconnections
shown here (Hanna et al., 2014).
Chapter 3

A Multi-scale Analysis of Drought and Pluvial Mechanisms for the southeastern U.S.

3.1 Background

Drought over the U.S. has been studied extensively in terms of its causes and impacts, with particular focus on the central U.S. because of its position as the bread-basket of the country, and the southwestern U.S., because of the fragility of water availability and increasing demands from a burgeoning population. The central U.S., in particular, has experienced severe droughts during the 1930s, 1950s, late 1980s and 2000s. The 1930s Dust Bowl drought was derived from persistent high pressure systems over the central U.S., anomalous tropical sea surface temperatures (SSTs), and strong land surface-atmosphere coupling (Schubert et al., 2004b) and human-induced land degradation (Cook et al., 2009). The 1950s drought has been associated with cold phases of the Pacific Decadal Oscillation (PDO) and the El-Nino and Southern
Oscillation (ENSO) (Barlow et al., 2001), and the water phase of the Atlantic Multidecadal Oscillation (AMO) (McCabe et al., 2004). The 1988 drought showed similar atmospheric and oceanic conditions to the 1930s drought (Trenberth et al., 1988).

Figure 3.1: (a): Time series of annual precipitation anomalies over 1979-2008. (b), (c), and (d): Spatial distribution of annual precipitation anomalies over the eastern US in 1981, 1986, and 2007. The anomalies are based on the climatology period of 1979-2008. The study region is shown by the box.

Droughts over the southeastern (SE) U.S. (box in Figure 3.1) have been less well understood, which is partly due to the complexity of its climate but also its relative abundance of water. Over the last three decades the region has experienced several severe droughts with different characteristics due to high inter-annual and intra-seasonal precipitation variability. In 1981, the SE experienced a severe “flash” drought (Figure 3.1 (b)), which was driven by below-normal summertime precipitation, the least over the last three decades, and above-normal temperature. The drought was alleviated by tropical storm Dennis and above-normal winter precipitation. A multi-year moderately severe drought occurred during 1984-1988 (Figure
3.1 (c)), which was initiated by below-normal precipitation in the autumn of 1984. Precipitation during the summer of 1985 was not enough to recover this drought, and this was exacerbated by the least wintertime precipitation in 1985 over the past three decades. The drought of 1988 extended to the central U.S. and covered 58% of contiguous U.S. (CONUS) during the summer (NCDC, 2012). The most recent long-term drought during 2006-2008 was initiated from below-normal precipitation during the autumn and winter of 2005 (Figure 3.1 (d)). The severity of the drought peaked in 2007, with the least annual precipitation over the last three decades. The expanding population over the region has compounded the impacts of the droughts on water supply, agriculture and ecosystem health (Keim, 1997; Silliman et al., 2005). During 2006-2008 drought, agricultural damages of one billion U.S dollars were incurred, and water storage and quality for municipal use deteriorated, prompting multi-state water resources restrictions and legal action (Manuel, 2008; Campana et al., 2012). This drought prompted a number of studies to investigate the causes of droughts over the SE using observations and modeling (Seager et al., 2009; Ortegren et al., 2011).

Precipitation over the SE is derived from diverse moisture sources that vary with driving mechanisms occurring at different temporal and spatial scales, including land-atmosphere feedbacks, variability in regional-scale circulation, and persistent oceanic conditions over the north Atlantic, and the north and tropical Pacific. Moisture sources are derived from 1) advected moisture from surrounding regions, and 2) recycled moisture from within the region from local evapotranspiration. Variations in moisture transport play a critical role in regional climate variability and have strong seasonality over the CONUS (Shukla and Mintz, 1982; Ruane, 2010). For the southern U.S., the Great Plains Low Level Jet (GPLLJ) plays a dominant role in moisture transport (Higgins et al., 1997; Mo et al., 2005; Mestas-Nunez and Enfield, 2007; Trenberth et al., 2011), bringing moisture from the Gulf of Mexico into the Great
Plains during the summer, but which may be deflected to the S.E. by high pressure systems over the central U.S. (Berbery and Rasmusson, 1999; Mo et al., 2005). Such synoptic scale circulations plays an important role in driving variability in moisture transport and precipitation over the SE (Seager et al., 2009; Wang et al., 2010; Diem, 2013), with important features including the western ridge of the North Atlantic Subtropical High (NASH) (Li et al., 2011) and the upper-level jet (Wang et al., 2010), and westerly and northwesterly lower-troposphere flux over the SE region (Diem, 2006). Seager et al. (2009) suggested that inter-annual variability of summer precipitation over the SE is forced by purely internal atmospheric variability. However, there are suggestions that this situation has been made more complex in recent years, because of the potential effects of climate change. For example, summer precipitation variability over the SE has intensified owing to changes in the NASH that have been linked to recent warming (Wang et al., 2010; Li et al., 2011).

The contribution of the land surface to moisture availability and precipitation depends on the regional hydroclimate and the strength of feedbacks with the atmosphere (Dirmeyer, 2011; Ferguson et al., 2012). The recycling ratio is the ratio of recycled precipitation from a local source via evaporation to total precipitation and can be used to quantify the strength of land-atmosphere coupling and its impact on precipitation variability. Several recycling models have been proposed over the past three decades based on various assumptions, such as a fully mixed atmosphere and negligible changes in atmosphere moisture at monthly to seasonal time scales (Brubaker et al., 1993; Eltahir and Bras, 1994; Burde and Zangvil, 2001; Dominguez et al., 2006; Dirmeyer and Brubaker, 2007; van der Ent et al., 2010). Using such a model, Dominguez and Kumar (2008) demonstrated that during the summer, the eastern U.S. exhibits strong coupling between the land and atmosphere with the strength of coupling depending on soil moisture availability. For the SE in particu-
lar, Dominguez et al. (2006) reported that about 30% of summertime precipitation is recycled on average.

At broader time and space scales, the oceans play one of the most important roles in driving climate globally. Marshall et al. (2001) review the ocean teleconnections for the North Atlantic and note that anomalous sea surface temperatures tend to force large scale circulation over the U.S. and Europe, contributing to anomalous precipitation. McCabe et al. (2004) found that the positive phase of the AMO generally introduces more frequent drought events over the CONUS and that the cold phase of the PDO coupled with the warm phase of the Atlantic Ocean influences the spatial extent of the droughts over the CONUS. Climate model studies (Feng et al., 2011; Nigam et al., 2011) show that the warm phase of the AMO decreases summer and autumn rainfall over the US, especially for the south and SE regions, but increases precipitation over the west African monsoon region. Furthermore, atmosphere-ocean interactions are also critical for precipitation variability over the SE. During the boreal winter (December-February; DJF), air-sea flux variability can drive anomalous SST patterns over the North Atlantic (Gulev et al., 2013). Anomalous SSTs over the Pacific can modulate the westerly jet stream in the mid-troposphere leading to anomalous large circulation across the CONUS (e.g. 1988 drought) (Trenberth et al., 1988). North Atlantic tropical cyclones (TCs) play a moderate role in impeding and alleviating droughts over the SE depending on their frequency and rainfall intensity (Chapter 4). However, despite that the influences of the Pacific and Atlantic oceans on continental scale droughts over the CONUS have been well studied, their roles in regional scale drought over the SE are less well understood.

Of course, these individual mechanisms do not act alone but interact at different time and space scales. For example, inter-annual variability of the north Atlantic is linked to development of TCs that play a role at intra-seasonal time scales in
modulating drought occurrence and recovery (Chapter 4). Therefore, a thorough understanding of drought mechanisms for the SE requires a multi-scale study of land-atmosphere-ocean interactions. Here, we use the North American Regional Reanalysis (NARR) to examine the land-atmosphere-ocean conditions and feedbacks that are associated with drought (and pluvial) conditions over the SE, how they vary seasonally and how consistent these conditions are between different events. The SE is defined here as 30°-35°N, 91°-82°W and is shown in Figure 3.1 (b)-(d) as outlined by the black box. This area primarily includes the states of Mississippi, Alabama, and Georgia with small portions of Arkansas and South Carolina. We focus on meteorological droughts and pluvials based on precipitation anomalies during 1979-2008, but also examine anomalies in other parts of the terrestrial water budget. Section 3.2 describes the data and methods used. In Section 3.3, we present the hydroclimate over the SE and compare its variation between the drought and pluvial years at regional scale. Section 3.4 investigates the typical conditions of the atmosphere and oceans during the development of droughts and their potential driving mechanisms, at ocean basin to planetary scales, and then Section 3.5 focuses on how these conditions evolved for the recent multi-year drought (2006-2008) across the various land, atmosphere, and ocean scales. In Section 3.6, we summarize our findings and discuss their implications for the predictability of SE droughts and its application to seasonal forecasting.

3.2 Data and Methods

3.2.1 Datasets

The NARR (Mesinger et al., 2006) is a state-of-the-art regional reanalysis for North America and adjoining oceans that was proposed to better capture weather and cli-
mate variability at regional and continental scales and associated hydrometeorological extremes at high resolution. The NARR has numerous improvements compared to previous global reanalyses. It is based on the NCEP Eta atmospheric model (Black, 1988), its data assimilation system (Zupanski and Mesinger, 1995), and the Noah land-surface model (Ek et al., 2003). It uses lateral boundary conditions from the NCEP-DOE Global Reanalysis (Kanamitsu et al., 2002). The NARR includes land surface, atmospheric, and oceanic variables at high temporal (3-hour) and spatial resolution (32 km). It has 29 vertical layers and covers from 1979 to the present. Notable improvements in the NARR over previous global reanalyses include finer resolution and additional/improved observational datasets, assimilation of precipitation and introduction of a comprehensive land surface model (Mesinger et al., 2006). Past studies have shown that the NARR is a useful data resource for exploring atmospheric and terrestrial water cycles at basin and regional scales (Mo et al., 2005; Ruiz-Barradas and Nigam, 2006; Dominguez and Kumar, 2008; Li et al., 2010; Ruane, 2010; Kumar and Merwade, 2011), for examining extreme hydrologic events such as pluvials and droughts (Mo and Chelliah, 2006; Weaver et al., 2009; Neiman et al., 2011; Sheffield et al., 2012a), and as an observational estimate for validation of other reanalyses and coupled models (Sun and Barros, 2010; Mei and Wang, 2012; Ferguson et al., 2012). We use the NARR precipitation and soil moisture data at monthly scale to calculate anomalies to represent the timing and severity of droughts and pluvials. Additional datasets of geopotential heights (GHs) and SSTs are used to analyze teleconnections with SE hydroclimate beyond the domain of the NARR. SSTs are taken from the monthly Extended Reconstructed SST version 3b (ERSSTv3b) dataset (Smith et al., 2008) and monthly GHs at 500 mb and 850 mb from the NCEP/NCAR global reanalysis (Kalnay et al., 1996).
3.2.2 Analysis of Moisture Fluxes

We compute the seasonal mean vertically integrated moisture fluxes (the surface to 250 mb; ignoring the negligible amount of water vapor above 250 mb) from the 3-hourly data over 1979-2008 as follows:

\[
\bar{Q}_\lambda = \frac{1}{g} \int_{p_0}^{250 \text{mbar}} \bar{q} \bar{u}
\]

(3.1)

\[
\bar{Q}_\phi = \frac{1}{g} \int_{p_0}^{250 \text{mbar}} \bar{q} \bar{v}
\]

(3.2)

The overbars represent the seasonal means over 1979-2008. In this study, other forms of liquid and frozen water in the atmosphere are ignored. The overbar represents a seasonal average derived from the three hourly data. We partition the total column moisture transport into the lower (1000 mb - 850 mb) and upper tropospheric levels (825 mb - 250 mb) in order to assess the contribution of low/high level jets to total moisture transport over the SE. We partition the moisture fluxes into their mean component and transient eddies (Peixoto and Oort (1992); Eqn.3.3 and Eqn. 3.4)

\[
\bar{Q}_\lambda = \frac{1}{g} \int_{p_0}^{250 \text{mbar}} \bar{q} \bar{u} + \frac{1}{g} \int_{p_0}^{250 \text{mbar}} \bar{q}' \bar{u}'
\]

(3.3)

\[
\bar{Q}_\phi = \frac{1}{g} \int_{p_0}^{250 \text{mbar}} \bar{q} \bar{v} + \frac{1}{g} \int_{p_0}^{250 \text{mb}} \bar{q}' \bar{v}'
\]

(3.4)

The transient eddy terms, \(\bar{q}' \bar{u}'\) and \(\bar{q}' \bar{v}'\), are the cross-covariance between the wind and humidity variations computed from the difference between the seasonal means of the product of the 3-hourly specific humidity and wind, and the product of the seasonal means of specific humidity and wind (\(\bar{q} \bar{u} - \bar{q} \bar{u}\) and \(\bar{q} \bar{v} - \bar{q} \bar{v}\)). For the zonal
eddy moisture fluxes, a positive (negative) value indicates higher moisture flux than
the average in the westerly (easterly) direction (Brubaker et al., 1994).

Sub-integrals for each boundary of the rectangular region were computed following
the method of Simmonds et al. (1999). Seasonal fluxes (mm/day) along the
four boundaries were computed by aggregating the 3-hourly moisture fluxes (ks/m/s)
along the respective boundary (kg/s) and dividing by the area of the region. Then,
the regional moisture flux convergence was computed by summing the representa-
tive fluxes along the four boundaries. Positive values represent convergence, that is,
the region is a moisture sink, and negative values represent divergence or a moisture
source. Despite the recommendation of a broader study area (> 10^6 km^2) (Rasmus-
son, 1977), our study area for the SE is 0.4 * 10^6 km^2, which is supported by the
high-resolution data of the NARR. To validate this, the moisture flux convergence
averaged over the long term (1979-2008) was compared with the long-term average
of the difference between regional precipitation and evapotranspiration. Under the
assumption that the precipitable water tendency is negligible, these two quantities
should be equal.

3.2.3 Precipitation Recycling

We use the concept of precipitation recycling as a measure of the strength of land-
atmosphere coupling and its impact on drought development and persistence. Recy-
cled precipitation is defined as precipitation originated from local sources via evap-
otranspiration and recycling ratio is a ratio of recycled precipitation to total pre-
cipitation over a region during a given time. We use a regional recycling ratio (r)
as an indicator for the strength of land-atmosphere coupling based on the Brubaker
et al. (1993) recycling model. It can be applied at monthly and seasonal time scales,
although it has been found to underestimate the strength of recycling (Burde and Zangvil, 2001). Our estimates of the recycling ratio for the SE from the NARR are consistent with the estimates across the CONUS from different recycling models (Dirmeyer and Brubaker, 2007; Dominguez and Kumar, 2008).

### 3.2.4 Water Budget Non-closure in the NARR

Uncertainties exist in estimates of regional moisture flux convergence from reanalysis products because of the assimilation of observations (Higgins et al., 1997; Ruane, 2010). The errors in an individual product can be estimated by the analysis increments, which are defined as the difference between the analysis field and the model first guess and in general represents the impact of observations on the model forecast (Higgins et al., 1997). In the NARR, assimilation of precipitation observations introduces moisture increments in the atmospheric column, which include water vapor and water condensate increments, to reduce errors in moisture tendencies. Here, we consider only the water vapor increment. For example, over the SE, overestimation in simulated precipitation introduces negative moisture increments into the atmospheric column without adjustment of the overestimates in evaporation and/or moisture convergence. The maximum value of the increments occurs in the summer with a magnitude comparable to the total convergence itself (Ruane, 2010). This means that there are large uncertainties in the NARR data especially during the summer. The increments are larger for pluvial years but relatively small for drought years due to overestimation of precipitation (Ruane, 2010) and evaporation (Sheffield et al., 2012a) by the NARR atmospheric and land models and the imbalance in the water budget because of the lack of feedback from the land surface (Dominguez et al., 2006; Luo et al., 2007; Bisselink and Dolman, 2008; Weaver et al., 2009; van der Ent...
et al., 2010; Ruane, 2010). Over our study region, the moisture increments are 1.29 mm/day on annual average, which contributes to most of the residuals from the atmospheric water budget (1.5 mm/day).

3.2.5 Selection of drought and pluvial years

To select pluvial and drought years, we focus on meteorological drought defined by anomalous precipitation relative to the 1979-2008 climatology, which is a precursor to agricultural and hydrologic drought. We choose the six wettest and driest years based on the ranks of seasonal precipitation anomalies (Figure 3.2) and then composite hydroclimate variables for the pluvial and drought years. The winter period is defined as from December in the previous year to February of the current year. For example, the winter of 2006, which was the fourth driest winter of the time period, is from December 2006 to February 2007.

Figure 3.2: Time series of precipitation anomalies over the southeastern US during 1979-2008 for (a) summer, and (b) winter.
3.3 Moisture Transport and Water Budgets over the Southeastern U.S.

We show results for different aspects of the water budgets over the SE: the transport of moisture into the region, the land and atmospheric water budgets, and the recycling of precipitation. We compare the climatologies of the moisture transport on a seasonal basis and the water budget and recycling ratio on a monthly basis with the composites for the six driest and wettest years.

