

Influence of Atlantic sea surface temperatures on persistent drought in North America

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Abstract This study investigates the relationship between North Atlantic sea surface temperatures (SST) and persistent drought in North America using modern observations, proxy paleo-data, and simulations from multiple climate models. The observational results show that persistent droughts in the Great Plains and the southwest North America are closely related to multidecadal variations of North Atlantic SST (Atlantic Multidecadal Oscillations, AMO). During the AMO warm (cold) phases, most of North America is dry (wet). This relationship is persistent since at least 1567 AD, as based on proxy SST for the North Atlantic and the reconstructed drought index for North America. On centennial timescales, proxy SST records from the North Atlantic and proxy drought records for North America suggest that major periods of AMO-like warm (cold) SST anomalies during the last 7.0 ka correspond to dry (wet) conditions in the Great Plains. The influence of North Atlantic SST on North American droughts is examined using simulations made by five global climate models. When forced by warm North Atlantic SST anomalies, all models captured significant drying over North America, despite some regional differences. Specifically, dry summers in the Great Plains and the southwest North America are simulated by all models. The precipitation response to a cold North Atlantic is much weaker and contains greater disagreement among the models. Overall, the ensemble of the five models could well reproduce the statistical relationship between the dry/wet fluctuations in the North America and North Atlantic

SST anomalies. Our results suggest that North Atlantic SSTs are likely a major driver of decadal and centennial timescale circulation, including droughts, in North America. Possible mechanisms that connect North Atlantic SST with North American drought, as well as interactions between North Atlantic and tropical Pacific SST and their relative roles on drought are also discussed.

1 Introduction

Drought is a ubiquitous feature of North American climate. The 1930s, 1950s and more recently the period from 1998 to 2005 were all characterized by severe drought conditions. The 1930s ‘Dust Bowl’ in particular has been considered as an icon of the toll that drought can take on humans and the landscape (Egan 2006). Despite the tremendous impact that these modern droughts have had on people and their economic, social, and agricultural systems, they are dwarfed in duration when compared to the megadroughts that affected the North America during the Holocene, e.g. the early to middle Holocene (Miao et al. 2007), and the medieval warm period (MWP, approximately 800–1300 AD) (Sridhar et al. 2006; Graham et al. 2007; Seager et al. 2007; Cook et al. 2004; Feng et al. 2008). Tree ring records show that the medieval droughts persisted for decades and even an entire century during MWP (Cook et al. 2004, 2007). During these severe medieval droughts, the grass that currently covers the sand dunes of the Nebraska Sand Hills, the largest sand dune body in the western hemisphere, disappeared and the dunes became mobile, indicating drought conditions much more severe than that during the twentieth century dust bowl (Sridhar et al. 2006). These records indicate that persistent droughts exceeding even the 1930s dust bowl have

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occurred frequently in the North America in the past, and therefore are likely not only to repeat in the future, but also be exacerbated by global warming due to increases in atmospheric greenhouse gases (Dai et al. 2004).

In the past decade, a growing body of evidence has shown a close relationship between North American drought and multidecadal variations of North Atlantic SST, i.e., the Atlantic Multidecadal Oscillation, or AMO (Kerr 2000). The AMO contains a roughly 60- to 80-year cycle. During the AMO warm (cold) phases, SST over the entire North Atlantic Ocean show basin-wide positive (negative) anomalies. Enfield et al. (2001) and Hu and Feng (2008) found that when the North Atlantic Ocean is warm (AMO warm phase), most of the US receives less than normal precipitation. Enfield et al. (2001) and Rogers and Coleman (2003) found that the AMO warm (cold) phase is linked to low (high) streamflow in the Mississippi Valley. McCabe et al. (2004) showed that AMO can significantly influence drought in the Midwest and the western US, contributing about 28% of the variance in drought frequency over the contiguous US for the period 1900–1999. They further noted that long-term predictability of North American drought may reside in the multidecadal behavior of AMO. The impact of AMO on North American drought has also been simulated by climate models (Knight et al. 2006; Sutton and Hodson 2005, 2007). Sutton and Hodson (2005, 2007) found that, when forced by an idealized AMO SST pattern for 1871–1999, the HadAM3 model is able to simulate the observed modulation of AMO on the boreal summer precipitation in the North America and the Europe.

These preliminary findings from both observational and modeling studies strongly suggest an important role for North Atlantic SST variations in severe, so called megadroughts in North America. The key issues are: (1) the existing observational results on the impact of the North Atlantic SST are limited to the past 50–100 years, just enough to cover one (or two) full cycles of the AMO. Whether the observed associations between North Atlantic SST and the North America drought in the past 50–100 years are physically robust, or just a statistically-random phenomenon is unknown. (2) Previous modeling studies (Knight et al. 2006; Sutton and Hodson 2005, 2007) are based on use of a single climate model. Ideally, multiple models should be used so as to avoid individual model biases (Krishnamurti et al. 2000).

This study addresses these issues by evaluating the impact of North Atlantic SST on persistent drought in North America across timescales from decadal to millennial using observational and proxy data. Use of proxy data for the last several thousands years to extend the observational record provides greater confidence in understanding the impact of North Atlantic SST on the persistent droughts in the North America (issue 1). The statistical

relationships obtained are also examined using simulations made by five global climate models. The ensemble of the five climate models could help overcome the potential biases inherent in any one model used to simulate the impact of Atlantic SST in single model (issue 2). In the next section, we describe the observed and proxy drought data and SST records. The statistical methods, the climate models and model experiments used to evaluate the relationship between Atlantic SST and North American drought are also described in this section. The relationships between Atlantic SST and persistent drought analyzed using observational and proxy data are presented in Sect. 3. The model responses to prescribed SST patterns in the North Atlantic are presented in Sect. 4. The results are discussed in Sect. 5. Possible physical mechanisms linking North Atlantic SST and persistent drought in North America are also discussed in this section. The study is summarized in Sect. 6.

2 Data and methods

2.1 Observed and proxy data

The Palmer Drought Severity Index (PDSI) and precipitation for 344 climate divisions in the continental US from 1895 to 2007 were obtained from the National Climate Data Center. The PDSI is frequently used to characterize drought variations over the North America. To evaluate the relationship between North American drought and Atlantic SST, the Global Sea Ice and SST (HadISST1) dataset on 1° latitude by 1° longitude for the same period is used (Rayner et al. 2003). The SST data were then aggregated to $5.0^\circ \times 5.0^\circ$ resolution to reduce the local noise for this analysis.

The observational relationship between Atlantic SST and the North American drought will be evaluated further using proxy data. The proxy SST used in this study include reconstructed annual SST with $5.0^\circ \times 5.0^\circ$ resolution over the Atlantic Ocean during 1750–1980 (Mann et al. 2000), and the reconstructed AMO index during 1567–1990 (Gray et al. 2004). Of particular interest for this study, Feng et al. (2009) analyzed 24 proxy SST records from the North Atlantic that span the last 10,000 years. The first empirical orthogonal function (EOF) of these 24 records accounts for 52.5% of the total SST variations during the Holocene. The spatial distribution of the first EOF (EOF01) shows dominant and coherent SST variations from the subtropical to the mid- and high-latitude North Atlantic, a pattern resembling the AMO as revealed by the first EOF of the instrumental data during the twentieth century. Feng et al. (2009) thus argued that the SST variations in the North Atlantic Ocean are dominated by an AMO-like pattern

throughout the Holocene. The first principal component (PC01) is thus a good index that represents the temporal variations of the AMO-like SST pattern during the Holocene. In this study, the PC01 of the 24 SST records in the North Atlantic (Feng et al. 2009) will be used to examine the relationship between the AMO-like Atlantic SST anomalies and persistent droughts in North America during the Holocene.

The proxy drought records used in this study include tree ring-reconstructed PDSI for the last 2,000 years (Cook et al. 2004). Beside the tree ring data, two sets of high resolution dry/wet records are also used. One is based on the $\delta^{18}O$ of speleothems from a core in Guadalupe Mountains, New Mexico (105°W, 32.5°N) for the last 10,000 years (Asmerom et al. 2007) and the other is the water-table depth reconstructed from a peat core in Minden Bog, Michigan (82.8°W, 43.6°N) for the last 3,500 years (Booth and Jackson 2003). The locations of the two sites are also shown in Fig. 1b. The variations of $\delta^{18}O$ are closely related to dry/wet fluctuations in New Mexico as revealed by the tree-ring reconstructed PDSI in the last 2,000 years (see Fig. 4b) that the records overlap. The variations of water-table depth in Minden Bog, Michigan in the last 3,500 years are closely related to lake level variations and sand dune activity in the central US (Booth and Jackson 2003; Booth et al. 2006). The comparison of these proxy drought records with the SST anomalies in the North Atlantic can assist our understanding of the impact of SST on the persistent droughts in the North America across various time scales.