3.3.1 Moisture Transport: Climatology, Drought, and Pluvial Years

The mean and transient eddy components from the NARR are presented in Figure 3.3. Across the eastern US, the mean moisture flux component dominates the total moisture transport via the Great Plain Low Level Jet (GPLLJ) during summer while the transient eddies play a minor role. During winter, the mean component dominates the total moisture transport in the zonal direction due to large scale meanders of the jet stream, that is, Rossby waves along mid-latitude regions (Schubert et al., 2011). In the meridional direction, however, the mean and transient eddy components are comparable (Brubaker et al. 1994). Surface temperature maps for the summer and winter climatology show large gradients along the boundary between moisture convergence and divergence by the transient eddy component (Figure 3.3 (e) and (f)). This indicates that transient eddy moisture fluxes are associated with heat exchange between the land and ocean and also via storm tracks, especially during the winter (Chen, 1985).

The summertime total-column moisture flux convergence over the SE is weakly
Figure 3.3: (a) and (b): Mean component of moisture fluxes during summer and winter. The box represents the study region. (c) and (d): Climatology of transient eddy component of moisture fluxes during summer and winter. The background shading is the moisture convergence by each component. (e) and (f): Spatial distributions of summertime and wintertime surface temperature averages over the eastern US and the adjacent oceans during 1979-2008. To aid visualization, the vectors are shown at a coarser spatial resolution (> 150km) than the original data (32km).

Divergent (very close to zero) in the climatology (1979-2008), due to higher evapotranspiration from the land surface during summer, while it is strongly convergence (2.5 mm/day) during the winter (Figure 3.4 (a)). The weak summer divergence is derived from weak zonal moisture transport deficit (surplus) between the west (north) side of the region and the east (south) side of the region. During winter, moisture convergence is dominated by the surplus of moisture transport in the meridional direction. During the six driest years, weak meridional moisture fluxes along the Gulf of Mexico drive relatively strong divergence around our study region during the summer and
relatively weak convergence during winter. This indicates that the moisture transport in the meridional direction plays an important role in moisture convergence over the SE and its inter-annual variability can influence moisture availability over the region. The total column mean and transient eddy components are both convergent over the SE during winter, indicating that advected moisture into the region modulates the variation of wintertime precipitation over the SE, and especially in relation to the negligible recycling of moisture (see below). Interestingly, strong (weak) high-level meridional moisture transport along the coastline is associated with summertime and wintertime pluvials (droughts) over the SE, while consistent moisture is transported by low level moisture flux from the Gulf of Mexico regardless of the season. This suggests that extreme summer and winter precipitation is related to fluctuations of high-level meridional moisture transport over the south side of region.

Figure 3.4: The magnitude of moisture transport for the SE in terms of regional moisture convergence and meridional and zonal moisture fluxes, for (a) the climatology (1979-2008), (b) six driest years, and (c) six wettest years, during summer (JJA) and winter (DJF). Total column moisture flux convergence into the region is represented by the middle box (positive: convergence and negative: divergence) and moisture fluxes at each of the regions four boundaries are represented by the boxes in each cardinal direction. Red (blue) bars are total column moisture transport in the zonal (meridional) direction. Positive values in the zonal (meridional) direction represent westerly (southerly) fluxes. Total column (dark shading) fluxes are separated into lower level (medium shading) and upper level (light shading).
Total column moisture flux is partitioned into the mean (Figure 3.5 (a), (b), and (c)) and transient eddy component (Figure 3.5 (d), (e), and (f)). Over 1979-2008, the moisture flux convergence by the mean component compensates moisture divergence by the transient eddies, resulting in weak moisture flux convergence during the summer. Based on the climatology, the transient eddy component plays a critical role in moisture flux convergence during the winter since most of the moisture flux convergence is derived from the transient eddy component. During the six driest summers and winters, the low-level mean component (from surface to 850mb) in the meridional direction is a major portion of moisture transport over the study region, and the direction of the mean component at low level and high level is opposite to each other. This indicates that there is heat exchange from land to ocean via transient eddies, which suggested warm SSTs in the Gulf of Mexico. During the six driest winters, transient eddies also play a dominant role in the regional moisture transport. Interestingly, strong (weak) high-level meridional moisture transport along the coastline is associated with summertime and wintertime pluvials (droughts) over the SE, while consistent moisture is transported by low level moisture flux from the Gulf of Mexico regardless of the season. This suggests that extreme summer and winter precipitation is related to fluctuations of high-level meridional moisture transport over the south side of region. This indicates that for the six driest (wettest) summers and winters, moisture flux divergence (convergence) is mainly derived from transient eddies, which are related to sub-seasonal meteorological phenomenon driven by atmosphere variability, as noted by Seager et al. (2009), rather than more persistent oceanic forcing. This suggests that the potential predictability of droughts at seasonal time scale over the SE might be limited. However, the results demonstrate that there exists a consistent contrast of meridional moisture transport along the coastline of the Gulf of the Mexico between drought and pluvial years, and better understanding
of its driving mechanisms can lead to improved predictability as discussed in section 3.4.

Figure 3.5: Same as Figure 3.4 except for the mean component (top) and the transient eddy component (bottom) of total moisture fluxes over the study region

3.3.2 The Regional Water Budget: Climatology, Drought and Pluvial Years

Monthly anomalies of moisture flux convergence and precipitation have large intra-seasonal variability (standard deviation = 3 and 2.5 mm/day, respectively; Figure 3.6 (a) and (b)) and are highly correlated (correlation coefficient > 0.7) (Table 3.1). The terrestrial water budget components (evaporation, runoff and soil moisture) have progressively lower short-term variability (Figure 3.6 (c), (d), and (e)) indicating that they have longer persistence compared to components of the atmospheric budget (precipitation and moisture flux convergence, MFC). Soil moisture at one-month lag
shows the highest temporal correlation with precipitation (Table 3.1), indicating that it takes about one month for the land surface to respond to atmospheric anomalies.

Figure 3.6: Time series of anomalies of the components of the land surface and atmosphere water budgets based on the climatology period of 1979-2007. Blue (red) represents positive (negative) anomalies of each component.

Figure 3.7 shows the climatology of components of the terrestrial and atmospheric water budgets for the SE and compares it with those during the six driest and wettest summers. We focus on the warm season (JJA) to examine the maximum interaction of the land and atmospheric components. To quantify the impact of precipitation assimilation in the NARR on the water budgets, the moisture increments are also represented in Figure 3.7. Over the SE, the overestimation of precipitation by the
Table 3.1: Lag correlation coefficients among the monthly averages of the components of land surface and atmospheric water budgets during 1979-2008. Negative (positive) indicates that precipitation (first five rows) and soil moisture (last four rows) lead (lag) other current components by the indicated month.

<table>
<thead>
<tr>
<th>Lag/Lead</th>
<th>-3</th>
<th>-2</th>
<th>-1</th>
<th>0</th>
<th>+1</th>
<th>+2</th>
<th>+3</th>
</tr>
</thead>
<tbody>
<tr>
<td>MFC</td>
<td>0.06</td>
<td>0.08</td>
<td>0.06</td>
<td>0.85</td>
<td>0.12</td>
<td>0.12</td>
<td>0.08</td>
</tr>
<tr>
<td>SM</td>
<td>0.45</td>
<td>0.48</td>
<td>0.53</td>
<td>0.36</td>
<td>0.17</td>
<td>0.16</td>
<td>0.11</td>
</tr>
<tr>
<td>E</td>
<td>0.30</td>
<td>0.35</td>
<td>0.28</td>
<td>0.03</td>
<td>0.08</td>
<td>0.05</td>
<td>-0.06</td>
</tr>
<tr>
<td>Q</td>
<td>0.18</td>
<td>0.26</td>
<td>0.38</td>
<td>0.59</td>
<td>0.23</td>
<td>0.13</td>
<td>0.04</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Lag/Lead</th>
<th>-3</th>
<th>-2</th>
<th>-1</th>
<th>0</th>
<th>+1</th>
<th>+2</th>
<th>+3</th>
</tr>
</thead>
<tbody>
<tr>
<td>MFC</td>
<td>0.01</td>
<td>0.07</td>
<td>0.79</td>
<td>0.23</td>
<td>0.38</td>
<td>0.36</td>
<td>0.35</td>
</tr>
<tr>
<td>E</td>
<td>0.54</td>
<td>0.59</td>
<td>0.63</td>
<td>0.61</td>
<td>0.52</td>
<td>0.47</td>
<td>0.40</td>
</tr>
<tr>
<td>Q</td>
<td>0.38</td>
<td>0.43</td>
<td>0.47</td>
<td>0.55</td>
<td>0.56</td>
<td>0.51</td>
<td>0.47</td>
</tr>
</tbody>
</table>

NARR atmospheric model is reduced by adding negative moisture increments in the atmosphere while there is no adjustment to the overestimates of evaporation and moisture convergence. Especially during the wetter seasons, the moisture increments are over -2.0 mm/day while they are less than -1.0 mm/day during the drier seasons. This indicates that the NARR overestimates evaporation and/or moisture flux convergence over the SE and therefore can overestimate the recycling of rainfall as analyzed in the next sub section. The residuals from the atmospheric water budget (Figure 3.7; P-E-MFC) are 1.5 mm/day averaged over 1979-2008, which are derived mostly from the precipitation data assimilation (the increments: -1.2 mm/day).

There is a large correlation between MFC and precipitation (Table 3.1). Evaporation has a marked seasonal cycle, which peaks in June and is at a minimum in December causing a transition from an overall moisture source (evaporation exceeding precipitation) during the summer to a moisture sink during winter. Runoff, which includes surface runoff and subsurface baseflow, peaks in the spring as driven by the peak in precipitation and low evaporation from winter to spring. The inferred change in terrestrial water storage (Figure 3.7; P-E-Q) is positive in the winter and early
Figure 3.7: Terrestrial and atmospheric water budget components over the SE region on a monthly basis for the climatology (1979-2008), six driest years, and six wettest years ((a), (b), and (c), respectively). The six driest and wettest years are selected based on summer precipitation. Residuals are shown from the terrestrial water budget components (open circle) and atmospheric water budget components (closed circle) and including the moisture increments (open rectangle) over the climatology (1979-2008), six driest years, and six wettest years ((d), (e), and (f), respectively). Components are precipitation (P), evapotranspiration (E), runoff (Q), moisture flux convergence (MFC), and analysis increment (INC).

During the summer drought years, the MFC decreases from spring onwards driving less precipitation over the SE with no peak in June. On the other hand, the summer pluvial years show a higher peak of MFC in June, which is related to a concurrent increase in precipitation. The NARR indicates that there is a significant increase in the atmospheric water budget during summer pluvial years and a decrease during summer drought years as expected. In drought years, higher solar radiation due to reduced cloud cover, and higher surface temperatures (not shown), causes evaporation to exceed precipitation from May through August and enhances the role of soil
moisture in providing moisture to the atmosphere to compensate the high evaporative
demand. In pluvial years, the opposite changes occur. Thus, drought years tend to
have stronger coupling between the land surface and atmosphere than normal years,
which can be quantified by the recycling ratio as shown in the next section in terms
of the inter-annual variability of recycling.

3.3.3 Recycling Ratio and Recycled Rainfall: Climatology, Drought Years and Pluvial Years

The recycling ratio and recycled precipitation over the SE are shown in Figure 3.8. Over
the course of the year, the recycling ratio increases from near zero in winter to a maximum
in August, when the recycling ratio is over 0.1, meaning that more than 10% of monthly
precipitation is derived from local evaporation. The variance of the recycling ratio also
increases during the warm season (Figure 3.8 (a)). The recycling ratio and its variance
are higher during summer droughts (mean of about 0.15) than during summer pluvials
(mean of about 0.06). To assess the actual impact of recycling on precipitation, the
recycled precipitation is computed from the monthly precipitation multiplied by the recycling ratio. Interestingly, there is more recycled precipitation on average (0.5 mm/day) than during extreme years. During summer pluvials, precipitation is recycled a little bit more (0.4 mm/day) than during drought years (0.3 mm/day), despite a lower recycling ratio, since the absolute size of precipitation is much higher. These estimates are likely overestimated, however, because of the overestimates in evaporation and/or moisture flux convergence discussed above. The low average recycling ratio indicates that the land surface plays a moderate role in summer precipitation over the SE during normal years, mainly due to its proximity to ocean sources of moisture.
Figure 3.8: Recycling ratio and recycled precipitation over the SE on a monthly basis over the climatology (1979-2008; (a) and (c)), six driest years ((b) and (e)), and six wettest years ((c) and (f)). (a) and (d): the boxes represent a range of one standard deviation of the recycling ratio and recycled precipitation. (b), (c), (e), and (f): the error bars represent the minimum and maximum of the recycling ratio and the recycled precipitation on a monthly basis during the six driest and wettest years. The circles represent their monthly means.

3.4 Multi-scale Atmospheric and Oceanic Drivers:

Warm and Cold Seasons

To identify typical conditions of large-scale circulation during drought and pluvial years, we plot the difference in moisture fluxes (vectors) and 500mb GHs (shading) over the CONUS between drought and pluvial years (Figure 3.9 (a) and (b), respectively). During summer drought years, moisture transport from the Gulf of Mexico into the SE is generally blocked by anomalous anticyclonic motion in the mid-troposphere induced by positive GH anomalies over the Central U.S. During
winter drought years, a dipole (high-low) anomalous pressure system brings dry air from Canada through the Great Plains down to the SE. Dong et al. (2011) also identified this pattern in their study of the winter 2006 Great Plains drought. These anomalous patterns suggest that different driving mechanisms for extreme events exist over the SE region at seasonal scale. We explore next the driving mechanisms of these patterns at different spatial scales, from regional up to planetary scales, first during warm seasons, and then during cold seasons.

Figure 3.9: The differences of total moisture fluxes (vector; mean + transient eddies) and the geopotential height at 500 mb (background) between the drought and pluvial years ((a): summer and (b): winter).
3.4.1 Warm season variability

During dry summers, a strong high pressure system at 850 mb (green contour in Figure 3.10) extends to the southern U.S. and thus its western ridge is shifted northward, which is derived from strong positive GH anomalies over the north Atlantic and the central U.S. This drives less precipitation along the Gulf of Mexico coastal states. This situation is reversed during wet summers (Figure 3.10 (b)). This high pressure system is well known as the NASH, which is a major driver of summer climate variability for the SE (Wang et al., 2010). The strength of the NASH is generally measured by the extent of the 1560 geopotential meter (gpm) lines near Bermuda, and quantified by the Bermuda High Index (BHI) and Western Bermuda High Index (WBHI). Here, the definition of the BHI and the WHBI are from the study of Diem (2013), where the BHI (WBHI) index is computed from the normalized difference of sea level pressures (850-hPa geopotential heights) between two regions marked as circle (cross) in Figure 3.10. During dry summers, the NARR show positive values of the BHI and WBHI, which derive from positive and negative surface pressure anomalies over the land and oceans, respectively. The positive values of the BHI and WBHI are related to the blocking effect of moisture transport from the Gulf of Mexico in the meridional direction and thus it results in below-normal precipitation. Also, the shift of the NASH westward reduces the number of landfalling TCs in the SE by deflecting their paths to the southern U.S. (Elsner and Kocher, 2000). Summer droughts are also associated with warmer SSTs in the north Atlantic (55°-25°W, 25°-55°N) and cooler SSTs over the north Pacific (160°-130°W, 30°-60°N) while positive anomalies of SSTs over the Nio 3.4 region are weakly associated in terms of the magnitude (Figure 3.10 (c)). We quantify the consistency of the anomalous patterns by counting the number of the years that show the same sign as the composited patterns with a
threshold of more than four years in agreement (that is 5 or 6 years). During dry summers (Figure 3.10 (c)), these anomalous SST patterns over the north Pacific and north Atlantic are consistent across individual dry summers (hatched areas in Figure 3.10 (a)), however a warmer phase of the tropical Pacific is consistent only during wet summers. The contrast in summertime SST anomalies between the north Pacific and north Atlantic is notable and is reminiscent of the pattern identified at decadal or longer scales (the cold phase of PDO-the warm phase of AMO) by McCabe et al. (2004) and Schubert et al. (2008) as favorable for drought conditions for the U.S. in general.

The center of the anomalous SST pattern over the north Atlantic is well matched with the center of the NASH, suggesting that either the ocean or atmosphere could be a predictor over the Atlantic for the other. To examine the mechanisms, the compositied maps for 500mb GH are calculated from two month lead time (-2 month) to two month lagged time (+2 month) during summer droughts and pluvials (Figure 3.11). Due to the small sample size, we chose Barnard’s exact test to estimate the statistical significance of the signs of the composite geopotential height anomalies during droughts and pluvials, as compared to during non-drought and non-pluvial years. In general, Barnard’s exact test is used to test the independence of rows and columns in a contingency table (Barnard, 1945). Here, the two rows of the table are: droughts (pluvials) and non-droughts (non-pluvials); and the two columns are: the signs of the composited 500mb geopotential height anomalies and their opposite sign. The null hypothesis is that the signs of the composite 500mb geopotential height anomalies are independent of droughts or pluvials conditions at the 0.1 significance level (hatched areas in Figure 3.11 and 3.13). During summer droughts, consistent and positive anomalies of 500 mb GH over the north Atlantic are greatest at two month lead time (AMJ; Figure 3.11 (a)) and persist until zero lead time, after which it
Figure 3.10: Geopotential heights anomalies at 850mb ((a) and (b)) and SST anomalies ((c) and (d)) during summer drought and pluvial years. The black contours represent climatological geopotential heights and the green contour, 1560 gpm, represents the drought or pluvial year anomalies. The Bermuda High Index (Western Bermuda High Index) is computed from the differences of sea level pressures (geopotential heights at 850mb) between the two circles (crosses). The north Pacific and Atlantic are represented as a box in Figure (c). Hatched areas represent areas with at least four of the six driest or wettest years having the same sign as their average. The climatology is computed based on 1979-2008.

disappears at one month lag time (Figure 3.11 (g)), and vice versa. This indicates that a strong NASH during late spring forces positive north Atlantic SST anomalies and thus leads to reduced precipitation over the SE. Therefore, variability in the NASH is one of main drivers of summer droughts over the SE. During summer pluvials, consistent 500 mb GH anomalies over the Gulf of Alaska and north Atlantic persist from two month lead time until zero time, and their locations are well matched with the summertime SST anomalies over the north Pacific and north Atlantic (Figure 3.11 (b)). This implies that the atmosphere forces the anomalous SSTs over the north
Pacific and Atlantic during late spring (AMJ) via Rossby waves at mid-latitudes and then these anomalous GHs and SSTs result in the summertime pluvials. Overall, the results suggest that SE hydroclimate has a symmetric connection with conditions over the Atlantic but an asymmetric connection with the Pacific. Zhao et al. (2011) suggested that during the summer, these subtropical anticyclones over the Pacific and Atlantic and anomalous SSTs might be associated with the Asian-Pacific Oscillation at interdecadal scale, which is predictable by some climate forecast systems (Chen et al., 2013).

Figure 3.11: Leading and lagged composited maps for geopotential height (GH) anomalies at 500 mb relative to climatology (1979-2008) for drought summers and pluvial summers at two month lead time ((a) and (b): AMJ), one month lead time ((c) and (d): MJJ), zero month lead time ((e) and (f): JJA), one month lagged time ((g) and (h): JAS), and two month lagged time ((i) and (j): ASO). Hatched areas represent areas with statistically significant positive (negative) anomalies at 10 per. level based on Barnard’s exact test.
3.4.2 Cold season variability

Winter droughts (pluvials) show a westward (eastward) shift of a low pressure system (green line in Figure 3.12 (a)), which is located between Iceland and southern Greenland for the climatology (the contour black line of -5 gpm in Figure 3.12 (a) and (b)). This low pressure system is the Icelandic Low, which is downstream of the mid-tropospheric wave trough towards Europe (Blackmon et al., 1984). Over the north Atlantic, the IL teleconnection with the pressure system over Bermuda was introduced by the study of Blackmon et al. (1984) and its inter-annual variability is linked with inter-annual Arctic sea ice extent variations and the North Atlantic Oscillation (NAO) (Deser et al., 2000). For example, eastern US winter droughts relate to higher values of the IL index, which resembles the negative phase of the NAO. The IL index is calculated from 500mb GH anomalies on a regional basis (55°-70°N, 60°W-10°E). Our results demonstrate that the westward shift of the IL results in a higher value of the IL and contributes to a reduction (increase) in precipitation over the SE.

The winter drought years are associated with weak but consistent cold phases of SSTs in the Nio 3.4 region and the north Pacific (Figure 3.12 (c)). The anomalous SST patterns during dry winters are close to the SSTs pattern of “the perfect ocean for drought” during 1998-2002 as identified by Hoerling and Kumar (2003). During winter pluvials, the El Nio phase over the tropical Pacific is strong but inconsistent across individual wet winters in the composite SST maps (Figure 3.12 (d)) since it is derived from two strong El Nio in 1982 and 1997. There is a horse-shoe like SST anomaly pattern over the central Pacific, similar to the negative phase of the PDO. This is consistent across years during winter pluvials (hatched areas in Figure 3.12 (d)). This suggests that the north Pacific is related to winter pluvials over the SE.
Figure 3.12: Same as 3.10 except for winter

and the signal from the North Atlantic is weak and inconsistent during dry winters.

During the winter droughts, the composited maps for mid-tropospheric GHs indicate that the IL (box in Figure 3.13) peaks at one month lead time (NDJ) due to a strong high-low dipole pattern of mid-tropospheric GH anomalies, which then weakens gradually and disappears during the early spring (FMA). At one month lead time there is a strong atmospheric forcing of SST anomalies over the Atlantic and winter droughts over the SE, with the opposite mechanism in place during the winter pluvials. Negative 500mb GH anomalies over the central Pacific appear abruptly at zero lead time and persist until early spring (FMA), indicating that during winter pluvials, SSTs over the central Pacific force the atmosphere, and induce more precipitation over the CONUS, especially the SE. Furthermore, we constructed composite maps for 500mb geopotential height anomalies at monthly time step (October-February;
The monthly composite maps demonstrate the north-south dipole pattern of anomalous 500mb geopotential heights (low-high) from November through February of the next year, which is representative of the typical pattern of a positive IL. Our results demonstrate that the IL at one month lead time step can be used as a predictor of winter droughts over the SE while a negative phase of PDO forces the atmosphere and then induces winter pluvials over the SE.

Figure 3.13: Same as Figure 3.11 except for winter ((a) and (b): OND; (c) and (d): NDJ; (e) and (f): DJF; (g) and (h): JFM; (i) and (j): FMA).

### 3.5 A Case Study of the 2006-2008 SE Drought

As a case-study, we examine the roles of the above-discussed climate drivers in the recent 2006-2008 SE drought by examining of the evolution of hydroclimate vari-
ables and atmospheric and oceanic drivers. Figure 3.14 shows monthly time series of hydroclimate variables normalized by their standard deviations for 1979-2008. It also shows related atmospheric and oceanic drivers: monthly time series of the IL index, the BHI and the WBHI, SSTs over the Nio 3.4 region, the north Atlantic SSTs averaged over the region 55°-25°W and 30°-60°N, and the difference between the regional averaged SSTs over the north Atlantic (55°-25°W, 25°-55°N) and north Pacific (160°-130°W, 30°-60°N), each normalized by its standard deviation. Positive values of the northwestern Atlantic/northeastern Pacific difference indicates a stronger SST contrast which is associated with less precipitation during summer as shown above. The IL index is computed from the regional average (60°W-10°E, 55°N-70°N, the dashed box in Figure 3.9) of the normalized 500mb GHs. Figure 3.14 also shows time series of TC related rainfall anomalies again normalized by the standard deviation to indicate the strength of TC activity (Chapter 4).

During the late summer of 2005, below-normal moisture flux convergence led to less precipitation (including close to normal TC-related rainfall). This period is coincident with a strong positive phase of the IL index. This induced negative soil moisture anomalies beginning in the fall of 2005, reaching a minimum during the late spring of 2007. During the spring and early summer of 2006, high potential evaporation increased evapotranspiration, but this turned to negative anomalies in subsequent years because of the lack of available soil moisture, despite persistently positive potential evaporation anomalies. These combined to force positive recycling ratio anomalies during the summer, of over 0.3. During the winters, the positive phase of the IL was derived from the southward shift of the minimum in GH anomalies, which brought a reduction in winter precipitation over the SE. The north Atlantic was in a warm phase during the entire period and the wintertime SSTs over the Nio 3.4 region (5°S-5°N, 170°W-120°W) showed the La-Nia phase (a cold phase) during the subsequent
Figure 3.14: Time series of standardized anomalies of (left) atmospheric and land surface conditions over the SE, and (right) atmospheric and oceanic conditions over surrounding oceans, for the multi-year SE drought of 2005-2008.

winters, which were also associated with a reduction in SE precipitation. During the summers of 2006-2009, differences between SSTs in the north Atlantic and north Pacific were positive and provided favorable oceanic conditions for less summertime precipitation over the SE (McCabe et al., 2004). Also, the WBHI index indicates that there existed a strong NASH during the summers of 2006 and 2007. From August 2008, the precipitation deficit over the SE region weakened gradually and more moisture transport and TC-related rainfall produced above-normal summertime precipitation, leading to drought recovery in the early spring of 2009. This was linked to negative phases of the IL, the BHI, and the WBHI through the winter of 2008. In summary, this drought was mainly forced by the atmosphere, especially a strong NASH and IL, and persisted by a combination of favorable conditions in different
aspects of the land and ocean.

3.6 Summary

This study has evaluated the mechanisms associated with drought over the SE U.S. from the point of view of the land, atmosphere and ocean, as well as the conditions associated with pluvial events. The evaluation is based on the NARR, which while providing a consistent and continuous dataset of relevant variables for exploring extreme hydrologic events over the CONUS, does have limited spatial and temporal coverage and is subject to uncertainties at all scales. Of importance is the non-closure of the water budget as derived from its data assimilation method that despite being novel in its indirect assimilation of observed precipitation results in significant closure residuals. This will impact the strength of the calculated anomalies, for example, the overestimate in moisture flux and recycling ratio computation, but is unlikely to affect the sign of the anomalies or the overall conclusions of this study.

The hydroclimate of SE U.S. is subject to different dominant moisture sources and different climate drivers depending on the season. Typical conditions for SE droughts (pluvials) are: 1) weaker (stronger) southerly meridional moisture fluxes and weaker (stronger) westerly zonal moisture fluxes, 2) strong moisture flux divergence (convergence) by transient eddies, that is, a strong (weak) heat and moisture exchange between tropical and extra-tropical regions, and 3) strong (weak) coupling between the land surface and atmosphere. For summer, a strong NASH at two month (AMJ) until zero month lead time (JJA) forces a warmer phase of the north Atlantic and SE drought conditions. For winter, a peak in the IL index in NDJ propagates a wave train in the mid-troposphere, which resembles the negative phase of NAO. This leads to less moisture convergence and less precipitation over the SE. For oceanic influences,
the roles of the Pacific and the Atlantic are different during droughts and pluvials depending on the season. The tropical Pacific induces more inter-annual variability in wintertime precipitation over the SE, the central Pacific forces the atmosphere during winter pluvials, and the Atlantic is forced by the inter-annual variability of the NASH during the summer. During summer droughts and pluvials, different Nio regions also play different roles in the inter-annual variability of SE precipitation.

Overall, the mechanisms for variability in SE hydroclimate are complex, especially during summer. Seager et al. (2009) suggested that SE precipitation variability is purely controlled by internal atmospheric processes and thus is unpredictable by a SST-forced global circulation model. This study also indicates that the symmetry of SE droughts and pluvials is mainly driven from the symmetrical atmospheric responses at one to two month lead time step despite asymmetric terrestrial and oceanic responses. Recent studies indicate that summer precipitation variability relates to the NASH (Wang et al., 2010) and to low frequency variability of the North Atlantic Ocean (Ortegren et al., 2011). This study has provided a better understanding of the land, atmosphere, and ocean conditions associated with historical SE droughts, which, despite the overall preponderance of atmospheric controls, shows potential for enhanced seasonal forecast skill via one to two month lead time of anomalies in the NASH and the IL. Further study is necessary to understand whether these associations are stationary. For example, variability in SE precipitation has increased in recent years as linked to warming of the north Atlantic (Wang et al., 2010), and Pacific SST teleconnections with CONUS precipitation and vary on multi-decadal time scales (Chapter 6).
Chapter 4

The Influence of Atlantic Tropical Cyclones on Drought over the eastern United States

4.1 Background

The North Atlantic basin provides a favorable environment for tropical cyclone (TC) genesis due to high sea surface temperatures from late spring until early winter (Emanuel, 1986). Landfalls of Atlantic TCs bring high winds and heavy rainfall to the eastern US on time scales of a few days causing significant societal and economic losses along the coastal states of the Gulf of Mexico and the North Atlantic (Pielke, 2009). For example, Hurricane Katrina in 2005 was the costliest hurricane in US history and the third deadliest since 1900 with 81 billion US dollars in economic losses and over 1800 fatalities in the hurricane and subsequent floods (Beven et al., 2008). Despite the direct devastation and economic costs that are caused by TCs, they also have a beneficial side in terms of the amount of water they bring to the
land and their role in drought recovery.

Drought is a naturally occurring and prolonged climate phenomenon that, like TCs, is one of the most costly of natural disasters (Wilhite et al., 2000). For example, the estimated agricultural losses from the Texas 2011 drought were about 7.6 billion US dollars and thus it was the costliest drought in the state history (Fannin, 2012). There is increasing awareness that TCs play a significant role in drought development and recovery, which has been highlighted in recent studies. Lam et al. (2012) discussed the role of TCs (typhoons) as drought breakers in Hong Kong, and their other benefits including contributions to wind energy and cooling effects. McGrath et al. (2012) observed that TCs are an important feature of vegetation dynamics and their quiescence along with other climate drivers may have induced the recent continental multi-year Australian drought from 1997 to 2011. For the example of the 2011 Texas drought, there were no landfalling TCs, which may have contributed to the development of the drought. Despite this growing awareness, the contribution of TCs to drought recovery has not been quantified and their role in drought dynamics is poorly understood, in part, due to their different temporal and spatial scales.

Previous studies investigating droughts over the contiguous US (CONUS) have focused mainly on identifying and characterizing historical drought events and have used monthly average time series of drought variables (Palmer, 1965; McKee et al., 1993; Sheffield et al., 2004; Andreadis and Lettenmaier, 2006; Mo, 2008; Sheffield et al., 2009). Most recently, Mo (2011) studied the historical onset and demise of drought over the US from monthly mean precipitation and simulated soil moisture. Since the lifetime of landfalling TCs is generally of the order of a few days to a week, the impact of TCs on drought is difficult to evaluate from monthly average time series.

The role of Atlantic TCs on the eastern US drought regime has not yet been quantified. Hence, this study was designed to answer the following questions: 1)
What is the contribution of Atlantic TCs to annual and seasonal total rainfall over the eastern US? 2) What is the impact of TCs on the eastern US drought regime at local to regional scales? 3) Do TCs play different roles in drought initiation, persistence, and recovery and how does this depend on different types of drought? 4) How do TCs change the characteristics of drought during either their more active years (2004-2005) or their more dormant years (2000-2001)?

In this study, the Variable Infiltration Capacity (VIC) land surface hydrologic model was run for two simulations with different rainfall forcing datasets. An EXP-TC simulation was forced by observed rainfall, including the contribution of TCs, and an EXP-NOTC simulation by rainfall excluding TCs. A drought was defined in terms of soil moisture as a prolonged period below a threshold corresponding to a certain soil moisture percentile and TC-related rainfall was defined as rainfall within a certain distance, 500 km, from the centers of TCs. (Shepherd et al., 2007; Jiang and Zipser, 2010; Barlow, 2011). Rainfall associated with TCs was determined from the Atlantic Hurricane Database (HURDAT) and the North America Land Surface Data System phase 2 (NLDAS-2). The analysis was carried out for 1980-2007 based on NLDAS-2 data availability. Detailed descriptions of the data and methods are given in section 4.2. Section 4.3 presents the historical contribution of Atlantic TCs during 1980-2007 to total rainfall, followed by an analysis of the impact of TCs on drought over the eastern US, firstly for the example of Hurricane Katrina on local drought and secondly for the impact of all TCs on regional drought events. In Section 4.4, we discuss uncertainties in our forcing data and potential impacts on the results, and the sensitivity of changes in drought to TC frequency and intensity and the implications of this under future potential climate change. In Section 4.5, we summarize the findings of this study and highlight the benefits of TCs to drought relief.
4.2 Datasets and Methods

4.2.1 Rainfall Forcing Dataset: The North Land Data Assimilation System Phase 2 (NLDAS-2)

The first phase of the North Land Data Assimilation System product (NLDAS-1) was initiated to support the development of more consistent and reliable initial land surface states for numerical weather prediction (Mitchell et al. 2004). Building on the experimental configuration of the first phase, the NLDAS-2 forcing data (Xia et al., 2012) covers a longer period from January 1979 to the present, via near real-time updates. The data are acquired from diverse sources including atmospheric reanalysis, satellite remote sensing and ground-based observations (Cosgrove et al. 2003; Xia et al. 2012). The NLDAS-2 has temporal and spatial resolutions of hourly and 0.125 degree (12.5 km), respectively. The NLDAS-2 precipitation forcing data is derived from the Climate Prediction Center (CPC) daily CONUS gauge data adjusted for topographical effect (Daly et al. 1994), hourly Doppler Stage II radar precipitation data (Crum et al. 1993), the CPC hourly CONUS/Mexico gauge data (Higgins et al. 2000), the half-hourly CPC Morphing method (CMORPH) data (Joyce et al. 2004), and three hourly North America Regional Reanalysis (NARR) rainfall data (Mesinger et al. 2006). Details of the hourly NLDAS-2 rainfall forcing data are described by Xia et al. (2012).