2.2 Model data

We use output from simulations conducted with global climate models of the US-CLIVAR project (Schubert et al. 2009). These simulations were designed specifically to examine the relationship between persistent SST forcing and drought. The SST forcings are idealized SST anomaly patterns, obtained using rotated EOF analyses. This study only analyze the model simulations forced by warm and cold SST anomalies over the North Atlantic Ocean, with patterns resembling the actual SST anomalies associated with warm and cold AMO phases (Schubert et al. 2009).

The five global climate models used in this study are from NCEP (GFS) (Campana and Caplan 2005), NASA (NSIPP1) (Bacmeister et al. 2000; Schubert et al. 2004), GFDL (AM2.1) (Delworth et al. 2006; Milly and Shmakin 2002), LDEO/NCAR (CCM3) (Kiehl et al. 1998; Seager et al. 2005) and NCAR (CAM3.5) (Oleson et al. 2008; Stockli et al. 2008). (They are labeled as the GFS, NSIPP, GFDL, CCM3 and CAM3.5 respectively.) The same SST anomalies were imposed to force the five models in all

seasons. The impact of the absence of a seasonal cycle is expected to be small because the SST anomalies associated with AMO are robust in both winter and summer (Sutton and Hodson 2005). The control run (which was forced by climatological SST over the global oceans) and the warm/cold North Atlantic experiments were run for 36 years for GFS, 50 years for NSIPP, GFDL and CAM 3.5, and 51 years for CCM3. To make the model results comparable, only the last 35 years of output from each model are used in this study.

Schubert et al. (2009) examined the ability of the five models to reproduce the observed annual mean precipitation and atmospheric circulations. They found that, when forced by climatological SST, all the models produced quite reasonable annual mean stationary wave and annual mean precipitation patterns, albeit with some differences in local details. As shown in their Fig. 2, the local differences in precipitation between observations and the model simulations are found mainly over the oceans, whereas the differences over the North America are quite small. These results help substantiate the models' capabilities in simulating the response of drought to the SST anomalies in the North Atlantic Ocean.

2.3 Statistical methods

The relationship between North Atlantic SST and North American drought (as measured by PDSI, or precipitation) is examined using singular value decomposition (SVD; Bretherton et al. 1992). This statistical technique identifies pairs of spatial patterns that have the maximum temporal covariance between the PDSI and SST. Because the global SST is dominant by a warming trend (e.g. Schubert et al. 2009; Wang et al. 2008a), the long-term trend of SST in the North Atlantic is usually linearly removed to isolate the decadal and multidecadal signals (e.g. Enfield et al. 2001; Hu and Feng 2008). Since global ocean warming is nonlinear (e.g. Schubert et al. 2009), linearly detrended SST in the North Atlantic may still contain some signal of global ocean warming (e.g. Trenberth and Shea 2006; Mann and Emanuel 2006). On the other hand, drought frequency in North America appears increased following global warming (Dai et al. 2004; McCabe et al. 2004). Therefore, the global warming signal in the North Atlantic SST and the North American PDSI was first removed by linear regression before the SVD analysis. In particular, the global warming signal in PDSI (or SST) during 1895–2007 was linearly removed by regressing the PDSI (or SST) with the observed temperature anomalies over the Northern Hemisphere (Brohan et al. 2006). The global warming signal in the tree ring-reconstructed PDSI and the reconstructed gridded SST in the North Atlantic

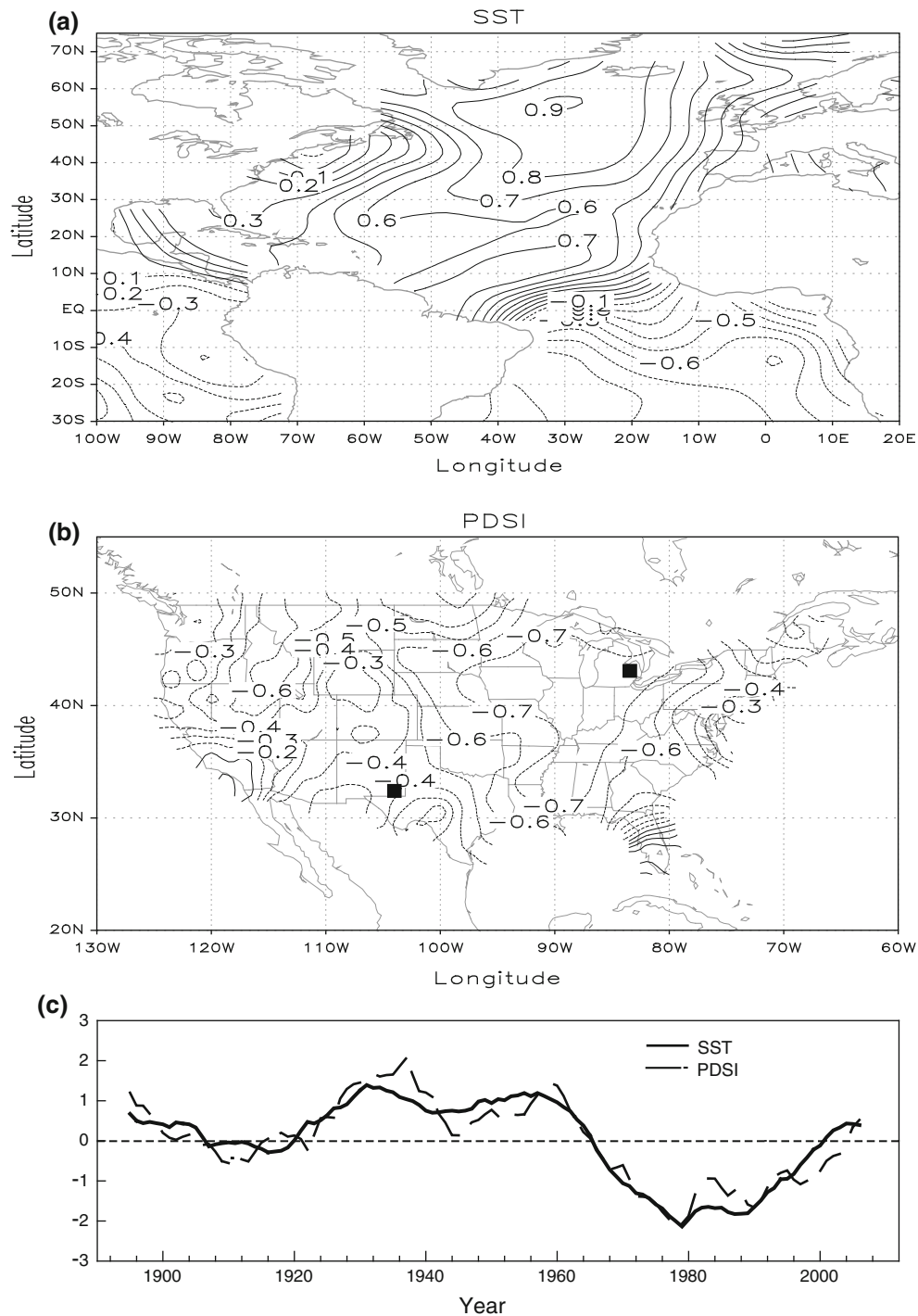


Fig. 1 The first SVD mode of the heterogeneous correlation of **a** observed annual SST in the North Atlantic Ocean, and **b** the annual PDSI on 344 climate divisions in the continental United States during 1895–2007. Squares show the location from where the high resolution proxy drought records during the Holocene were obtained.

c Normalized expansion coefficients of the first SVD mode for SST (solid line) and PDSI (dashed line). Large positive PDSI expansion coefficients suggest that the regions with negative correlation in the North America are drier

during 1750–1980 were also removed by regressing with the reconstructed temperature anomalies over the Northern hemisphere (Mann et al. 1999). To examine the impacts of the Atlantic SST on the persistent droughts in

North America on decadal and longer timescales, a 15-year smoothing was also applied to the observed and reconstructed PDSI and SST data after the global warming signal was removed.

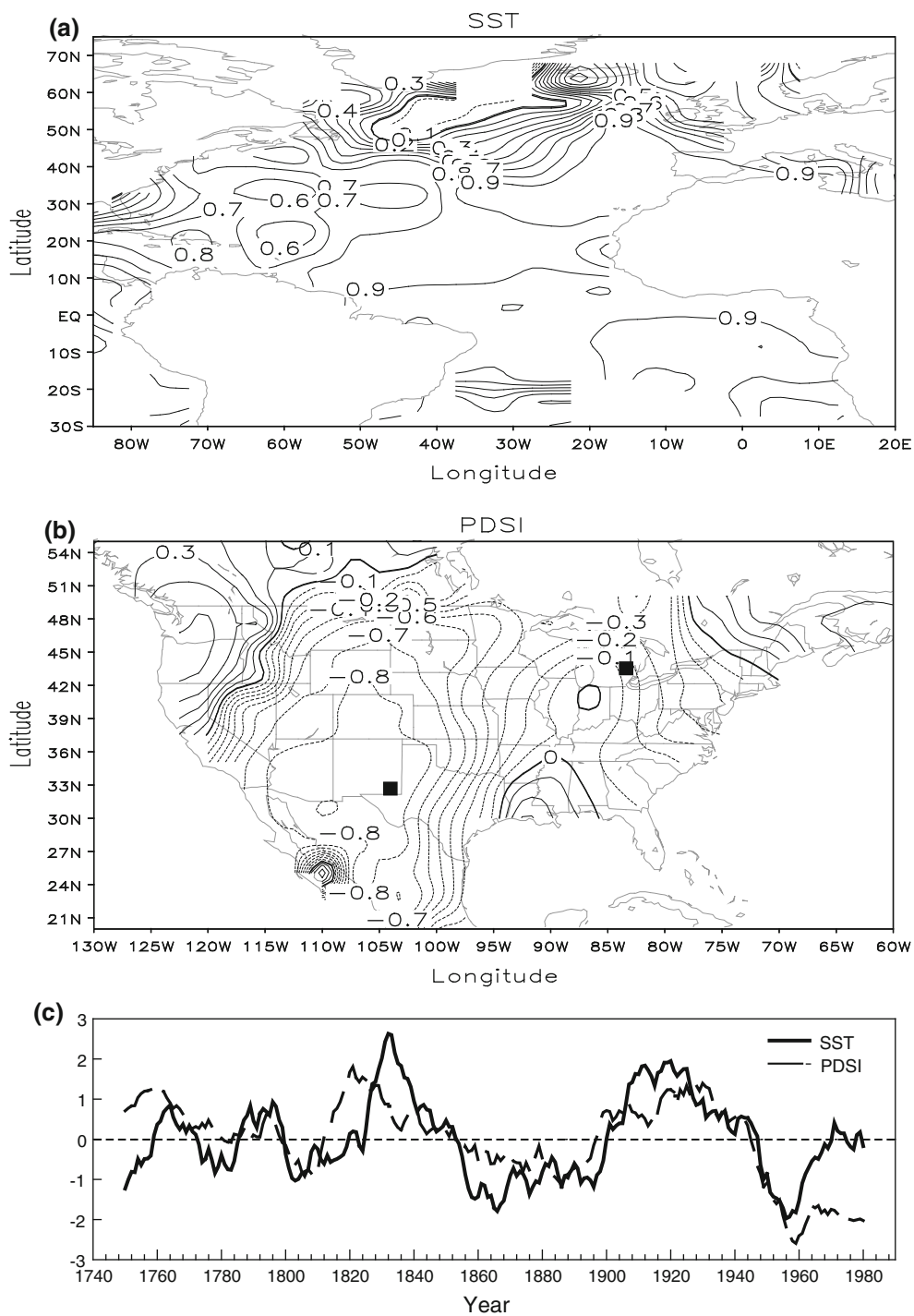


Fig. 2 The first SVD mode of the heterogeneous correlation of **a** reconstructed annual SST in the North Atlantic Ocean, and **b** the tree ring reconstructed PDSI in the North America during 1750–1980. **c** Normalized expansion coefficients of the first SVD mode for SST

(solid line) and PDSI (dashed line) for the same period. Large positive PDSI expansion coefficients suggest that the regions with negative correlation in the North America are drier

3 Statistical relationship between Atlantic SST and persistent drought in North America

To explore connections between North Atlantic SST and persistent drought in North America, the SVD analyses

were performed by using the covariance matrices of annual PDSI in the US and the SST in the Atlantic Ocean. The SVD modes were obtained using instrumental observations for 1895–2007, and proxy reconstructions for 1750–1980. Table 1 lists the statistics for the first SVD mode between

Table 1 Statistics of the first leading SVD mode between SST in the North Atlantic Ocean and PDSI (or precipitation) in the North America

	Square covariance (%)	SST variance (%)	Drought variance (%)	Temporal correlation of the expansion coefficients
SST-US PDSI (1895–2007)	67.3	38.4	36.0	0.93
SST-US Precipitation (1895–2007)	64.1	38.3	30.4	0.93
SST-North America PDSI (1750–1980)	68.1	51.9	19.4	0.58

the PDSI and SST, including the percentage of covariance explained, the temporal correlation between the expansion coefficients, and the variance in individual fields (PDSI, or SST) explained.

The spatial patterns of the first SVD mode based on instrumental records are shown in Fig. 1. During 1895–2007, the first SVD mode in the Atlantic Ocean is characterized by an AMO pattern, with large positive correlations over the entire North Atlantic and weak positive or neutral correlations in the tropical region (Fig. 1a). The corresponding PDSI (Fig. 1b) displays strong negative correlation (dry conditions) over the Midwest and the Great Plains. A moderate negative correlation is found over the southwest US. This is the typical drought pattern associated with AMO, also identified by the correlation between the AMO index and PDSI (or precipitation) over the US (e.g. McCabe et al. 2004, 2008; Hu and Feng 2008, Enfield et al. 2001). This mode explains 67.3% of the covariance between SST in the Atlantic Ocean and the PDSI in the US. The SST pattern itself explains 38.4% of the total annual SST variance over the Atlantic Ocean, while the PDSI pattern itself explains 36.0% of the annual PDSI variance over the US.

The temporal variations of the expansion coefficient for the first SVD mode during the instrument period are shown in Fig. 1c. The temporal correlation between the Atlantic SST and the associated PDSI time series is 0.93. The SST time series displays clear decadal variations, with large positive fluctuations during the 1920s to 1960s and after the middle 1990s, consistent with the known phase changes of AMO (Enfield et al. 2001). The expansion coefficients of PDSI exhibit coherent fluctuations with the Atlantic SST, indicating a strong association between the PDSI pattern and the Atlantic SST. Large positive PDSI expansion coefficients occurred in the 1930s, 1950s and after 2000, concurrent with severe drought periods in the US, are all associated with warm SST anomalies in the North Atlantic.

We also performed the SVD analysis between the annual North Atlantic SST and the annual precipitation on 344 climate divisions in the US using data for the period 1895–2007. The total covariance explained by the first

SVD mode and the correlation between the expansion coefficients of the precipitation and SST are very similar to the relationship between PDSI and SST (Table 1). The spatial distributions of the first SVD mode between precipitation and SST (figure not shown) are also very similar to that of PDSI and SST (Fig. 1), suggesting that the relationship between North Atlantic SST and drought in North America is robust and not sensitive to the specific drought records used.

For the period 1750–1980, the first SVD explained 68.1% of the covariance between the Atlantic SST and North America PDSI. The SST pattern (Fig. 2a), which explains 51.9% of the total SST variance, features warm SST anomalies across the North Atlantic, except for a small region south of Greenland. This SST pattern is similar to the first SVD mode (Fig. 1a) and the SST anomalies associated with AMO. The first mode of PDSI, which explains 19.4% of the total variance, has a well-defined pattern with negative correlations in the Great Plains and the southwest North America, and weak positive correlations in New England and the Pacific Northwest. Compared to Fig. 1b, the negative correlation in the Great Plains is slightly weaker, while the negative correlation in the Southwest North America is stronger during the past 250 years. Such differences may be related to fewer tree ring chronologies being available to reconstruct the past PDSI variations in the Great Plains (Cook et al. 2004). Time series of the expansion coefficients of the first SVD mode during 1750–1980 are shown in Fig. 2c. The temporal correlation between Atlantic SST and the associated PDSI in North America is 0.58. Figure 2 also suggests that when the North Atlantic is warm, there is a tendency for dry conditions in the Great Plains and the southwest North America.

Such a connection between North Atlantic SST and the PDSI is also obvious when extended further back through the last 420 years. Figure 3 shows the reconstructed AMO index (Gray et al. 2004) and the PDSI averaged over the Great Plains and the southwest North America since 1567 AD. Inspecting Fig. 3 we find that, when the North Atlantic is warm, the two regions are dry, and vice versa. The correlation between AMO index and the drought in the

Great Plains and the southwest North America are -0.374 and -0.425 , both significant at 95% confidence level by Monte Carlo test.