4.2.2 TC-related Rainfall from the Atlantic Hurricane Database (HURDAT)

The Atlantic Hurricane Database (HURDAT) (Jarviene et al., 1984; Neumann et al., 1993) provides best storm track data for 1851-2011. We combined it with the NLDAS-
forcing hourly rainfall dataset to identify TC-related rainfall for 1980-2007. The
Atlantic HURDAT include track information for the center of each TC at six hourly
time step, including maximum wind speed, minimum pressure, and their classification
as a tropical storm (TS) or hurricane with category between 1 and 5 based on their
sustained winds. Recent studies (Landsea et al., 2010; Kunkel et al., 2010) have used
the Atlantic HURDAT to analyze the trend in the duration of TCs and to examine the
contribution of TCs to heavy rainfall. Since the NLDAS-2 forcing data has an hourly
temporal resolution, the HURDAT TC track data were linearly interpolated from six
hourly to hourly. The data thus describes the evolution of the TC at hourly time
step (Villarini et al., 2011). The interpolated rainfall data have some uncertainties
for a number of reasons. The centers of the TCs do not move linearly in time along
the tracks between the six hourly positions from the HURDAT database and the
associated rainfall may therefore be higher or lower than estimated depending on these
dynamics. Furthermore, the spatial footprint of TC-related rainfall is generally not
symmetric and rainfall may fall preferentially on one side of the TC center. Therefore,
there are some uncertainties in the estimation of TC-related rainfall from our method
and the datasets used. Although several studies (Shepherd et al., 2007; Part and
Nelson, 2013) evaluated the contribution of TCs on total rainfall using the Tropical
Rainfall Measuring Mission (TRMM) TC database, which has temporal and spatial
scales of three hourly and 0.25 degree, the temporal coverage is only from 1998, which
is too short for our purposes and therefore, the NARR is the best candidate dataset
for this study. TC-related rainfall was assumed to fall within 5° from the center
of TCs as used in previous studies (Shepherd et al., 2007; Jiang and Zipser, 2010;
Barlow, 2011). The use of a threshold based on a fixed distance rather than fixed
latitude would make a slight difference to the results but would not affect the overall
conclusions.
We computed several metrics on a grid cell basis to describe the spatial distribution of TC-related rainfall and its contribution to total rainfall: frequency of TCs, total TC-related rainfall and its contribution to total rainfall, and TC rainfall intensity. TC frequency for each grid cell was calculated as the number of TCs whose center passed within 5 degrees of the cell. The TC rainfall intensity is the total TC-related rainfall divided by the frequency of TCs. We also calculated the total number of affected TC days, for each grid cell.

4.2.3 Off-line Land Surface Hydrologic Model Simulations

The VIC model has been developed and previously applied to better understand land surface hydrologic systems at catchment, regional and continental scales (Liang et al., 1994; Cherkauer et al., 2003; Nijssen et al., 1997, 2001; Pan and Wood, 2006). It has also been used to identify and monitor drought events at continental and global scales (Andreadis and Lettenmaier, 2006; Sheffield and Wood, 2007; Luo and Wood, 2007; Wang et al., 2011). The soil and vegetation parameters used in this study were calibrated by Maurer et al. (2002). The soil moisture at the beginning of 1980 was derived by running the model using forcing data from 1949 to 1979. From the initial condition on 1st January 1980, two VIC simulations were run: the first forced by the NLDAS-2 rainfall data and identified as the EXP-TC simulation and the second forced by non-TC related rainfall data and identified as the EXP-NOTC simulation. The EXP-NOTC rainfall data were calculated by removing the TC-related rainfall, as described in section 4.2.2, from the NLDAS-2 rainfall data. VIC simulations have temporal and spatial resolutions of daily and 0.125 degree, respectively, from 1st January 1980 until 31st December 2007 (28 years). The VIC model was run in so-called water balance mode, which requires forcing data including daily total rainfall, daily
maximum temperature, daily minimum temperature, and daily average wind speed. Therefore, the hourly NLDAS-2 rainfall and non-TC rainfall data were aggregated to daily.

Precipitation was the only forcing variable considered to be affected by TCs. However, other variables can be altered over a broad area around the center of TCs. For example, TC maximum sustained wind speeds are classified as being greater than 17 m/s (Blake et al., 2011). Latent heating can be significant with high evaporation rates and surface cooling (Trenberth et al., 2007). Therefore, TCs affect not only precipitation but also the three remaining forcing variables that are used as drivers of VIC. To generate the non-TC forcing scenarios more precisely, daily maximum and minimum temperature should be increased and daily average wind speed should decrease. Increased temperatures will likely increase evaporative demand and decreased wind speeds will reduce it. Therefore, their combined effects on evaporative demand and soil moisture may be negligible, and their impact will likely be overwhelmed by the rainfall forcing anyway.

4.2.4 Definition and Characteristics of Drought

A universal definition of drought is difficult because of the complex physical mechanisms and diverse effects on societal and economic sectors (Dracup et al., 1980; Mishra and Singh, 2010). Several drought indices have been developed and applied to detect and characterize historical drought events based on precipitation, soil moisture, streamflow, or other hydrological variables and combinations of these (Palmer, 1965; McKee et al., 1993; Byun and Willhite, 1999; Sheffield et al., 2004; Narasimhan and Srinivasan, 2005; Sheffield and Wood, 2007). Soil moisture from observations, remote sensing, and model simulation is a useful drought variable and has been used in past
studies (Andreadis and Lettenmaier, 2006; Hunt et al., 2009; Sheffield et al., 2009; Wang et al., 2011). In general, the shallow soil layer (from surface to 0.3m) responds quickly to short-term meteorological events, for example, rainfall, snowfall, and evaporation, whereas the deep soil layer (from 0.3m to 3m) is driven mainly by longer-term drivers including moisture redistribution and the seasonal cycle of evapotranspiration (Hunt et al., 2009). Therefore, the total column soil moisture in VIC embeds signals of meteorological and climate phenomena at short and long time scales and is a good indicator of drought, and especially agricultural drought (Sheffield et al., 2004; Narasimhan and Srinivasan, 2005). Agricultural drought is defined as a prolonged period of soil moisture under a certain threshold level that stresses vegetation and adversely impacts on its productivity. The threshold level can be set arbitrarily to detect a drought event and the relative magnitude of the drought can be defined for other parameters, such as, the intensity, duration, and severity of the drought (Dracup et al., 1980). Other drought indices could also be examined, such as the commonly used Palmer Drought Severity Index (PDSI) (Palmer, 1965). However, the PDSI is calculated at weekly or monthly time step and so is likely unsuitable for capturing the correct hydrological response to heavy rain rates that are typical of TCs (Alley, 1984).

This study evaluated drought based on VIC simulated daily total column soil moisture depth. The use of monthly averages, as often used in previous studies, would not allow the true impact of TCs, which act at much shorter time scales, to be properly evaluated. We used the threshold method for identifying droughts as described in Dracup et al. (1980). Here, two different types of threshold values were chosen to characterize the magnitude and duration of drought. First, we calculated three low percentiles (10th, 15th, and 20th percentiles) based on daily total column soil moisture depth for each model grid cell in the EXP-TC simulation. These corresponding soil
moisture depths were used as threshold values in measuring the magnitude of drought as a critical level of water availability in soil. We defined a drought as a period that has more than 30 continuous days (short-term drought) or 90 continuous days (long-term drought) below the soil moisture depth corresponding to each percentile threshold. We evaluated six types of drought ranging from moderately severe, short-term drought to extremely severe, long-term drought by combining the two durations (short-term and long-term) with the three thresholds (10 percentile: extremely severe; 15 percentile: severe; 20 percentile: moderately severe). Since droughts were defined relative to the local soil moisture climatology as represented by a percentile threshold value, the total number of droughts will be similar among the study regions. To describe of the temporal characteristics of drought, three metrics were introduced as follows and shown in Figure 4.1:

1. Drought initiation time \( (T_{ini}) \) is the day when a drought event starts
2. Drought recovery time \( (T_{rec}) \) is the day when a drought event terminates
3. Drought duration \( (D_d) \) is the total number of continuous days under a certain threshold level for one drought event.

To characterize the impact of TCs on drought, two further metrics were defined based on the differences between the EXP-TC and EXP-NOTC simulations:

4. Late drought initiation duration is defined as the total difference of drought initiation times between the two simulations \( (T^{INI}; \text{ Eqn. } 4.1). \)
5. Early drought recovery duration is defined as the total difference of drought recovery times between the two simulations \( (T^{REC}; \text{ Eqn. } 4.2). \)

For each model grid cell and each type of drought, we counted the total number of drought events \( (N_d) \) in both simulations and every occurrence of late drought
Figure 4.1: Example of total column soil moisture time series from 1st July 2005 to 5th October 2005. The solid line represents soil moisture from the EXP-TC simulation and dashed line represents the EXP-NOTC simulation. Drought characteristics are described in the text.

initiation ($N_i$) and early drought recovery ($N_r$) during the period 1980-2007. Example time series illustrating these indices are shown on Figure 4.1. Furthermore, we also computed averaged drought durations per event for each simulation ($D_{ave}$), the total difference in drought durations from both simulations ($D_{diff}$) averaged late drought initiation duration per event ($T_{late}$) and early drought recovery duration ($T_{early}$) during 1980-2007 for each grid cell.

\[ T^{INI} = \sum_{i=1}^{N_i} T_{ini,EXP-TC}(i) - T_{ini,EXP-NOTC}(i) \]  
\[ (4.1) \]

\[ T^{REC} = \sum_{r=1}^{N_r} T_{rec,EXP-TC}(r) - T_{ini,EXP-NOTC}(r) \]  
\[ (4.2) \]

\[ D_{ave} = \frac{\sum_{i=1}^{N_d} D_d(i)}{N_d} \]  
\[ (4.3) \]
\[ D_{diff} = \sum_{i=1}^{N_d} D_{d,EXP-NOTC}(i) - D_{d,EXP-TC}(i) \] (4.4)

\[ T_{late} = \frac{T^{INI}}{N_i} \] (4.5)

\[ T_{early} = \frac{T^{REC}}{N_r} \] (4.6)

### 4.3 Results

#### 4.3.1 Summary of the Atlantic Hurricane Data (HURDAT) database

The majority of landfalling TCs during 1980-2007 were tropical storms (33\%) and category 1 hurricanes (27\%), with 30\% of all TCs occurring in August and September (Figure 4.2). The greatest number of landfalling TCs during 1980-2007 occurred in 2004, with eight out of the total of nine TCs occurring between August and September. The number of landfalling TCs has increased since 1995 (numbers of hurricanes in 1980-1994 = 19; 1995-2007 = 26) and there have been more during the early TC season from May to June (numbers of TCs in 1980-1994 = 6; 1995-2007 = 16). Summaries of landfalling TCs can also be found in several previous studies (Landsea et al., 1996; Hart and Evans, 2001; Goldenberg et al., 2001; Shepherd et al., 2007; Landsea et al., 2010; Part and Nelson, 2013).
Figure 4.2: Annual total number of Atlantic TC landfalls by category (a) and by timing (b) by month over the United States for 1980-2007. Numbers stand for the Saffir/Simpson Hurricane Wind scales and TS represents tropical storm. The Saffir/Simpson Hurricane scale is based on the speed of sustained winds (1:119-153 km/h, 2:154-177 km/h, 3 (major): 178-208 km/h, 4(major): 209-251 km/h, and 5(major): 252 km/h or higher).

4.3.2 Contribution of TC-related Rainfall to Total Rainfall

Four study regions were identified for further analysis based on the intensity and frequency of TC-related rainfall and their contributions to total precipitation: South-eastern (SE), South, Northeastern (NE), and Midwestern (MW) regions (Figure 4.3 (a)). The South region had the most intense TC rainfall over our study regions due to the longer residence time as derived from the total number of TC days (not shown) divided by the total number of TCs. There is a local maximum in TC rainfall intensity (>50mm/event) over Oklahoma and Texas.

TCs have strong inter-annual and spatial variability because they are randomly occurring meteorological phenomena (Goldenberg et al., 2001). TCs contributed from 1% (1.5%) to 4% (7%) of annual (TC-season) rainfall (Figure 4.4 (a)) and the seasonal timing of landfalling TCs varied (Figure 4.4 (b)) depending on the region. The maximum annual (TC-season) contribution of TCs were 20% (28%), 17.7% (33%),
Figure 4.3: Maps of Atlantic TCs and related rainfall for 1980-2007 over the eastern US: (a) total TC-related rainfall, (b) the contribution of TC-related rainfall to total rainfall, (c) the frequency of TCs and (d) TC rainfall intensity. The study regions and the units for the maps are shown in the boxes and the text, respectively.

11% (17%), and 10% (17%) for the SE, South, NE, and MW regions, respectively, with the year of maximum occurring mostly in 2004 or 2005. The SE, NE, and MW regions showed increased contributions of TC-related rainfall in recent years while the South region showed relatively consistent contributions of TC-related rainfall over time. Rodgers et al. (2001) found that TCs contribute 4% of cumulative rainfall over the western North Atlantic and the south and southeastern US, which is in line with our results.

While TCs play an important role over the eastern US for heavy rainfall events
Figure 4.4: (a) Regional contribution of TC-related rainfall to annual (blue) and TC-season (red) total rainfall. The average contribution of TCs to annual and TC-season total rainfall is given in the top-left corner of each box. (b) The relative size of TC-related rainfall over each region compared to total TC-rainfall over the Southeastern region (100%) by month.

(Shepherd et al., 2007; Knight and Davis, 2009; Barlow, 2011), our results show that their contributions to annual and TC-season total rainfall are small compared to non-TC related rainfall (Table 4.1). Furthermore, there is large inter-annual variability in the TC related rainfall. Except for the SE region, the regional standard deviations in TC-related rainfall are larger than their averages (Table 4.1) and the contribution for active TC years can be considerable. For example, TCs contributed 10% and 17% of annual and TC-season total rainfall, respectively, over the Midwestern region during the most active TC year of 2005.

4.3.3 Climatology of Precipitation and Soil Moisture

In general, the seasonal cycle of precipitation drives the seasonal vegetation and soil moisture dynamics (Eagleson, 1978) and is modulated by the seasonal cycle of solar radiation through atmosphere-land interactions that drive evapotranspiration (Yeh
Table 4.1: Climatology for annual total rainfall, non-TC related rainfall, and TC-related rainfall and their standard deviations for the period 1980-2007. The unit is mm.

<table>
<thead>
<tr>
<th>Area</th>
<th>Annual Mean</th>
<th>Std</th>
<th>No-TC related Mean</th>
<th>Std</th>
<th>TC-related Mean</th>
<th>Std</th>
</tr>
</thead>
<tbody>
<tr>
<td>MW</td>
<td>394</td>
<td>977</td>
<td>104</td>
<td>968</td>
<td>105</td>
<td>9</td>
</tr>
<tr>
<td>NE</td>
<td>387</td>
<td>1148</td>
<td>141</td>
<td>1122</td>
<td>134</td>
<td>26</td>
</tr>
<tr>
<td>S</td>
<td>389</td>
<td>1395</td>
<td>174</td>
<td>1355</td>
<td>182</td>
<td>40</td>
</tr>
<tr>
<td>SE</td>
<td>475</td>
<td>1319</td>
<td>152</td>
<td>1258</td>
<td>157</td>
<td>61</td>
</tr>
</tbody>
</table>

et al., 1998; Dai et al., 2004; Amenu et al., 2005). Our results indicate that the study regions have a seasonal cycle of soil moisture that has a peak during spring and then decrease until late summer and early fall due to high evaporative demand (Figure 4.5, (b)), while they have diverse seasonality of precipitation (Figure 4.5, (a)), that is, the sub-tropical region (30-33°N) has two peak wet seasons (January to March and June to August) but the mid-latitude region has a single peak wet season from April to August. Although TC-related rainfall is small (< 5% of annual total), the TC season is generally coincident with the driest soil moisture conditions and thus is likely to have a significant impact. Furthermore, the high inter-annual variability in TC activity means that their influence will also vary greatly.

4.3.4 Impact of TCs on Drought

We demonstrate the impact of TCs on drought by showing results for Hurricane Katrina in 2005 and its impact on drought at local scales by a comparison between the EXP-TC and EXP-NOTC simulations. In Figure 4.6, the shaded regions indicate the spatial extent of drought on 31st August 2005.

Here, drought was defined in terms of soil moisture lower than the 20th percentile of monthly average soil moisture from the EXP-TC simulation. Along the path of the hurricane, three points (Point 1, Point 2, and Point 3) were selected to examine its...
maximum impact on local scale droughts as it tracked inland. Each point experienced a different impact on drought: 1) late drought initiation, 2) weak drought persistence, and 3) early drought recovery, which is shown in Figure 4.7. The time period of the drought analysis is the same for each point from 1st May 2005 to 31st December 2005. Point 1 (89.6°W 31.1°N; Figure 4.7 (a)) had a large peak in total soil moisture on 29th August due to the intense rainfall from Katrina (> 100 mm). This caused soil moisture to take 11 days longer to reach the threshold value for drought in the EXP-TC simulation than from the EXP-NOTC simulation. The soil moisture time series from both simulations merged together in mid-November because of losses to evapotranspiration during the summer and fall, and lack of rainfall after Katrina. Winter rainfall resulted in drought recovery in both simulations. Point 2 (88.0°W 35.6°N; Figure 4.7 (b)) had multiple TCs in 2005. Katrina brought the second most intense rainfall (> 45mm) out of all TC events that was much lower than at Point 1 because Point 2 was further inland. Tropical Storm Arlene (8th - 13th June) introduced the most intense rainfall (> 50mm) to Point 2. During May, low non-TC
related rainfall led to large decrease in soil moisture with drought initiation in mid-May. Without the contribution of TCs, the drought did not recover until winter with a total duration of drought of about seven months. With multiple TCs the drought initiated three months later and terminated 14 days earlier and therefore Point 2 in the EXP-TC simulation experienced only a three month drought. At point 3 (40.0°N 82.9°W; Figure 4.7 (c)), low non-TC rainfall initiated drought conditions in mid-July before Hurricane Katrina alleviated drought in late August by one month earlier than in the EXP-NOTC simulation. That is, heavy rainfall from the TC can terminate the drought immediately while non-TC related rainfall accumulates slowly and thus local drought recovers later.

The average duration of drought per event is shown in Figure 4.8 as the distribution of values over each region. We show results for six types of drought during
two periods, 2004-2005 and 1980-2007. Due to the large inter-annual variability of Atlantic TCs as shown above, we focused on the most active TC years (2004-2005) to show the maximum impact on drought due to TC frequency alone. Rainfall from Atlantic TCs sometimes affects soil moisture until early spring of the next year while the TC season usually starts in late-May and ends in early-November (not shown). Hence, our drought analysis period for 2004-2005 was from 1st May 2004 to 30th April 2006. For each region, the average duration of drought per event (Eqn. ??) for 1980-2007 was used as a reference to evaluate the impact of TCs during the most active TC years, 2004-2005. For moderately severe, short-term (long-term) droughts, the medians of the average durations derived from Eqn. 4.2 were 53 (119), 54 (120), 56 (122), and 52 (122.5) days over the SE, South, NE, MW regions, respectively, in the EXP-TC simulation for the period 1980-2007. Over all regions, the median drought duration was reduced by two weeks for increasing drought magnitude from moderately severe to extremely severe drought. In the EXP-NOTC simulation, however,
the MW and South regions for 2004-2005 had longer median values of the average durations of long-term droughts than those of their regional climatology. Comparison of the EXP-TC and EXP-NOTC simulation showed that Atlantic TCs in 2004-2005 reduced the duration of moderately severe, long-term droughts over the Midwestern and South regions by one month, respectively. However, their impact on the Southeastern and Northeastern droughts was still minor since enough rainfall was provided during the warm season from other weather systems including squall lines and mesoscale convective complexes (Knight and Davis, 2009).

Figure 4.8: Regional distributions for average duration of drought per event during 2004-2005 (first two bars) and 1980-2007 (last two bars) for six different types of drought and four regions. Black bars represent the EXP-TC simulation ($D_{ave,EXP-TC}$) and green are for the EXP-NOTC simulation ($D_{ave,EXP-NOTC}$). Each column represents a different type of drought and each row represents a different region. The horizontal line represents the median and the box represents the inter-quartile range (third quartile, $q_3$, minus first quartile, $q_1$) over the region. The vertical line represents the range between $q_1 - 1.5 \times (q_3 - q_1)$ and $+1.5 \times (q_3 - q_1)$.

Given a drought in the EXP-TC simulation, its duration in the EXP-NOTC simu-
lation will be extended due to the lack of TC-related rainfall. The extended duration of drought in the EXP-NOTC simulation is partitioned into early drought initiation and late drought recovery. Comparison of the regional distributions of the average late drought initiation and average early drought recovery induced by TCs (Figure 4.9) shows almost no difference in the regional median values for the Southeastern and Northeastern regions between 2004-2005 and 1980-2007. The Southeastern and Northeastern regions had a slightly higher median value of late drought initiation days than the median of early drought recovery days in 2004-2005, while TCs during 1980-2007 give a higher median value of early drought recovery than late drought initiation. This implied that TCs can play different roles in drought initiation and recovery over a region because of the large inter-annual variability in the number and timing of TCs. Over the South region, the spread of the distribution for average late drought initiation days for 2004-2005 increased as the drought threshold increases, whereas the distribution for average early drought recovery did not change its shape regardless of the threshold. In other words, TCs caused the initial seasonal drying of the daily soil moisture in the South US to be shifted later.

Our results indicated that more than 50% of the NE, ME, and South regions experienced short-term drought in both simulations during 2004-2005 (Figure 4.10). However, TCs forced a decrease in the spatial extent of extremely severe, short-term drought in the Southeastern and Midwestern regions to decrease from 50% to 25% and from 80% to 50%, respectively. This indicates that TC-related rainfall during 2004-2005 plays a critical role in modulating extremely severe drought over these regions. About half of the long-term droughts (>90 days) over all the regions were terminated by rainfall from TCs. For example, there were two or three long-term droughts without TCs and one or two droughts with TCs, depending on the region. A greater fraction of the SE and NE regions experienced late drought initiation rather
Figure 4.9: First two bars (last two bars) represent the difference between the EXP-TC and EXP-NOTC simulations for 2004-2005 (1980-2007) for drought initiation ($T_{late}$; green bars) and drought recovery ($T_{early}$; black bars) times. Each column represents a different type of drought and each row represents a different region. The box and whiskers are as in Figure 4.8.

than early drought recovery while 40% of the South region experienced early drought recovery than late drought initiation. The Midwestern region had an equal benefit of late drought initiation and early drought recovery of extremely severe, short-term drought from TCs during 2004-2005.

During the period 1980-2007, the average numbers of short-term and long-term droughts over all four regions were 17 and 6 (22 and 7.5), 15 and 3 (16 and 4), and 11 and 1.7 (12 and 2.6) for moderately severe, short-term and long-term drought, respectively, in the EXP-TC (EXP-NOTC) simulation (Figure 4.11). Over the Mid-
Figure 4.10: Fractions of grid cells over the four study regions that experienced drought, late drought initiation, and early drought recovery at least once in 2004-2005 from the EXP-TC and EXP-NOTC simulation. $D_{d,\text{EXP-TC}}$ and $D_{\text{ave,EXP-NOTC}}$ represent the areal fractions that experienced drought from the EXP-TC simulation and the EXP-NOTC simulations, respectively. $T^{\text{INI}}$ and $T^{\text{REC}}$ are the areal fractions that experienced late drought initiation and early drought recovery days, respectively, during 2004-2005.

In the western region, TCs decreased the occurrence of extremely severe, long-term droughts from three to one over 1980-2007. The South region had the same number of events of severe and extremely severe, long-term droughts in both simulations because only two types of drought occurred over the South: either moderately severe or extremely severe, long-term droughts.

Figure 4.12 shows the area fraction of each region that exceeds a certain difference in the total duration of drought ($D_{\text{diff}}$), drought initiation times ($T^{\text{INI}}$), and drought recovery times ($T^{\text{REC}}$) over the period 1980-2007 between the EXP-TC and EXP-NOTC simulations for different types of drought. TCs reduced the duration of
Figure 4.11: Total number of short-term (a) and long-term (b) drought events by region average and types of drought. TC (NOTC) stands for the EXP-TC (EXP-NOTC) simulation for the period 1980-2007.

moderately severe, short-term droughts by 90 days (10%), 130 days (9%), 150 days (14%), and 180 days (16%) over 50% of the MW, NE, South, and SE regions, respectively. The ordering of the regions follows that of the regional contribution of TCs to total rainfall. The percentages in parenthesis are calculated as the difference of total duration of drought between the two VIC simulations relative to EXP-NOTC simulation. For moderately severe, long-term drought, the area of each region that exceeds a certain difference in total drought duration between the two simulations decreased rapidly as the threshold increased with less spatial variation among the regions. For more than 50% of our four study regions, TCs decreased the total duration of moderately severe, long-term drought by more 150 days. For moderately severe, short-term drought, more than half of the Southeastern region had total late drought
initiation days of greater than 30 days and more than 50% of the South region had total early drought recovery days by 42 days.

Figure 4.12: The areal fraction of each region that exceeds certain differences of total drought duration ($D_{diff}$; (a) and (d)), drought initiation times ($T^{INI}$; (b) and (e)) and drought recovery times ($T^{REC}$; (c) and (f)) from the EXP-TC simulation and the EXP-NOTC simulation for 1980-2007.

Late initiation and early recovery of drought can provide economic benefits to agriculture especially if they coincide with key times during the growing season, although TCs can also impart agricultural losses through direct physical damage on crops and flooding. In Figure 4.13, we showed spatial distributions of differences in drought duration ($D_{diff}$; Figure 4.13 (a) and (e)), the area with significant changes in drought durations based on a Wilcoxon signed rank test at the 90% level (Figure 4.13 (b) and (f)), the total late drought initiation durations ($T^{INI}$; Figure 4.13 (c) and
(g)), and total early drought recovery durations \(T^{REC}\); Figure 4.13 (d) and (h)) during the period 1980-2007 for moderately severe, short-term and long-term droughts, respectively. We presented the results only for moderately severe drought, since the spatial patterns are similar for other types of drought (not shown).

Figure 4.13: Spatial distributions of each region that exceeds certain differences of total drought duration \(D_{diff}\); (a) and (e)), the area with significant changes in drought duration based on a Wilcoxon signed rank at the 90% level ((b) and (f)), the total late drought initiation duration \(T^{INI}\); (c) and (g)), and total early drought recovery duration \(T^{REC}\); (d) and (h)) for the period 1980-2007.

During the period 1980-2007, TCs reduced the duration of moderately severe, long-term droughts by more than 150 days (>15% of the total drought duration from the EXP-TC simulation during the period) while their impact on short-term droughts was small (<10%) on a regional basis (Table 4.2). The Wilcoxon signed rank test showed that there were statistically significant reductions in the annual total duration of short-term droughts for the SE region at the 90% significance level. The results for moderately severe, long-term droughts are limited by the small sample size of
coincident TCs and drought events at the grid scale. TCs delayed drought mainly along the Gulf of Mexico and the mid-Atlantic coasts and the South and Northeastern regions have more early recovery days than other regions. There were local minimums for drought initiation around Atlanta, Georgia and for drought recovery along the Appalachian Mountains, respectively. Possible explanations for the location of these minimums included the local minimum of the contribution of TCs over the region (Figure 4.4) and the topographic effect on the tracks of TCs as shown in model simulations by O’Handley and Bosart (1996). Still, understanding the mechanisms of how TCs move across land is a challenge for predicting and forecasting not only tracks of TCs, but drought initiation and recovery as well (Lin et al., 2006).

Table 4.2: Total number of days under the 10th, 15th, and 20th percentile moisture storage for the period 1980-2007. The difference as a percentage of the total duration of drought for EXP-TC simulations is shown in brackets.

<table>
<thead>
<tr>
<th></th>
<th>EXP-TC</th>
<th>EXP-TC - EXP-NOTC</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>&gt; 30days</td>
<td>&gt; 90days</td>
</tr>
<tr>
<td><strong>Southeastern</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>20 pct.</td>
<td>2404</td>
<td>1074</td>
</tr>
<tr>
<td>15 pct.</td>
<td>1831</td>
<td>849</td>
</tr>
<tr>
<td>10 pct.</td>
<td>1290</td>
<td>625</td>
</tr>
<tr>
<td><strong>South</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>20 pct.</td>
<td>2299</td>
<td>974</td>
</tr>
<tr>
<td>15 pct.</td>
<td>1755</td>
<td>784</td>
</tr>
<tr>
<td>10 pct.</td>
<td>1238</td>
<td>572</td>
</tr>
<tr>
<td><strong>Northeastern</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>20 pct.</td>
<td>2192</td>
<td>968</td>
</tr>
<tr>
<td>15 pct.</td>
<td>1670</td>
<td>820</td>
</tr>
<tr>
<td>10 pct.</td>
<td>1202</td>
<td>615</td>
</tr>
<tr>
<td><strong>Midwestern</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>20 pct.</td>
<td>2167</td>
<td>815</td>
</tr>
<tr>
<td>15 pct.</td>
<td>1641</td>
<td>784</td>
</tr>
<tr>
<td>10 pct.</td>
<td>1158</td>
<td>564</td>
</tr>
</tbody>
</table>
We explored whether the SE droughts were more severe when TCs were less frequent by comparing data for 2000-2001 and 2004-2005, which were less and more active TC periods, respectively. The SE region had total TC-related (non-TC related) rainfall of 90 mm (2531 mm) over 2000-2001, and 318 mm (2609 mm) over 2004-2005. In 2000 and 2001, a high pressure anomaly over the central US blocked moisture fluxes from the Gulf of Mexico, reduced moisture convergence over the southeastern US (Liu et al., 2004) and thus non-TC rainfall was below-normal. At the same time, it introduced less cloudiness and more evaporative demand. This period also experienced a lack of TCs (number of tropical storms: 3; number of hurricanes: 2). As a result, severe drought was reported over the Southeastern region and its estimated cost from agriculture losses was 689 million US dollars (State of the Climate: Drought for Summer 2000). What if the TCs of 2004-2005 (a very active TC year) occurred over the SE region during 2000-2001? We compared the average duration of drought per event during 2000-2001 and 2004-2005 (Figure 4.14) to gain an insight into how much more severe droughts were when TCs were less frequent over the region. This showed a large change in average duration. The SE region experienced more severe short-term and long-term drought during 2000-2001 than during 2004-2005 due to less TC related precipitation for 2000-2001 than for 2004-2005.

4.4 Discussion

4.4.1 Uncertainties In Data Sources

There are a number of uncertainties in the approach that we use which warrant discussion. We use a single land surface model, VIC, to assess the impact of TCs from the Atlantic basin on drought over the eastern US. Other models could be
used and would give different values of the statistics of the impact of TCs because of differences in their soil moisture parameterizations. However, the differences are likely to be small because different models produce similar patterns of drought at regional scales when forced by the same meteorological data (Mo, 2008; Sheffield et al., 2012c) and the overall conclusions of this study would likely be the same. Although multi-model comparisons indicate that there are broad scale similarities, there still exist uncertainties from modeled soil moisture itself due to lack of knowledge about the role of soil moisture in the interaction between the land surface and atmosphere even though the capability of VIC to detect drought has been validated during the last two decades. The results are likely sensitive to the depth of the soil column considered, as upper soil layers respond more quickly to rainfall events than lower soil layers. This is pertinent to the choice of model because different models have different soil column depths, which affects the representation of drought dynamics (Wang et al., 2009, 2011), and therefore care must be taken when comparing results between models. We chose the total soil column as this is most representative of agricultural drought
in terms of depth, but note that the impact of TCs on drought is likely higher for shallower soils and lower for deeper soils.

Uncertainties in the NLDAS-2 forcing data have been discussed by several previous studies, especially for heavy rainfall events. Villarini et al. (2011) found that the NLDAS-2 rainfall data have lower rainfall amounts than the Stage IV multi-sensor radar data product. However, both datasets were able to detect TC-related rainfall events and the timing of their peaks for multiple TCs events in 2004. In that study (Villarini et al., 2011), the NLDAS-2 rainfall was found to be smaller than accumulated runoff interpolated from observed streamflow during September 2004 over regions affected by TCs, which could be considered as a lower boundary on accumulated rainfall. One of the reasons for the underestimation of rainfall in NLDAS-2 was gauge undercatch during heavy rainfall events with high winds (Xia et al., 2012). The NLDAS-2 rainfall forcing data was derived mainly from daily CONUS gauge data and these could be corrected with wind speed information using empirical corrections such as (Medlin et al., 2007), although such methods are still uncertain. We therefore regard our results as conservative estimates of the rainfall associated with TCs and their impact on drought initiation and recovery was likely underestimated.

4.4.2 Sensitivity of Regional Drought to TC Frequency and TC rainfall intensity

It is important to understand the sensitivity of drought to changes in TC frequency and TC rainfall intensity, given that TC frequency is highly variable from year-to-year, and in its spatial impacts. This is especially important in the context of potential future changes in TCs due to climate change (Trenberth, 2005; Knutson et al., 2010). We evaluated the impact of changes in TC frequency and TC rainfall intensity on
decreases in drought on a regional basis over 1980-2007. The frequency of TC events over the eastern US ranged from zero to 74 with the maximum occurring over the Florida peninsula. The TC rainfall intensities were 29, 30.8, 20, and 17.1 mm/event for the SE, South, MW, NE regions, respectively during the period. For each grid cell, we computed the Spearman rank correlation to measure the significance in the relationship between the TC metrics and drought duration. For all regions, there was a statistically significant linear relationship between TC frequency and decreases in duration (Figure 4.15 (a); 99% level). Furthermore, the TC rainfall intensity was significantly positively correlated with the decrease in short-term droughts for all regions (Figure 4.15 (b); 99% level). Only the Midwestern region has a significant relationship between TC rainfall intensity and decreased duration of moderately severe, long-term drought (correlation coefficient: 0.43). According to these results, more frequent intense TCs in the future would be expected to ameliorate short-term droughts over all regions and long-term droughts over the Midwestern region.