The North America suffered from a severe megadrought during the MWP (Graham et al. 2007; Seager et al. 2007; Feng et al. 2008). Figure 4 shows the difference between the tree ring reconstructed PDSI averaged for the period 900–1200 AD, and the twentieth century. Most of the US, northern Mexico, and the Canadian Prairie provinces during medieval times were drier than during the twentieth century. The largest changes occurred over the Great Plains and the southwest US. The occurrence of medieval drought was also supported by proxy data from terrestrial eolian deposits and alluvial stratigraphic evidence, and lake sediment chemical and salinity reconstructions (Graham et al. 2007; Seager et al. 2007; Feng et al. 2008). Previous studies suggested that those medieval droughts are related to a cool tropical Pacific during medieval times. Others (Booth et al. 2006; Feng et al. 2008) have argued that the medieval droughts are also related to warm temperature in the North Atlantic. Figure 4 shows the proxy temperature anomalies during the medieval warm period over the North Atlantic (see Feng et al. 2009 for detail of those data). It is clear that there is a basin wide warming pattern over the North Atlantic region during the MWP. Associated with this AMO-like warm SST pattern, there were severe droughts in the Great Plains and southwest US, and slightly

wetter conditions in New England and the west coast of Canada. Comparison with Figs. 1 and 2 suggests that the relationship of centennial-scale medieval drought to basin-wide SST warming in the North Atlantic Ocean are very similar to that on multidecadal timescales.

To examine the impact of the North Atlantic SST on the drought in North America during the Holocene, Fig. 5 shows the PC01 for the Holocene SST (Feng et al. 2009) and the two dry/wet records in North America (Asmerom et al. 2007; Booth and Jackson 2003). The centennial timescale dry/wet variations in the Midwest and the southern Great Plains are clearly shown. The megadrought during the MWP is just one of several such droughts on centennial timescales that affected North America. Additionally, though the two sites are about 2,250 km away from each other, the dry/wet fluctuations on centennial time scales in those two regions varied simultaneously, especially before 1200 AD, suggesting that the centennial dry/wet variations in the Midwest and the southern Great Plains are controlled by a common set of forcings, presumably the AMO-like SST variations in the North Atlantic Ocean. As marked by the vertical gray bars in Fig. 5, there were seven major AMO-like warm periods in the North Atlantic during the last 7.0 ka, each corresponding to dry conditions in the two regions of North America. It is also obvious that the cold periods in the North Atlantic during the last 7.0 ka align with wet conditions in North America. When interpolating the drought index to a 50 year resolution (same temporal resolution as the PC01 of the Atlantic SST), the correlation between PC01 and the water table in Michigan during 3.6–1.0 ka BP is -0.35 , and the correlation with $\delta^{18}O$ in New Mexico during 7.0–1.0 ka BP is -0.235 , both significant at the 95% confidence level, suggesting a significant impact of Atlantic SST on drought/wet fluctuations in the two regions during the last 7.0 ka.

This relationship, however, is reversed before 7.0 ka BP, with AMO-like warmer (colder) Atlantic associated with wetter (drier) conditions in the North America (Fig. 5b, c). This change is likely related to the presence of the Laurentide Ice Sheet (LIS) that covered the mid- and high-latitude North America prior to this time. Figure 5d shows the variations of the area of LIS during the Holocene (Kaplan and Wolfe 2006). During the earlier Holocene (e.g. 10.0 ka BP), there are about 8×10^6 km² of the North America covered by ice. As the temperature became warmer from the early to mid-Holocene, the LIS melted abruptly. The area of LIS decreased to about 2.0×10^6 km² during 6.0–7.0 BP and has remained nearly unchanged since then. Because the LIS can significantly alternate the surface energy balance of the high latitude North America (Bromwich et al. 2004), the larger area of North America covered by LIS before 7.0 BP likely overrode the effect of

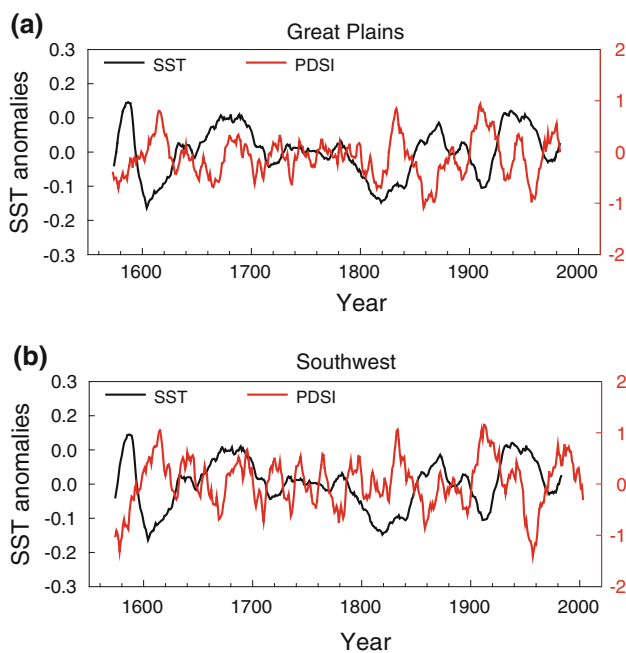


Fig. 3 Regional averaged PDSI for **a** the Great Plains (30° – 50° N, 90° – 110° W) and **b** the southwest North America (25° – 40° N, 95° – 120° W) and their relationship with the reconstructed AMO index (dark line) during 1567–1990

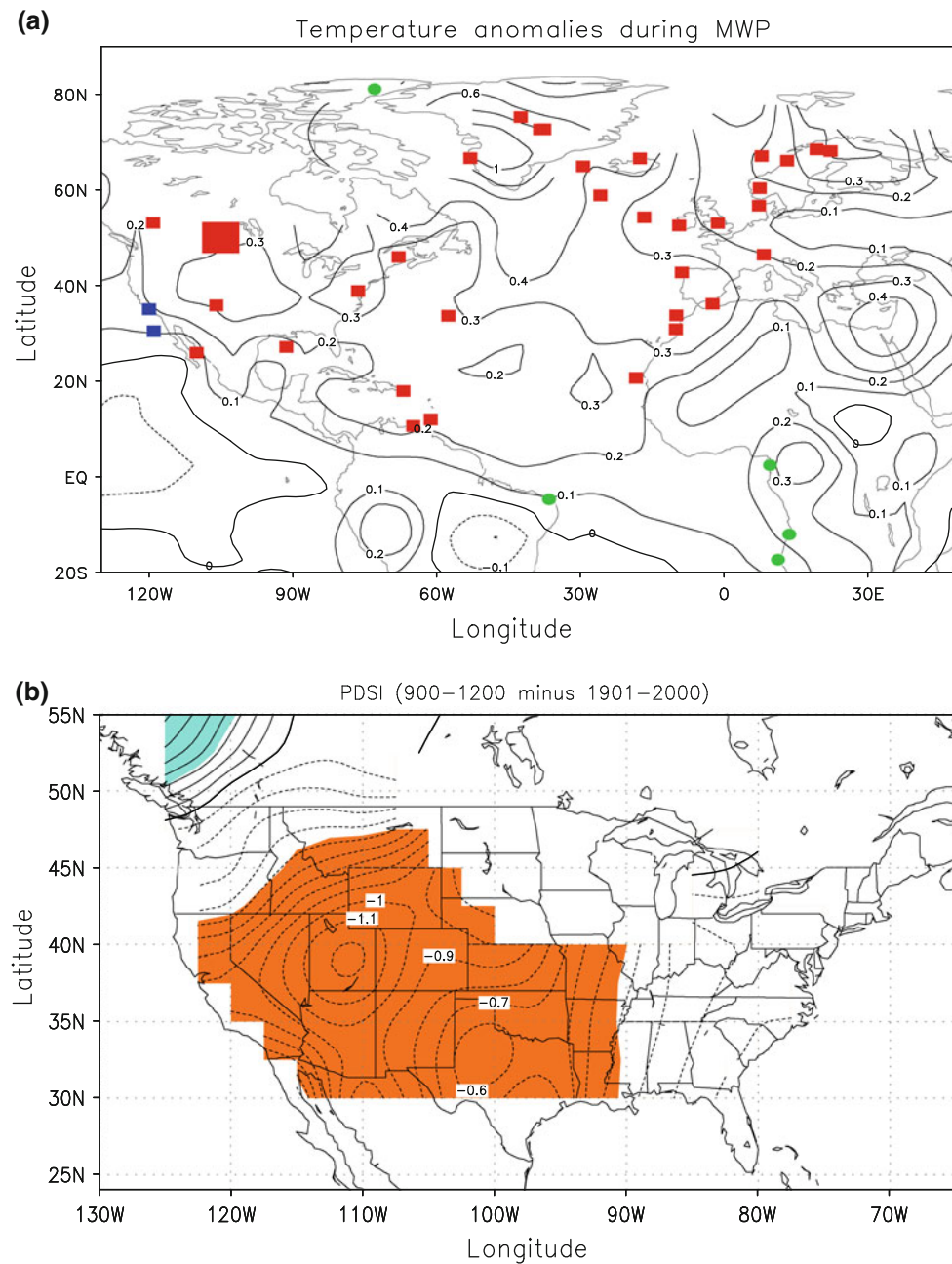


Fig. 4 a Spatial distribution of the proxy data that recorded temperature changes during MWP. *Blue squares, green circles and red squares* indicate cooling, no changes and warming during MWP, respectively. The contour lines are the observed temperature anomalies associated with AMO warm phases for the period 1901–2006.

Detail of those proxy data can be found at Feng et al. (2009). **b** Difference in tree ring reconstructed PDSI for 900–1200 AD minus 1901–2000. *Shadings* indicate the differences are significant at 95% confidence level by two-tailed student-test

Atlantic SST, thereby weakening the relationship between the SST and drought.

The above results indicate that the AMO and AMO-like SST variations in the North Atlantic Ocean is likely a major and consistent driver of decadal and centennial timescale circulation and drought variability in North America during the last 7.0 ka. Warmer SST in the North Atlantic can cause persistent droughts and colder SST can

cause wet conditions. Results from previous studies (e.g. Gupta et al. 2003; Feng and Hu 2008) have also indicated consistent and significant relationships between summer rainfall in Asia and Africa and North Atlantic SST anomalies during the Holocene. A relatively warm North Atlantic is associated with enhanced monsoon rainfall in Asia and West Africa. This evidence further supports the notion that North Atlantic SST anomalies may have

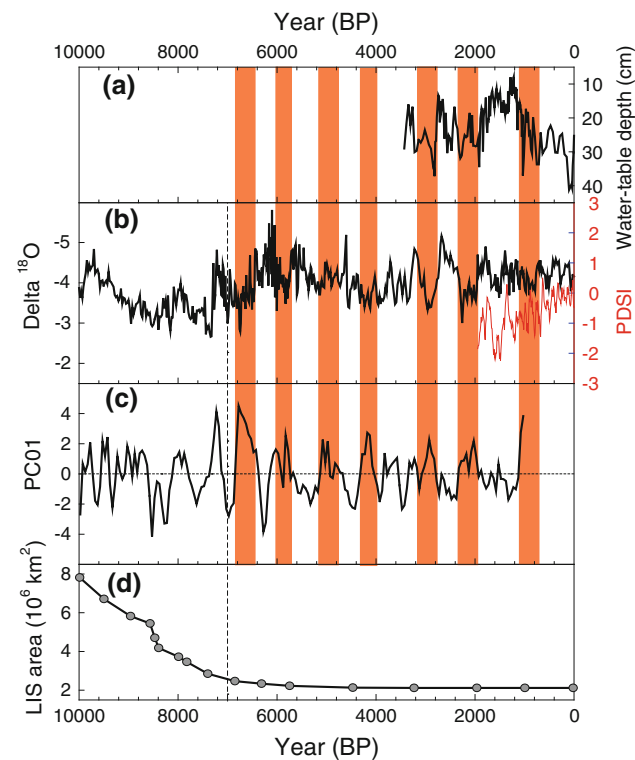


Fig. 5 Centennial scale variations of **a** the water table in Minden Bog, Michigan (Booth and Jackson 2003). Deeper water table indicates dryer condition. **b** The $\delta^{18}O$ of speleothems from Guadalupe Mountains, New Mexico (Asmerom et al. 2007). Smaller $\delta^{18}O$ values indicate enhanced precipitation and wetter conditions. Variations of 50-year smoothed PDSI for New Mexico are shown by the thin line for the past 2,000 years. Positive (negative) PDSI indicate wet (dry) condition. **c** Centennial variations of the PC01 of the 24 proxy SST in the North Atlantic Ocean (Feng et al. 2009). The vertical bars mark the centennial epochs of warm anomalies of the Atlantic Ocean SST and the corresponding dry conditions in the North America. **d** Variations of the Area of Lanrentide Ice Sheet during the Holocene (Kaplan and Wolfe 2006)

provided key climate forcing at decadal and centennial timescales in the Northern Hemisphere.

4 Model simulations

The proxy drought data used in this study were mainly PDSI values based on dendrochronological analysis. Although the PDSI measures the combined effect of temperature and precipitation on drought, precipitation is probably the major factor controlling the variations of PDSI on interannual and longer time scales (Guttman 1998). This is also supported by the SVD analysis between annual SST in the North Atlantic and annual precipitation in the US (Table 1). Therefore, in the following, we will focus on the responses of modeled precipitation to the Atlantic SST anomalies, and compare them with the statistical relationships shown in Sect. 3.

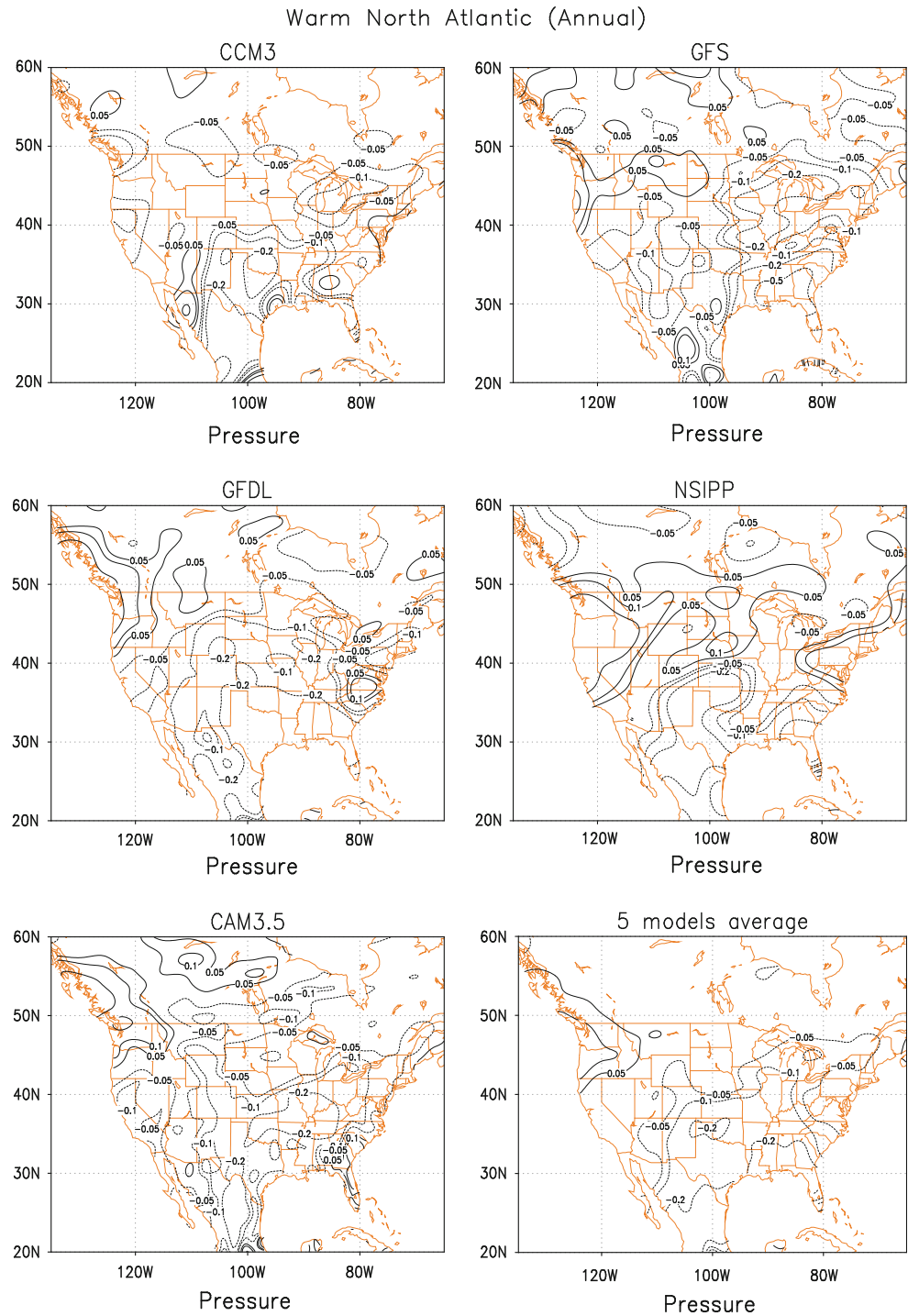
4.1 Annual precipitation

The difference in the simulated annual precipitation forced by a warm North Atlantic Ocean (a pattern similar to the AMO warm phase) and the control run is shown in Fig. 6. The large-scale results show considerable similarity among the models. In particular, all models capture the major dry features over the North America. Simulations with a warm North Atlantic Ocean yield 0.1–0.3 mm/day precipitation deficits (about 10–20% of the annual average) over the Great Plains (30°–50°N, 90°–110°W) and the southwest North America (25°–40°N, 95°–120°W), consistent with the statistical results shown in Sect. 3. When precipitation is averaged over each region, all models showed a tendency towards dry condition. The reductions in annual precipitation, though small in magnitude, are nonetheless all significant at the 95% confidence level (Table 2).