4.4.3 Past and Potential Future Changes in TC Impacts on Drought

Changes in the impact of TCs on drought over the past few decades are an important issue and especially given potential future climate change (Knutson et al., 2010; Sheffield and Wood, 2007). This is complicated by the multiple competing factors on changes in TC impact, including changes in the frequency of landfalling TCs, their associated tracks and footprint, changes in the intensity of TC rainfall, and changes in drought frequency, timing and severity. It is also important to put this in the context of whether observed changes in TCs and drought are due to natural variability and/or anthropogenic influences, as this can help inform evaluations of climate models and
Figure 4.15: Fractions of grid cells over the study regions that experienced the corresponding differences in drought durations from both simulations ($D_{diff}$) along the frequency of TCs ((a) and (b)) and the intensity of TCs ((c) and (d)), respectively, for 1980-2007. Colors represent the percentage of a region. Each scatter plot represents total number of grid cells and affected grid cells by TCs and the Spearman rank correlation coefficients between differences in drought duration and TC metrics are shown in the text.

assessments of their future projected changes. The attribution of changes in TCs to natural variability (Goldenberg et al., 2001; Elsner et al., 2008; Knutson et al., 2010), anthropogenic causes (Trenberth, 2005), and other factors (Klotzbach, 2006) (e.g. improved observational technology) has been discussed extensively, but there remains considerable uncertainty as to the variability and anthropogenic impacts. Studies on trends over the last several decades have shown ambiguous results with both decreases (Chan and Shi, 1996; Landsea et al., 1996) and increases (Goldenberg et al., 2001; Webster et al., 2005; Klotzbach, 2006; Briggs, 2008) in TC frequency depending on
the study period, data sources, and metrics used. Analyses of the most recent 10-15 years indicate that the frequency and intensity of TCs have remained unchanged worldwide (Knutson et al., 2010) but have been more intense with a large increase in the North Atlantic (Knutson and Tuleya, 2004; Elsner et al., 2008; Knutson et al., 2010).

Whilst this implies that the impact on drought in the eastern US has not changed due to TC frequency alone and especially over the period of this study for which observational data are more reliable and uniform, the other factors may be relevant and warrant further discussion. There appears to have been no trend in TC rainfall over 1980-2007 from our study, although this is masked by the high inter-annual variability in the TC statistics and subject to our interpretation of TC-related rainfall. However, during the last few decades, Atlantic TCs have contributed increasingly to extreme rainfall events over the coastal states along the Gulf of Mexico and the North Atlantic (Knight and Davis, 2009). Based on the HURDAT database, land-falling tropical storms (TS) for the CONUS have become more frequent during May and June since 1995, when Atlantic TCs have been more active during May and June since 1995, when Atlantic TCs have been more active due to the warmer phase of the Atlantic Multi-decadal Oscillation (Trenberth, 2006). Also, the destructiveness of tropical cyclones in the North Atlantic has increased over the past three decades (Emanuel, 2005). Droughts over the U.S. have not changed much over the 20th century (Andreadis and Lettenmaier, 2006), although regional changes are evident. For example, the southwest and parts of the interior west have become drier that has been attributed to warming temperatures, and it is estimated that drought has decreased in frequency, severity and spatial extent over much of the rest of the US, including the eastern US, due to increases in precipitation over the second half of the 20th century (Andreadis and Lettenmaier, 2006). Therefore, although drought appears to have decreased in the region and TCs have remained stationary in their
frequency, it is unclear whether TCs have contributed to these changes in drought because of uncertainty in the data, high inter-annual variability in TCs and the intermittent temporal and spatial nature of TCs.

Although a climate change signal is not detectable in recent years for the North Atlantic, the future risk of TC-related disasters is expected to increase because warmer SSTs are expected to enhance the frequency of more intense TCs (Knutson and Tuleya, 2004; Emanuel, 2005; Trenberth, 2005; Webster et al., 2005; Elsner, 2008; Elsner et al., 2008; Knutson et al., 2010). Our analysis suggests more intense tropical cyclones, delivering more rainfall per event, would reduce the severity of droughts slightly across the region.

Future changes of TC impacts are also dependent on how drought might change. Projected changes in drought over the eastern US indicate an overall increase in drought, although only changes in short-term droughts (4-6 month duration) are statistically significant, because of the large range of projected changes across climate models (Sheffield and Wood, 2008). The attribution of these changes is unclear at present because of the uncertainty across climate models and the complexity of controls on the eastern US climate (Seager et al., 2009), but is partly due to an increase in the evaporative demand from a warmer atmosphere. Furthermore, climate models (Sheffield and Wood, 2008) are generally unable to simulate realistic TC events and their historic frequencies mostly due to their coarse resolution, and so most current climate models such as those that contributed to the recent Climate Model Intercomparison Project phase 5 (CMIP5) (Taylor et al., 2012) are generally not able to give a realistic assessment of the impact of TCs on future drought. However, it is likely that TCs will play a more important role in the future as drought becomes more frequent and severe, irrespective of whether TCs change in frequency and intensity.
4.5 Summary

This study is the first that are aware of to quantify the influence of TCs from the Atlantic basin on drought regime over the eastern U.S., as far as we know. Daily simulated soil moisture from a land surface model were used to show the impact of TC events, which were short-term and localized, on the longer term and broad spatial scale of drought. We argued that the impact of TCs on drought cannot be appreciated properly without evaluation at the scale of the TCs, rather than at coarser scales such as month time step (Maxwell et al., 2012).

We found that the contribution of TCs to total rainfall was small over all four study regions (<6%) for the period 1980-2007, but was moderate (>15% of rainfall during the TC season) for very active TC years. The impact of TCs on drought included shorter drought duration, late drought initiation, early drought recovery and an overall decrease in the spatial extent of drought. This result was expected from our experimental setup, because removal of rainfall, of any type, would increase drought duration, severity and spatial extent. Other types of rainfall, such as convective systems and frontal storms, and even antecedent snowfall, will have (potentially more significant) impacts on drought than TCs. However, by their nature, TCs have a unique and identifiable impact on drought that is important to understand in isolation from other types of rainfall.

At local scale, a single TC (e.g. Hurricane Katrina in 2005) can play very different roles in drought initiation, persistence, and recovery depending on the location, the timing of the TC relative to the drought event, the number of other TCs during the season and the antecedent land surface state. At regional scale the impact of TCs on drought varied depending on the region, drought type, and year-to-year variation in TC activity. TCs decreased the duration of moderately severe, long-term droughts
by more than 150 days (15%) while their impact on short-term droughts was smaller (less than 100 days) on a regional basis. Over our study regions, they removed at least two short-term and one long-term drought events during the period 1980-2007. Also, they impeded drought initiation mainly along the coastline of the Gulf of Mexico and the Atlantic Ocean and advanced drought recovery especially in the South and NE regions.

The linkages between TCs and drought have been poorly understood in the past due to their different temporal and spatial scales. This study used a quantitative approach to highlight the major role of TCs in drought relief at local to regional scales and therefore the possible benefits of what are usually devastating natural hazards to local communities and economies. A pertinent question is how future changes in TCs may affect drought regimes. The findings of Knutson et al. (2010) indicate that multiple climate models predict an increase in the frequency of intense tropical cyclones globally in the future but no increase in their overall number. Based on these results, we speculate that TCs will play a more crucial role in pluvial and drought regimes over the eastern US in the future, both in terms of direct losses to infrastructure, property and agriculture but also societal benefits from drought relief. Whether, the direct benefits of TCs to drought relief will be offset by projected increases in drought frequency and intensity due to warming temperature and changes in non-TC rainfall (Sheffield and Wood, 2008) is unclear.
Chapter 5

Changes in Drought Risk over the Continental United States: the Influence of the Pacific and Atlantic Oceans

5.1 Background

Drought is a naturally occurring climate phenomenon that may persist from one season through several seasons and even multiple years. It is defined as a period of abnormally dry weather sufficiently long enough to cause a serious hydrological imbalance (AMS 2003) and is classified into four main types: meteorological, hydrological, agricultural, and socio-economic droughts as defined by deficits of precipitation, streamflow, soil moisture, and water supply, respectively. Drought is one of the most costly natural hazards, bringing adverse economic, social, and ecological impacts. For example, the 2012 Midwestern U.S. drought caused 12 billion U.S. dollars in damages
mainly from agricultural losses (Henderson and Kauffman, 2012). Human and ecological communities are also exposed to risk from associated hazards including wildfires and heatwaves, because of the favorable environment (high land surface temperature and water deficits), for example during the 2013 winter drought and wildfires over California. To mitigate these adverse effects, implementation and improvement of drought monitoring and forecasting capabilities is crucial (Pozzi et al., 2013).

However, drought is one of the least understood natural hazards because of its diverse origins, with processes driving drought initiation, persistence, and recovery happening at multi temporal (weekly, seasonal and multi-year) and spatial (local, regional and continental) scales. Due to this complexity, drought forecasting at seasonal and longer time scales remains problematic, and better understanding of the mechanisms is necessary to improve forecast skill. In recent years, dynamical forecasting using atmosphere-ocean coupled climate models has shown potential to improve drought seasonal forecasting based on “teleconnections” with ocean states due to the large heat capacity and long-term memory of the ocean (Hoerling and Kumar, 2003; Findell and Delworth, 2010; Yuan and Wood, 2013).

The El-Nino Southern Oscillation (ENSO) provides much of this potential globally. For example, the cold La Niña phase of ENSO induces warm SSTs over the north-central Pacific and abnormal Rossby wave motions at the mid-troposphere via an atmosphere bridge along mid-latitudes. This leads to a northward shift of the ridge of mid-tropospheric geopotential heights and thus a reduction in precipitation over the contiguous U.S. (CONUS) (Alexander et al., 2002). This is a typical drought scenario for the CONUS during La Niña episodes. However, at the decadal scale, the north Pacific and Atlantic modulate these ENSO teleconnections (Enfield et al., 2001; Mo, 2010) via the Pacific Decadal Oscillation (PDO) and Atlantic Multi-Decadal Oscillation (AMO), respectively. For example, Dettinger and McCabe (1999) found
stronger associations of ENSO with inter-annual variability of western U.S. precipitation during the positive phase of the PDO (1951-1980) than during the negative phase of the PDO (1921-1950). Furthermore, McCabe et al. (2004) found that over 1901-1999, a coincident positive phase of the AMO and a negative phase of the PDO were associated with drought over most of the CONUS. Since the late 1990s, the AMO has switched from a negative to a positive phase and the PDO has switched from a positive to a negative phase with implications for changes in drought risk (Figure 5.1). In this paper, we extend these previous analyses using the latest observational updates to 2012, and apply a Bayesian framework for quantifying the uncertainties in the teleconnections and the stationarity of drought risk over the CONUS. The small number of samples for different phases of the AMO and PDO means that the robustness of the teleconnections is inherently uncertain, which has not previously been taken into account when quantifying risk. The Bayesian method proposed in our study enables us to assess the uncertainties of the influence of SSTs on droughts in an objective way.

5.2  Data and Methods

5.2.1  Precipitation Data and Climate Indices

We use precipitation data from the latest version of Climate Research Unit monthly dataset (CRU TS3.21) that has 0.5 degree spatial resolution and covers 1901 to 2012. We use time series of annual PDO and AMO for 1901-2012 taken from the monthly indices of, respectively, the Tokyo Climate Center and the NOAA Earth System Research Laboratory. For ENSO, we use annual values of the Southern Oscillation Index (SOI) from the monthly index of the Bureau of Meteorology, Australian Government,
Figure 5.1: Time series of (a) annual mean global SST warming trend from monthly HadISST version 1.1 over 1901-2012, (b) annual Pacific Decadal Oscillation (PDO), (b) annual Atlantic Multi-Decadal Oscillation (AMO), and (c) annual El Nino Southern Oscillation (ENSO) indices. Solid lines are annual averages and ten-year moving averages for PDO and AMO and five-year moving averages for ENSO, respectively.

which is calculated from the pressure differences between Tahiti and Darwin (SOI (-): El Nio episodes and SOI (+): La Nia episodes).

5.2.2 Uncertainty Analysis Using a Bayesian Approach

We focus on meteorological drought, which is defined as the precipitation amount under a threshold value during a given time. We choose a threshold value as the lower quartile of annual precipitation during 1901-2012 at each grid cell. We treat drought occurrences, X (0 = no occurrence; 1 = drought occurrence), as samples of
a Bernoulli process derived from the time series of precipitation. The frequency of drought occurrence is defined as the fraction of the total number of drought occurrences over all the years. Based on the threshold value, the frequency \( p \) is expected to be 0.25, which translates to 28 drought occurrences over the 112 years.

One of the benefits of representing drought occurrence as a Bernoulli process is that the prior distribution is a conjugate of its likelihood function and thus the posterior distribution is derived from the same family as the prior (Benjamin and Cornell, 1970). We can compute the posterior distribution for the unknown parameter (drought frequency, herein) given a Bernoulli process sample, \( X \), based on the Bayesian inference that the posterior distribution of the unknown parameter is proportional to its likelihood function multiplied by the prior.

\[
Pr\{p|X\} \approx Pr\{X|p\} Pr\{p\} \tag{5.1}
\]

, where \( p \) represents drought frequency.

Previously, studies have used Bayesian methods to analyze drought, but generally in the context of forecast uncertainty (Raje and Mujumdar, 2010; Madadgar and Moradkhani, 2013). If we assume that the prior distribution is a uniform distribution, known as an uninformative prior, which is equivalent to a beta distribution with both parameters (alpha and beta) equal to 1, we can simply compute the posterior distribution from another beta distribution with different parameters, alpha \((s+1)\) and beta \((n-s+1)\), where the number of drought occurrences \( s \) and the number of years \( n \) are from a Bernoulli process sample (Figure 5.2). These two statistics \( s \) and \( n \) are the sufficient statistics of a Bernoulli process sample (Benjamin and Cornell, 1970).
\[
f(p; s, n - s) = \binom{n}{s} p^s (1 - p)^{n-s} \text{constant}
\]
\[
= \frac{\Gamma(n)}{\Gamma(s)\Gamma(n - s)} p^s (1 - p)^{n-s} \text{constant}
\]
\[
= \frac{1}{\beta(s + 1, n - s + 1)} p^s (1 - p)^{n-s}
\]

where \(\Gamma(n)\) is the gamma function and \(\beta(s, n-s)\) is the beta function.

We compute the conditional posterior distribution from a subset of the observational data given a certain phase of the PDO, AMO, and ENSO (\(\Pr\{p | X, Y\}\), where \(Y\) is the phase of SST index, Figure 5.2) in order to examine the impact of the Pacific and Atlantic Oceans on drought occurrence. We can compare the statistics (location and dispersion) of those conditional posterior distributions with those of the posterior distribution from the full observational data. We also quantify the certainty of the impact of the Pacific and Atlantic Ocean states as the probability to have higher drought frequency than expected (0.25) from the conditional posterior distributions (\(\Pr\{p > 0.25 | X, Y\}\)). This probability is expected to be 0.5 from the posterior distribution from all the observational data. We use a probability threshold value of 90% and 95 percent (hereafter, PTV= 90% and PTV = 95%) to identify areas with a more certain increase in drought frequency given a phase of the SST index, which is equivalent to or higher than a 99.9 percent confidence level from the Kolmogorov-Smirnov test with the posterior distribution from climatology (1901-2012).
5.3 Results

5.3.1 The impact of PDO, AMO, and ENSO on annual drought frequencies

We construct the maps for annual drought frequency over the U.S. for each phase of ENSO, AMO, and PDO (1900-2012; Figure 5.3). Red (blue) colored regions have higher (lower) drought frequency than the expected drought frequency of 0.25. More vulnerable regions given negative or positive phases of PDO, AMO, or ENSO are represented by contour lines and hatched areas, for PTV = 90% and PTV = 95%, respectively. The results indicate that a negative phase of the PDO induces higher risk of annual drought frequency than expected over the U.S., while the area under
drought risk ranges from 30% (PTV = 90%) to 10% (PTV = 95%) depending on the probability threshold value. These regions include Kansas, Oklahoma, Texas, Wyoming, Colorado, New Mexico, and the Southern part of California, increasing the drought frequency to more than 0.35 (shorter than a 3 year return period). Given a positive phase of AMO, most of the U.S. is exposed to elevated risk for annual drought, although regions of high certainty are limited. Given a positive phase of the SOI, 19% and 10% of the U.S. are more vulnerable to annual drought at PTV = 90% and 95%, respectively, with drought frequency above 0.35. These regions include Alabama, Florida, Louisiana, Mississippi, Texas, New Mexico, and Arizona. Based on all years of the record (1901-2012), the U.S. has generally been exposed to higher risk for drought occurrence, when the PDO has been in a negative phase, respectively, and a coincident La Niña year brings higher risk, especially in the southwestern U.S.

These spatial features are consistent whether for all or just for strong negative and positive phases (defined as below the 33rd percentile and above the 66th percentile, respectively) of the PDO, AMO, and ENSO (Figure 5.4). Interestingly, the central Great Plains including Kansas, Oklahoma, and Wyoming, have drought frequency close to expected during strong negative phases of the PDO, suggesting that drought risk over these regions is influenced more by weak negative phases of the PDO (Figure 5.3 and 5.4).

5.3.2 Recent changes in U.S. drought risk

We examine the stationarity of Pacific teleconnections and the influence on the risk of annual drought using an 80-year moving window (1901-1980 through 1933-2012), which is long enough to capture both inter-annual and decadal SST variations. Figure 5.5 shows drought risk maps during the negative phase of the PDO and the positive
Figure 5.3: Annual drought frequency map during each phase of PDO ((a): + and (d): -), AMO ((b): + and (c) -), and ENSO ((c): + and (f): -). Contour lines and hatched areas represent the grid cells that have 90% and 95% to have higher drought frequency (p) than 0.25, respectively, from each conditional posterior distribution for drought frequency during 1901-2012.