While all the models show a tendency for dry conditions in response to a warm North Atlantic Ocean, there are some differences. For example, the precipitation response in GFS is largely located in the Midwest and the southeast United States, while the responses in the other models are more evident in the southern Great Plains. On the other hand, the CCM3 simulated widespread dry conditions for most of North America, while the other models produced wetter conditions over the northwest US and southwest Canada. When the five models are averaged, these small, individual regional responses were smoothed out (Fig. 6). The five-model average shows dry conditions over most of the North America, except the northwest US and southwest Canada. Comparing and contrasting the (averaged) precipitation anomalies of the five models with the statistical results shown in Figs. 1, 2 and 4 suggests that the 5-model ensemble can well reproduce the impact of warm North Atlantic SST anomalies on drought in North America.

Compared to those for a warm North Atlantic, the model responses to a cold North Atlantic Ocean (a pattern similar to the AMO cold phase) are much weaker and less spatially coherent (Fig. 7). All models except for CAM3.5 produced wetter conditions over the south-central Great Plains, albeit the magnitudes differ. Over the Midwest, three models (CCM3, GFS and GFDL) showed drier conditions, whereas the other two models (CAM3.5 and NSIPP) showed weakly wet to neutral conditions. The CAM3.5 is very different from the other models in simulating drier conditions over most of the continental US. This suggests a potentially large model bias in simulating the influence of cold North Atlantic SST anomalies. The average of the precipitation anomalies for the five models yielded slightly wetter conditions over the Great Plains and the southwest North America, generally consistent with the statistical results shown in Sect. 3. When the precipitation in the Great Plains is averaged, the response to a cold North Atlantic is overall

Fig. 6 Simulated annual averaged daily precipitation anomalies (mm/day) forced by warm North Atlantic SST anomalies (a pattern resemble the SST associated with AMO warm phase) made by five climate models and their ensemble. The data shown here are the difference between the modeled precipitation and the control runs



small and insignificant (Table 2). Precipitation in the southwest North America in response to cold North Atlantic in CAM3.5 is somewhat less biased than in the Great Plains. As shown in Fig. 7, the CAM3.5 has drier conditions in the southwest US, but wetter conditions in northern Mexico. All other models simulated wetter conditions in both regions. Compared to results from the warm North Atlantic simulations (Fig. 6; Table 2), the

precipitation response to a cold North Atlantic is much weaker in the Great Plains, but of a similar magnitude in the southwest.

4.2 Summer precipitation

In the Great Plains and the Southwest North America, most of the annual total precipitation occurs during the warm

Table 2 Averaged annual precipitation anomalies (mm/day) in the Great Plains (30°–50°N, 90°–110°W) and the southwest North America (25°–40°N, 95°–120°W, shown in the parenthesis) simulated by the five climate models

Models	Warm Atlantic minus the control runs	Cold Atlantic minus the control runs	Warm minus cold
CCM3	−0.07* (−0.10*)	0.08* (0.10*)	−0.15* (−0.21*)
GFS	−0.07* (−0.06)	0.05 (0.11*)	−0.13* (−0.16*)
GFDL	−0.14* (−0.16*)	0.08* (0.13*)	−0.22* (−0.29*)
NSIPP	−0.04* (−0.06*)	0.05* (0.12*)	−0.09* (−0.18*)
CAM3.5	−0.10* (−0.12*)	−0.00 (0.00)	−0.10* (−0.12*)
5-model average	−0.09* (−0.10*)	0.05 (0.09*)	−0.13* (−0.16*)

* Indicate the precipitation anomalies are significant at 95% confidence level by two tailed student-test

season. To understand the warm season precipitation changes in response to North Atlantic SST variations, the summer (June–August) precipitation simulated by the five models are analyzed. Because the responses of summer precipitation to warm/cold North Atlantic in each individual model are very similar to those for annual precipitation (figure omitted), only the ensemble results of the five models are shown here (Fig. 8). In the control runs, the ensemble average shows peak precipitation in southern Mexico and the central and eastern United States, primarily fueled by moisture fluxes from the Gulf of Mexico and the Caribbean Sea. The west coast is dry and influenced by the subtropical Pacific high and the cold north/northwest air flow from the cold North Pacific (Fig. 8d).

Compared to the control runs, summer precipitation of the five models shows wide spread dry conditions over the contiguous US when the North Atlantic is warm (Fig. 8a). Large reductions (about 0.5 mm/day) occur over the southern Great Plains, while slightly wetter conditions occur in the southwest Canada. These precipitation anomalies resemble the responses of annual precipitation to the warm North Atlantic (Fig. 6). Figure 8a shows anomalous easterly flow dominates the eastern half of the US. The east wind anomalies then curve to south (north) in the southern (northern) Great Plains, creating a wind divergence over the Great Plains. Because wind anomalies are associated with similar variations in moisture transport (Wang et al. 2008b), these wind anomalies in the southern Great Plains and eastern Mexico indicate that the low level jet and hence the moisture transport from the Gulf of Mexico is reduced. These cyclonic wind anomalies in the Gulf of Mexico also suggest that the subtropical high in the Gulf of Mexico and Caribbean Sea is weaker (and/or shifted east) when the North Atlantic is warm. Less moisture transport from the Gulf of Mexico and the divergence in the Great Plains together cause the dry conditions in most of the United States.

Compared to a warm North Atlantic, the response of summer precipitation to a cold North Atlantic is weak and less spatially coherent (Fig. 8b), similar to the annual

precipitation (Fig. 7). Weak wetter conditions occur in the southern Great Plains and northern Mexico. There is an anticyclonic wind anomaly in the Gulf of Mexico, which brings more moisture and excess precipitation to the southern Great Plains and eastern Mexico (Wang et al. 2008b; Hu and Feng 2008; Feng et al. 2008).

4.3 Seasonal precipitation in the Great Plains and Southwest North America

The precipitation response to a warm (cold) North Atlantic Ocean also varied by season (keeping in mind the seasonally-invariant SST anomalies used to force the models). Figure 9 shows the signal to noise ratio (SNR) for precipitation in the Great Plains in response to warm (cold) North Atlantic SST forcing for each season and each model. The SNR is used to measure the year-to-year scatter in precipitation response to SST forcing, and is defined as (Schubert et al. 2009):

$$\text{SNR} = (\bar{x} - \bar{y}) / S_{xy}$$

Where x and y represent seasonal precipitation values from the warm (or cold) North Atlantic SST forcing and the control runs, respectively, and the bar denotes the 35 years average of model precipitation. The $S_{xy}^2 = (S_x^2 + S_y^2) / 2$, where S_x^2 and S_y^2 are the variances of x and y , respectively. The mean precipitation difference between the warm (or cold) North Atlantic SST forcing and the control runs ($\bar{x} - \bar{y}$) is a measure of the signal, so that SNR measures the size of the signal in units of standard deviation (Schubert et al. 2009). A positive (negative) SNR indicates the North Atlantic SST forcing produced a wetter (drier) condition compared to the control run. The confidence level of SNR is examined using two-tailed student-test.

Figure 9 shows that the response of precipitation to North Atlantic SST contains strong seasonal variations. The Great Plains response to a warm North Atlantic is significant (or marginally significant) for all five models during the warm seasons (May–August). However, the