Phase of the SOI for the two end periods, 1901-1980 and 1933-2012. These periods share negative phases of the PDO during 1933-1980 and thus the differences between the risk maps are related to the different response to the negative phase of the PDO between the early 1900s and early 2000s. Figure 5.4 also shows the time series of the 80-year moving window areal fraction of the U.S. with elevated drought frequency at probability threshold value of 90% and 95% during the negative phases of the PDO and the positive phase of SOI. During 1901-1980, the southern and midwestern U.S. were more vulnerable to annual drought occurrence during negative phases of the PDO (more than 30% and 20% of the U.S. at PTV = 90% and 95%, respectively), whereas the southwestern U.S. (20% and 10% of the U.S. at PTV = 90% and 95%, respectively,
Figure 5.4: Same a Figure 5.3 except for during strong positive (above 66th percentile) and strong negative (below 33th percentile) phases of each annual climate index. Including Arizona, California, and New Mexico) was exposed to higher risk for annual drought during 1933-2012. Droughts over the southwestern and midwestern U.S. were associated with positive phases of SOI (La Ninas) during 1901-1980, but the area susceptible to this level of drought risk shrank and moved slightly northwards (Colorado, Utah, and Wyoming). The time series of the 80-year moving window areal fraction shows that overall the influence of the negative phase of the PDO and the positive phases of the SOI on drought occurrence over the U.S. has weakened.

### 5.4 Discussion and Conclusions

The Bayesian approach allows us to quantify the uncertainties of the impact of SSTs on annual and seasonal droughts over the U.S. in an objective way. Our findings
suggest that a combination of a positive phase of the AMO and a negative phase of the PDO and ENSO is a worst-case scenario for droughts over the southern U.S. Since the late 1990s, the phases of the PDO and AMO have changed from positive to negative, and negative to positive, respectively, which suggests that the southern U.S. will be exposed to higher drought risk until another phase shift occurs. This risk will be modulated at annual and seasonal time scales by the inter-annual variability of ENSO and short-term meteorological events such as atmospheric rivers (Dettinger, 2013) and tropical cyclones (Chapter 4).

During the last century, ENSO teleconnections with U.S. summer droughts have been non-stationary (McCabe and Dettinger, 1999). Here, analysis of recent observational data shows that associations of ENSO and PDO with U.S. drought have
weakened over time, although an increased influence is apparent for some regions such as the southwest. This weakening is consistent with the attribution of recent drought events in the U.S. to atmospheric variability (Kumar et al., 2013; Seager et al., 2014; Wang et al., 2014) and suggests that droughts are becoming less associated with oceanic variability. Whether this has contributed to overall changes in drought is unclear, in part because long-term trends are regionally and epoch dependent. Long-term analyses for the southern U.S. (Chen et al., 2012) and the U.S. (Andreadis and Lettenmaier, 2006; Sheffield et al., 2012b) show that little change or even a decrease in drought, which may be related to the weakening of oceanic controls. On the other hand, the increasing influence of the negative phase of the PDO in the southwest may be linked to the reported increased in drought over the past three decades in this region (Damberg and AghaKouchak, 2014).

These results also have implications for seasonal predictability. The non-stationarity in observed teleconnections and the possible switch to greater atmospheric control may partly explain the limited skill of seasonal climate forecast models (Chapter 6), which tend to have stronger coupling between SSTs and U.S. hydroclimate than the observational data, especially during summer and for the Midwestern U.S. Better understanding of non-stationarity in the impact of SSTs on drought occurrence over the U.S. can better help translate the information from SSTs into improving seasonal and annual drought forecasts.
Chapter 6

Understanding the Role of Sea Surface Temperatures in Forecasting of the Midwestern U.S. Summer Droughts

6.1 Background

With improved understanding of ocean-atmosphere-land interactions and representation of such interactions or teleconnections in climate models, seasonal forecasting of drought using coupled climate models has shown potential for forecasting severe drought several months in advance (Luo and Wood, 2007; Yuan et al., 2013). The rationale for this is that the oceans can serve as a source of seasonal predictability because of its long memory, and the fact that drought onset usually takes several months to emerge. For example, the El Nino-Southern Oscillation (ENSO), which is the major driver of climate variations globally, is now predictable at seasonal and
perhaps longer time scales depending on the initial month (e.g. after the spring predictability barrier) (Webster and Yang, 1992). Observations and climate models suggest that variability in the Pacific and Atlantic ocean conditions are associated with drought in the Midwestern US (MW, 100-85°W and 37-46°N) at inter-annual (Ting and Wang, 1997; Wang et al., 2007; Hu and Feng, 2012) and decadal scales (Hoerling and Kumar, 2003; McCabe et al., 2004).

The MW region of the US is susceptible to severe drought due to its highly variable climate than spans semi-arid to continental humid climate zones. Persistent and/or spatially extensive droughts over the MW have plagued the region during the 1930s, 1950s, and 1988, each with different characteristics. The long-term Dust Bowl drought of the 1930s is thought to have been forced by cold SSTs over the tropical Pacific and warm SSTs over the North Atlantic, which induced anomalous large scale circulation patterns, e.g., strong positive height anomalies in the upper troposphere in extra-tropical regions and upper-level anticyclone anomalies, respectively (Schubert et al., 2004b; Bronnimann et al., 2009) and thus hindered moisture supply from the Gulf of Mexico. The Dust Bowl drought was exacerbated by high surface temperatures and increased evaporative demand, as well as poor agricultural practices that resulted in soil erosion and large dust storms (Cook et al., 2007).

During the 1950s, another major drought persisted over much of the southern and central US for several years, with Texas the worse hit and declared a federal drought disaster area (Sheffield and Wood, 2011). Based on the Palmer Drought Severity Index, moderate to extreme drought extended up to about 58% of the contiguous US (CONUS) including the MW region in December 1956 (NCDC, 2012). This drought was associated with low-tropospheric anti-cyclonic anomalies over the mid-Pacific and a cold phase over the northeastern Pacific (Namias, 1982).

The 1988 MW summer drought was relatively short-term but was comparable to
the 1930s drought in terms of severity and spatial extent (Schubert et al., 2004a), with the worst agriculture losses on record in the US of 39 billion U.S. dollars, mainly because its location and timing coincided with the major crop growing region and season of the US (Sheffield and Wood, 2011). Below average precipitation initiated drought over the MW during the early summer of 1988, and an anomalous high pressure system maintained the drought through the summer, contributing to high surface temperature and high evaporative demand (Sheffield and Wood, 2011). The La Niña phase of ENSO, the cold phase of the Pacific Decadal Oscillation (PDO) and the warm phase of the Atlantic Multi-decadal Oscillation (AMO), conspired to form a favorable environment for the drought (McCabe et al., 2004).

The latest drought to hit the MW region, in 2012, was driven by the least summer precipitation for the last three decades, with over 70% of the CONUS experiencing below-normal summer precipitation (Figure 6.1), which is more severe than the 1988 drought (not shown). Total agricultural losses were over 20 billion U.S. dollars based on crop insurance losses (Booton, 2012). In the summer of 2012, the north Pacific and north Atlantic were in a cold and warm phase, respectively, which was also observed in the 1930s and 1988 droughts. The tropical Pacific was, however, in a weak warm phase, differing from the strong cool phase in the 1930s and 1988. The role of these anomalous SST conditions was therefore ambiguous in weakening of CONUS precipitation since ENSO was out of phase with the PDO (Hu and Huang, 2009) and AMO (Hu and Feng, 2012) during 2012. Two recent studies suggest that the 2012 drought was driven by atmospheric noise. A modeling study (Kumar et al., 2013) found that atmospheric variability alone can reproduce the 2012 MW drought, and Hoerling et al. (2013) found that the drought was caused by a reduction in atmospheric moisture transported from the Gulf of Mexico due to mid-troposphere anticyclonic patterns over the Atlantic.
Recently, the National Multi-Model Ensemble (NMME) project (Kirtman et al., 2014), a multi-institutional collaborative seasonal forecasting system, was developed with the goal of improving seasonal prediction of extreme events over the CONUS and globally. Phase I of the project includes seven seasonal forecast models (Kirtman et al., 2014). Here we examine forecasts from six of the models: NCEP-CFSv2 (CFSv2), COLA-RSMAS-CCSM3 (COLA), GFDL-CM2.1 (GFDL), IRI-ECHAM4.5-AnomalyCoupled (IRI-AC), IRI-ECHAM4.5-DirectedCoupled (IRI-DC), and NASA-GMAO (NASA). In 2012, all the NMME models predicted negative precipitation anomalies over the MW. They also predicted a cold phase of the eastern Pacific and a
warm phase of the North Atlantic, which are associated with more frequent drought occurrence over the CONUS (McCabe et al., 2004), and a cold phase of ENSO during the antecedent winter, which is also a favorable condition for meteorological drought in the central US during the summer (Wang et al., 2007). However, the magnitude and spatial distribution of the predicted anomalies over the CONUS varied widely (Figure 1). The most skillful model was the Geophysical Fluid Dynamics Laboratory (GFDL) model, which forecasted the drought with comparable magnitude to the observational data, and showed the most skillful prediction for SSTs. This raises the following questions: 1) Does the apparent skill for SSTs translate to other historical drought events? In other words, what is the level of potential predictability? 2) How well do the forecast models represent the observed association of conditions in the north Pacific and Atlantic with MW precipitation (actual predictability)? 3) Are the role of the Pacific and Atlantic in driving MW summer drought robust over time?

To answer these questions, we conduct a singular value decomposition (SVD) analysis of the cross-covariance matrices between northern hemisphere summertime global SSTs and CONUS precipitation for 1982-2012. We do this using observational data and the six NMME climate forecast models to assess and compare Pacific and Atlantic SSTs teleconnections with CONUS precipitation. We focus on meteorological drought, which is a precursor to agricultural drought and is defined as a precipitation deficit over a given time period. We also carry out an SVD analysis for earlier periods (1922-1951, and 1952-1981) to examine whether the observed teleconnections are stationary. This is because the influence of the Pacific and Atlantic oceans on CONUS precipitation has been documented to have changed in early 1950s (McCabe and Dettinger, 1999; McCabe et al., 2004) and there is evidence that the characteristics of ENSO have changed since the 1970s with less frequent but stronger events (An and Jin, 2000), which could have modulated teleconnections with CONUS hydroclimate.
6.2 Data and Methods

Observed precipitation and SST data are taken from the Climate Prediction Center (CPC) Unified Gauge-Based Analysis (0.5-degree resolution, 1948-present) (Chen et al., 2008) and the Extended Reconstructed Sea Surface Temperature v3b (ERSST v3b; 2.0-degree, 1854-present) (Smith et al., 2008), respectively. We focus on northern hemisphere summer averages (JJA) since this is coincident with the growing season and high water resources demand, and focus on 1982-2012. Furthermore, the summertime is characterized by poor predictive skill (Wang et al., 2007). The ensemble means of the first three month (JJA) precipitation forecast initialized on different dates depending on the model (5th day of every June: CFSv2 and NASA-GMAO; 1st day of June: COLA, GFDL, IRI-AC, and IRI-DC) are used because the zero month lead time forecast has the best seasonal skill (Zhang, 2011).

We perform an SVD analysis (Wallace et al., 1992; Bretherton et al., 1992) of the cross-covariance matrix between the CPC precipitation and ERSST SSTs at the 4369 (CONUS) and 10855 (global oceans) grid points for the period 1982-2012. The number of grid points for the six NMME forecast models are 1456 for the CONUS precipitation but vary for the SSTs from 33891 (COLA) to 65341 (IRI-AC and IRI-DC) due to different spatial coverage. Before the SVD analysis, linear trends, reflecting global warming, in the SST and precipitation data are removed. From the SVD analysis, we compute the loadings and the scores for each mode for the left and right fields (sea surface temperature and precipitation, respectively). The loadings of the SVD modes from the SST (SVD SST) and the precipitation (SVD PREC) fields are computed from the left and right eigenvectors, respectively, and represent spatial patterns for the corresponding mode. The scores of the SVD modes are computed by projecting the data on to the loadings of the corresponding field and often are used
to define a climate index (e.g. PDO index). One of the benefits from the SVD is that the order of the SVD modes is based on the strength of the couplings, that is, the first mode is the strongest and the last mode, the weakest. Here, we produce the scores of the SVD modes from SSTs (SVD SST modes) and precipitation (SVD PREC modes). We also construct homogeneous and heterogeneous correlation maps (Wallace et al., 1992) for each mode, which are maps of the temporal correlation between the scores of the SVD SST mode and gridded global SSTs (homogeneous correlation) and between the scores of the SVD SST mode and gridded CONUS precipitation (heterogeneous correlation). The statistical significance of the correlation is calculated at a significance level of 0.1. These maps are useful to examine the association of global SSTs with CONUS precipitation.

We explore the multi-decadal variations in the strength of coupling of global SSTs with CONUS summer precipitation using another observational dataset: the University of Washington Extended (UW-Extended) dataset (0.5-degree, 1915-2008) (Andreadis and Lettenmaier, 2006), because the CPC dataset starts in 1979. It should be noted that the ERSSTv3b SST data in the first half of the 20th century have larger uncertainties due to heterogeneities and sparse sampling, which may affect the results. This dataset has been shown to be consistent with other similar SST datasets (Yasunaka and Hanawa, 2008) for ENSO after 1880 and but only after 1950 for the PDO. Over the MW region, the UW-Extended dataset is consistent with the CPC dataset for their overlap period (1982-2008) with temporal correlation coefficient $> 0.9$. The SVD analysis is carried out separately for three different periods: 1922-1951, 1952-1981, and 1982-2008.


6.3 Results

6.3.1 Covariability between global SSTs and JJA CONUS precipitation in observations and NMME models

The covariability explained by the three leading SVD modes between the observed global SSTs and CONUS precipitation for 1982-2012 as represented by the squared covariance fraction (SCF) values are 12%, 11.6%, and 7%, respectively (Table 6.1). For the models, the first SVD mode, S1, explains 20-32% of the covariability (SCF), which is twice to three times as high as for the observational data.

Table 6.1: Squared covariance fractions (SCFs) and temporal correlation coefficients of the three leading SVD modes from global sea surface temperatures and contiguous United States precipitation from observational data and six climate forecast NMME models. Bold represents the pan-Pacific ENSO-like SVD modes from observational data and climate forecast NMME models.

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<td>SCFs</td>
<td>7.0</td>
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<td>S3</td>
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The observational data indicate that the first SVD SST mode homogeneous correlation map shows a positive correlation with tropical Pacific SSTs and a negative correlation with north central Pacific SSTs, which is well-known as the pan-Pacific ENSO-like pattern (Figure 6.2). This pattern is related to inter-annual (e.g., ENSO) and decadal (e.g. PDO) variability (Schubert et al., 2009). The pan-Pacific ENSO-like SST mode is also identified in the data for the CFSv2, GFDL-CM2.1 and NASA-
GMAO models for the first mode, and the COLA, IRI-AC, and IRI-DC models for the second mode.

The heterogeneous correlation map of the pan-Pacific ENSO-like SVD SST mode from the observational data shows a positively weak correlation with precipitation over the southern and northwestern regions of the CONUS (right, first row of Figure 6.2). This suggests that a warm phase of ENSO and the cold phase of PDO are linked to more precipitation over some parts of CONUS, although these teleconnections are generally weak for this period, 1982-2012. However, the correlation maps for the NMME models indicate either much stronger or much weaker teleconnections than those in the observations with model-specific spatial patterns, while all of them are of the correct sign. The CFSv2 pan-Pacific ENSO-like SVD SST mode is strongly positively correlated with its forecasted precipitation over the mid-western and central US. The pan-Pacific ENSO-like mode from COLA is weakly associated with variations in CONUS precipitation since the mode explains only 8% of SCF (Table 6.1). For the GFDL model, the pan-Pacific ENSO-like mode plays a major role in inter-annual variations of summer precipitation over the MW explaining 37% of the variances (the regional average of the square of temporal correlation coefficients from the heterogeneous map). The IRI-AC and IRI-DC pan-Pacific ENSO-like SST modes are positively correlated with their respective forecasted precipitation across the CONUS. Finally, the NASA-GMAO model pan-Pacific ENSO-like SST mode relates to precipitation over the central US and from the southeast through to the northwest. For the MW region, the pan-Pacific ENSO-like SVD mode from observations explains only 4% of summer precipitation variances for 1982-2012 while the same mode from the climate models explain 19% (CFSv2), 31% (COLA), 37% (GFDL), 11% (IRI-AC), 19% (IRI-DC) and 17% (NASA) of their forecasted summer precipitation variances.

We also carried out an SVD analysis for spring (Figure 6.3) to see whether the
models were able to reproduce prior conditions. During spring, the observational data show a weak coupling pattern between global SSTs and CONUS precipitation. However, the NMME models show a much stronger coupling. This is similar to the SVD results for summer, with the model coupling much stronger. The spring SVD analysis of observations confirms that SSTs were not a dominant driver for the 2012 drought, but more than half of the NMME models forecasted the 2012 drought for the wrong reason: a strong coupling with SSTs. The models also predicted the drought to start earlier in the spring than observations. MAM spatial precipitation anomalies initialized in February (one month lead time) and skill maps for MAM computed against the observational data (NCEP Climate Prediction Center, 2014), show that
most of the NMME models forecasted a severe drought over the CONUS, but with poor skill relative to observations. The onset of the 2012 drought actually occurred in mid-May, and therefore the strong coupling with SSTs also hampered the forecast of the onset of the drought.

6.3.2 Current level of forecast skill for MW drought

To assess actual predictability for MW drought, we plot a Taylor diagram for observed and forecasted MW precipitation anomalies for 1982-2012 (Figure 6.4 (a)). The standard deviations are derived from observed and forecasted MW summertime precipitation anomalies. Overall, the climate forecast models show generally poor predictive skill for MW summer precipitation, with the best performing model,
NASA-GMAO, having a temporal correlation coefficient slightly over 0.5. All models but one underestimate the inter-annual variability of observed MW precipitation (standard deviation of observed precipitation is 0.6 mm/day) by 30-60%. In particular, the variability from the IRI-AC and IRI-DC models is only 30% of the observed value. Only the GFDL overestimates the inter-annual variability of MW precipitation by about 30% (standard deviation = 0.8 mm/day).

Figure 6.4: Taylor diagrams for time series of observed and forecasted precipitation over the MW regions of US (left) and time series of observed (forecasted) MW precipitation anomalies and the loadings of the pan-Pacific ENSO-like SVD mode for precipitation from observational data (climate forecast models). The standard deviations are computed from time series of observed and forecasted MW precipitation over the 1982-2012. Taylor diagrams for time series of observed and forecasted precipitation over the MW regions of US (left) and time series of observed (forecasted) MW precipitation anomalies and the loadings of the pan-Pacific ENSO-like SVD mode for precipitation from observational data (climate forecast models). The standard deviations are computed from time series of observed and forecasted MW precipitation over the 1982-2012.

To understand the importance of pan-Pacific conditions in generating MW summer precipitation, we plot another Taylor diagram from the time series of MW precipitation and the loadings of the pan-Pacific ENSO-like SVD PREC mode from the observational and model data (Figure 6.4 (b)). The loadings are derived by projecting
the MW precipitation anomalies onto the mode. The loadings indicate strong coupling (temporal correlation coefficients of 0.76-0.85) but not necessarily in the same order as the SVD mode number (Table 6.1), which is not guaranteed using SVD. The standard deviations are used in the same way as for Figure 6.4a since here we use normalized time series of the loadings of the pan-Pacific ENSO-like SVD PREC mode. Some climate models overestimate the role of this mode in producing MW summer precipitation while other models show comparable strength of the coupling between their forecasted MW summer precipitation and the pan-Pacific ENSO-like SVD PREC mode. CFSv2 and GFDL show strong coupling with temporal correlation coefficients greater than 0.8. These strong temporal and spatial couplings might explain why the GFDL model shows stronger precipitation anomalies over the MW region than other models depending on the strength of the pan-Pacific ENSO-like SVD PREC mode, regardless of the coincidence with actual extreme events over the CONUS over the last three decades. The 1983 and 1988 drought events are good examples of this. CFSv2 and GFDL fail to forecast the 1983 drought showing a strong positive signal in the loadings of the pan Pacific ENSO-like SVD PREC mode, while they succeed to predict the 1988 drought by showing the opposite signal (not shown). In fact, the cause of 1983 drought has been explained by the uniqueness of a combination between a very strong negative phase over the north Pacific and a warm phase over the tropical Pacific (Ting and Wang, 1997).

6.3.3 Changing strength of pan-Pacific teleconnections (1915-2008)

The role of pan-Pacific SSTs in modulating CONUS precipitation has been variable over time. For example, McCabe and Dettinger (1999) found that western US
precipitation was weakly related to ENSO during 1920-1950 compared to 1950-80. Wang et al. (2007) demonstrated that the La-Nina phase of ENSO in recent years forces a continental-scale persistent high pressure anomaly during the summer which is related to central US drought. Also, SSTs over the north Pacific contribute to inter-annual summer precipitation over the Central US with a comparable magnitude with SSTs over the tropical Pacific (Ting and Wang, 1997). The North Atlantic driven atmospheric circulation is modified by ENSO (Hu and Feng, 2012). Over the recent decades, the characteristics of ENSO have changed (An and Jin, 2000), the phase of PDO has shifted to the negative phase, and a warm phase of AMO has been persistent. The findings from these previous studies and the current study raise the following question: how have pan-Pacific teleconnections with MW summer precipitation changed over decadal time scale and has this impacted the skill of the NMME models?

To answer this question, we repeat the SVD analysis for three different periods, 1922-1951, 1952-1981, and 1982-2008 (Figure 6.5 (a), (b), and (c), respectively), between the ERSST v3b SSTs and the UW-Extended precipitation dataset. Again, the datasets are detrended over each period. During 1922-1951, the first SVD mode relates to regions of the tropical and north Pacific (top, Figure 6.5 (a)) and to MW precipitation (bottom, Figure 6.5 (a)). The sign of the coupling is different to that from the SVD analysis for 1982-2012 (top, Figure 6.2), indicating that the La Nia state of the tropical Pacific and the cold phase of PDO are related to above-normal MW summer precipitation. During 1952-1981, the homogeneous correlation map of the first SVD mode from the SST data shows the opposite sign to 1922-1951, while it is still related to MW summer precipitation. This suggests that the sign of pan-Pacific ENSO-like teleconnections changed in the early 1950s (top, Figure 6.5 (b)). During 1982-2008, the pan-Pacific ENSO-like pattern is the first SVD SST mode (top, Figure
6.5 (c)), however its influence is only on northwestern US summertime precipitation. The sign of the homogeneous and heterogeneous maps from the first SVD SST mode are consistent with the sign of the SVD analysis for 1982-2012. These results reaffirm that the role of global SSTs in inducing summer precipitation over the CONUS has varied with different coupled spatial patterns over the last 100 years (McCabe and Dettinger, 1999; McCabe et al., 2004; Hu and Huang, 2009; Hu and Feng, 2012).

Figure 6.5: Same as Figure 6.2 except for three different periods: (a) 1922-1951, (b) 1952-1981, and (c) 1982-2008.

6.4 Summary

The observational data indicate that there is a great deal of variability in the mechanisms for reductions in MW summer precipitation. Previous studies have shown that La Nia is the leading driver of drought in the MW, especially in the winter to spring, but the strength of this teleconnection in the summer has weakened and the precipitation footprint has shifted over the past decades. However, most of the NMME climate models underestimate the MW summer precipitation variances, leading to predictions of relatively normal precipitation. The GFDL model stands out as having strong coupling between SSTs and precipitation, and thus overestimates
MW summer precipitation variances, and generally predicts either extreme dry or wet anomalies more often than those from the observational data (e.g. droughts of the severity of 2012 are predicted six times during 1982-2012). Therefore, the models are limited in their ability to predict MW precipitation anomalies even at zero month lead time, despite their skill in predicting pan-Pacific SST variability at seasonal or longer time scales. Because Pacific SSTs did not play a major role in the 2012 drought (Kumar et al., 2013; Hoerling et al., 2014), and droughts in general over the last 30 years, the skillful prediction of Pacific SSTs may actually hinder the prediction of MW summer precipitation due to the underestimation of the variances in the most models and overly strong coupling in two models. The prediction of the 2012 drought by the NMME models was in fact fortuitous due to the erroneous coupling with pan-Pacific SSTs.

This study shows that the role of pan-Pacific ENSO-like variability on CONUS precipitation has changed over the last century with teleconnections weakening over the past 30 years. The attribution of these changes is still unclear: whether it derives from either multi-decadal variation of SSTs or changes in large scale circulation and SSTs forced by anthropogenic impacts. Since the late 1970s, ENSO has been modulated by the interdecadal climate shift (Wang, 1995). However, global warming has contributed to the modulation of ENSO due to the “delayed response” from deeper ocean temperatures in the tropical eastern Pacific (S. et al., 2008). Also, there has been a shift to more frequent central Pacific type ENSO events in recent years which have contributed to the weakening (strengthening) the impact on the MW (southwestern) US precipitation at seasonal scale (Mo, 2010). Due to poor understanding of the attribution of these changes in ENSO characteristics and their impact on precipitation over the CONUS, seasonal forecast skill based on SSTs remains limited.

As mentioned above, forcing from the Pacific does not matter for the 2012 drought
and for most previous droughts over the past 30 years. However, SSTs are important from the model perspective and will sometimes fortuitously give rise to skillful predictions. It should be noted, however, that droughts before the 1980s do show a connection with Pacific SSTs from observations (Figure 6.5), suggesting that model skill may be improved before the 1980s. Confirmation of this would require hindcasts back to the early part of the 20th century. Addressing these issues is critical for improving the current state of predictive skill for seasonal drought not only for the MW, but globally as well.
Chapter 7

Conclusions

7.1 Summary and Conclusions

This thesis addresses U.S. drought risk and predictability by providing a comprehensive understanding of U.S. drought mechanisms across different temporal and spatial scales. It finds that drought risk over the U.S. varies depending on the region. It is also determined that the current predictability of U.S. drought is limited at seasonal scales, resulting from an insufficient understanding of changes in the drought risk.

Chapter 2 shows that the low flow regime over the eastern U.S. has been changed with a north-south (increasing-decreasing) dipole pattern, which mainly follows the spatial patterns of the trends in antecedent precipitation. However, the mid-Atlantic region shows a strong consensus of trends between low flows and potential evapotranspiration, while it shows a weak consensus of trends between low flows and antecedent precipitation. Over the southeastern U.S., inter-annual variability of low flows is associated with large-scale atmospheric circulations, showing a negative correlation with the Pacific North-Atlantic (PNA) pattern and a positive correlation with the North Atlantic Oscillation (NAO) pattern at two-month and one-month lead time steps,
respectively. It informs us of the elevated low flow drought risk over the southeastern U.S. during positive PNA and negative NAO in summer, compounding with the long-term decreasing trends. Their concurrence can cause severe adverse effects on society (e.g. food and electricity production, ecological impact, public health). Other potential drivers (e.g. groundwater withdrawal and land cover changes) for the decreasing trends in low flows over the mid-Atlantic region need to be studied for better water resources management and policy.

In Chapter 3, the generating mechanisms for the southeastern U.S. droughts and pluvials are studied, which previously have not been fully understood due to ample water resources and a partially complex climate system. A recently published regional reanalysis product (NARR) enables this study to establish the favorable terrestrial, atmospheric and oceanic conditions for the southeastern U.S. droughts and pluvials. A combination of these favorable conditions warns water resources managers and stakeholders of the elevated risk of hydroclimate extreme events. In the conclusion of this chapter, drought risk over the southeastern U.S. is related to changes in large-scale atmospheric circulation, which intensifies precipitation over the region, induced by global warming. It suggests that a further study for anthropogenic impact on the southeastern U.S. climate regime and hydroclimate extremes is necessary.

In Chapter 4, the contribution of tropical cyclone-related (TC-related) precipitation to total precipitation is quantified using precipitation from NLDAS-2 and TC track information from HURDAT. Even though their contribution to total precipitation is minor, tropical cyclones bring important water resources during the onset of drought. This study shows the beneficial effects of TCs on drought occurrences, which had not been fully recognized before. Especially, this study highlights the impact of TCs on drought processes from initiation through recovery across multi spatial scales (from local to regional scales) from daily simulated soil moisture. In a changing cli-
mate, future changes in TCs and drought are expected, and thus, it is imperative to address the interplay between TCs and drought using a dynamical tropical cyclone model and a land surface model forced with different scenario climate projections.

In Chapter 5, the impact of the Atlantic and Pacific Oceans on U.S. drought is addressed using observational data within a Bayesian framework. Before the 1980s, decadal and inter-annual SST variability of the Pacific Oceans (PDO and ENSO) showed strong association with the central U.S. drought. In recent decades, their associations with central U.S. drought have been weakened, but their associations with southwest U.S. drought have been strengthened. In other words, strong SST forcings are not required to produce central U.S. drought, which warns us of over-confidence in SST teleconnections with the central U.S. drought in previous studies. This overconfidence can hinder U.S. drought forecasting, depending on the region.

In Chapter 6, the performance of the North American Multi-Model Ensemble (NMME) seasonal hindcasting for midwestern U.S. summer drought is addressed over the last three decades. Out of six climate models, only one climate model can produce strong inter-annual variability in summer precipitation anomalies that is comparable to that found in the observational data. Two of the NMME models show much stronger SST teleconnections with midwestern U.S. drought than those in the observational data. These strong teleconnections in the NMME models result in a high false alarm rate of hydroclimate extreme events (floods and droughts). The NMME seasonal forecasting models still predict floods and drought for the wrong reasons (e.g. a strong pan-Pacific El Nio-Southern Oscillation (ENSO)like pattern).

By crossing the spatial-temporal scales in studying drought, this thesis indicates that there are previously hidden processes involved in drought-generating mechanisms compared to a fixed scale (one dimensional) approach. A regional drought study not only over the U.S., but also over other regions is necessary within a multi-scale
approach, because key variables and processes can be found at different scales based on the findings of this thesis; the predictability of drought can come from different sources depending on the region. In addition, this thesis demonstrates that our current understanding does not account for the historical changes in the mechanisms that are revealed by testing the stationarity of the known mechanisms with the updated observational data. As a result, this insufficient understanding of recent changes in the mechanisms causes limited predictability of U.S. drought from the current seasonal forecasting models, which show overconfidence in associations between the Pacific Ocean and U.S. drought. This thesis demonstrates the need for updating the current version of climate models, taking into account changes in the drought mechanisms at decadal and multi-decadal scales, which will lead to robust improvement in drought predictability.

7.2 Next Steps and Future Work

This study provides guidance on how to conduct a multiple scale assessment of drought and reveals many benefits across different temporal and spatial scales. However, there are some limitations in this study. In this section, two major limitations from this study are discussed: 1) temporal and spatial limitations and 2) a generality problem (uncertainty assessment).

This study focuses on past major droughts because it only uses the observational data and hindcastings. This results in limited understanding of changes in drought risk due to climate change and human activity, and in evaluating the current predictability of future drought. In this context, the study in Chapter 2 can be further explored with a focus on the attributions of changes in drought risk by using a fine resolution hydrological model forced with climate projections. The hydrological models
account for a more realistic representation of the land surface (sub-grid heterogeneity, soil dynamics, and stream network) while global climate models simplify the land surface processes. Also, the hydrological model enables us to determine an optimal solution for water resource management and policy by testing and comparing alternatives with different scenarios.

Chapter 3 and Chapter 4 can be extended to the global scale. Initially, the recent reanalysis and data assimilation products have been developed over the U.S. with finer spatial (few kilometers) and temporal resolutions (sub-daily), and longer temporal coverage (longer than 30 years) than the previous data. In recent years, some of the regional projects have been extended to the global scale, and improved global datasets have been developed. For example, the National Land Data Assimilation System (NLDAS) project has been extended to the global scale as the Global Land Data Assimilation Systems project, which brings us a great opportunity to explore other regional and global drought mechanisms. Also, it allows us to extend the study in Chapter 4 to the global scale, which can address the contribution of tropical cyclones from other ocean basins to drought processes, by combining the track information and hydrometeorological data.

Another limitation of this thesis can be considered to be a result of the uncertainties inherent to single observational data or land surface model. There are multiple meteorological forcing datasets using different numbers of stations and methodologies. For example, precipitation is an important factor in the global terrestrial hydrological cycle and hydroclimate extreme events. However, the measurement of precipitation itself is uncertain due to underestimation of wind effects and limited spatial coverage.

In recent years, the quality of precipitation data has been improved by incorporating more precipitation data from in-situ measurements, remotely sensed data, and new methods for undercatchment correction. There are multiple precipitation data
products using different numbers of stations, different sources, and different methods. These datasets can be used for examining the sensitivity of uncertainties in our findings and the generalization of the findings across the products.

There are also different kinds of land surface hydrological models which treat land surface processes in different ways. These multiple land surface models can provide uncertainties (a spread) in the estimates of hydrological variables due to the different treatment of land surface processes, which is hard to assess using a single model. For example, the study of the impact of tropical cyclones on U.S. drought in Chapter 4 can be further studied using the multiple land surface model ensembles forced with or without TC related precipitation, which would enable analysis of the uncertainties of their impact on drought and to test the generalizations. Therefore, inter-comparisons among different datasets and land surface models are necessary to address the uncertainties of the findings in this thesis.

In a changing climate, the risk of drought is uncertain all over the world (Dai et al., 2004; Sheffield et al., 2012b; Trenberth et al., 2014). The Earth climate system has experienced significant changes in terrestrial hydrology and drought mechanisms. In recent decades, the hydrological cycle has been intensified leading to the “wetter is wetter, drier is drier” situation (Held and Soden, 2006). The roles of atmosphere and ocean in the drought mechanisms over the globe have become comparable due to global warming (Kumar et al., 2013). These changes cause more uncertainties in our knowledge of drought mechanisms.

To advance our robust prediction of drought over the globe, this thesis concludes that a multi-scale study of drought across temporal and spatial scales is beneficial in order for us to investigate the hidden processes and to test the consistency of recent events with the current understanding of the mechanisms. Inter-comparisons among datasets and land surface models are also suggested for a future study to
reduce the uncertainties in our knowledge of drought. This improved understanding of the drought mechanisms informs us of how to improve current climate forecasting models, resulting in the increased robustness in our prediction skill. Finally, these more realistic climate models can help to design and prepare better water resource management strategies and policies for future drought mitigation.


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