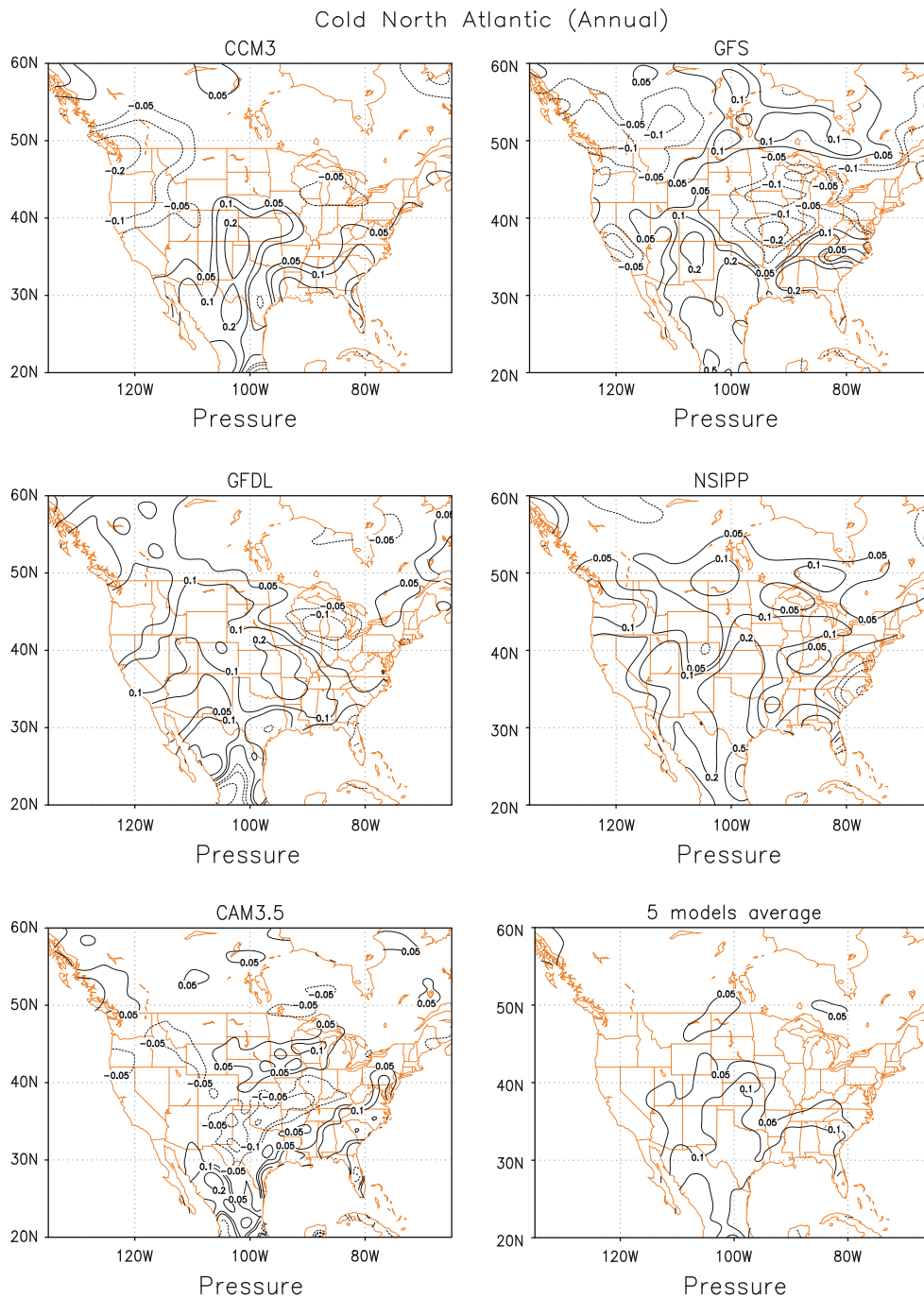


Fig. 7 Same as Fig. 6, but for cold North Atlantic SST anomalies

precipitation response to a cold North Atlantic shows less agreement among the models. For example, the warm season precipitation in the Great Plains shows virtually no response to a cold North Atlantic in GFDL model. The CAM3.5 produced dry conditions in summer in the Great Plains when forced by either a warm or cold North Atlantic SST. Figure 9 also shows considerable disagreement among the models in term of the seasonality of the

response. For example, large and significant responses to warm and cold Atlantic Ocean in the GFDL model appear during winter to early summer (December–June), whereas for the NSIPP model, large and significant response occurred during the summer and fall. When the results of the five models are averaged, the Great Plains is generally dry (wet) for all seasons when the North Atlantic is warm (cold). The response of precipitation to North Atlantic SST

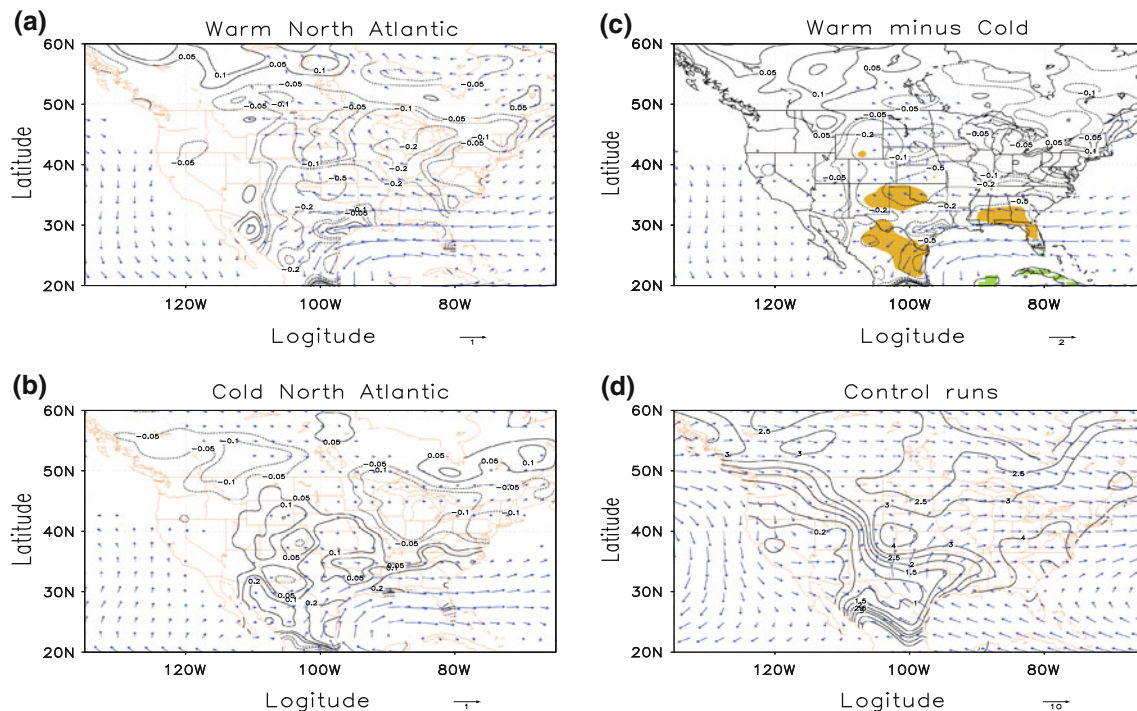


Fig. 8 Simulated summer-averaged daily precipitation (mm/day, contours) and the 850 hPa wind (m/s, vectors). Results shown are for the ensemble of the five models. **a** Difference between warm North Atlantic SST forcing and the control run. For clarity, only wind speed anomalies larger than 0.2 m/s are shown. **b** Same as **a** but for

cold North Atlantic SST forcing. **c** is the difference between **a** and **b**. *Shadings* (vectors) indicate the precipitation differences (wind anomalies) between warm and cold North Atlantic SST forcing are significant at the 95% confidence level. **d** is the control run, which was forced by climatological SST

is generally weaker in the cold season, possibly reflecting the indirect impact of the Atlantic SST on the North American climate during the wintertime (Kushnir et al. 2010; more details are discussed in Sect. 5).

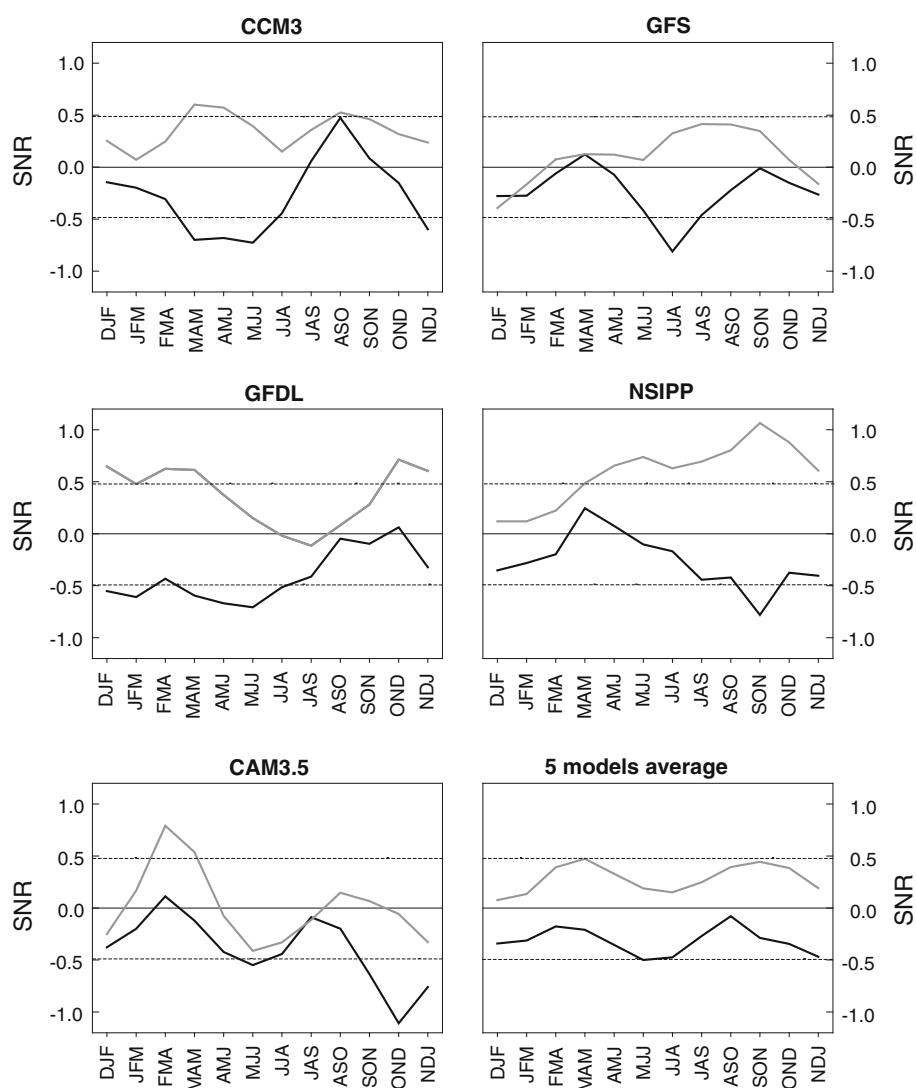
Precipitation in southwest North America in response to Atlantic SST for each season and for all models is shown in Fig. 10. All models simulated a significantly drier summer when the North Atlantic is warm. Compared to the Great Plains, precipitation in southwest North America shows a similar seasonality to cold North Atlantic SST forcing, except the response is stronger. The CAM3.5 again simulated a dry summer in southwest North America when forced by either warm or cold North Atlantic. When the results of the five models are averaged, the southwest North America is dry for all seasons when North Atlantic SST is warm, whereas the precipitation response to cold North Atlantic is generally wet, consistent with the statistical results shown in Sect. 3.

5 Discussion

The results in the previous sections demonstrate a strong and persistent effect of Atlantic SST on dry/wet fluctuations in North America. A warm (cold) North Atlantic is

associated with persistent droughts (wet periods) in the Great Plains and the southwest North America on multi-decadal and centennial time scales. This relationship is persistent during the last 7.0 ka and is also simulated by five atmospheric models when forced by SST anomalies in the North Atlantic Ocean. Because the intensity and size of the tropical west North Atlantic warm pool closely follows the variations of AMO (Wang et al. 2008a). A warm North Atlantic means a stronger and larger tropical west North Atlantic warm pool, which is associated with a weaker North Atlantic Subtropical High (NASH) in the Gulf of Mexico and southeast US, thereby helping to induce northerly wind anomalies (weaker low level jet) in the Great Plains (Wang et al. 2006; Hu and Feng 2008; Feng et al. 2008; Enfield and Cid-Serrano 2010; see also Fig. 8). Such an anomalous circulation transports less moisture into the Great Plains and Southwest (Wang et al. 2006, 2008b). Additionally, warming in the tropical Atlantic can cause northerly subsiding (drying) flow over the Great Plains and the southwest North America (Kushnir et al. 2010). This subsiding flow and reduced moisture transport to North America combine to cause dry conditions in the Great Plains in summer. On the contrary, a cold North Atlantic SST results in a weaker and smaller Atlantic warm pool (Wang et al. 2008a). Associated with these SST changes

Fig. 9 Seasonality of the signal to noise ratios (SNR, see text) of the 3-month average precipitation responses in each model in the Great Plains (30°–50°N, 90°–110°W) to the warm (dark lines) and cold (grey lines) SST anomalies in the North Atlantic Ocean. Results are based on the last 35-year simulations. The horizontal dashed lines denote the 95% confidence levels based on the two-tailed student-test

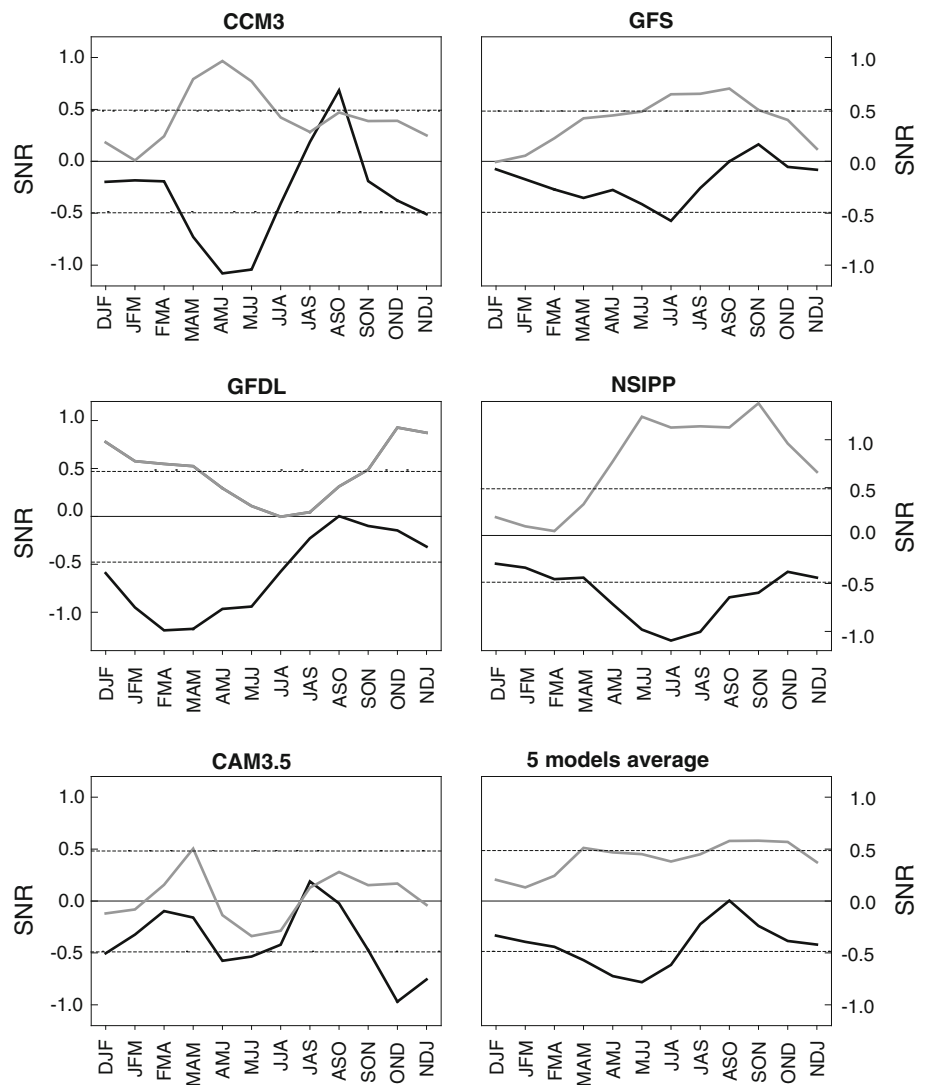


are positive low level height anomalies (stronger NASH) in the tropical North Atlantic and southeast US, and a stronger southerly wind to the Great Plains (Fig. 8). The stronger southerly wind brings more water vapor to the Great Plains and the southwest North America; resulting in excess precipitation in these regions (Wang et al. 2006, 2008b; Hu and Feng 2008).

The influence of the Atlantic in winter is different from summer. Kushnir et al. (2010) recently analyzed the influence of Atlantic SST using a linear primitive equation model. They argued that a warm tropical North Atlantic in winter could warm the upper tropical troposphere and help stabilize the atmosphere over the tropical Pacific. This suppressed heating over the Pacific will then induce a remote response (teleconnection) influencing the North Pacific and North America. This indirect impact of wintertime North Atlantic SST likely explains the weaker response of the cool season precipitation to the Atlantic SST forcing as shown in the models (Figs. 9, 10).

This study focuses on the influence of North Atlantic SST on North American drought. SST anomalies in the tropical and eastern North Pacific can also influence the drought in North America (e.g. Trenberth and Guillemot 1996; Hu and Feng 2001). Currently, there is an excellent understanding that historical droughts in the western North America are closely linked to cool La Nina-like SST anomalies in the tropical Pacific Ocean (e.g. Cook et al. 2004; 2007; Seager et al. 2005, 2007; Herweijer et al. 2006, 2007; Graham et al. 2007; Feng et al. 2008; Hoerling et al. 2009; Conroy et al. 2009). Recent statistical and modeling studies (McCabe et al. 2004; Feng et al. 2008; Schubert et al. 2009; Conroy et al. 2009) have examined the influence of both Atlantic and Pacific SST anomalies on the development of drought in the North America. Their results suggest that either a cold tropical Pacific or a warm North Atlantic can produce droughts. Schubert et al. (2009) concluded that the Atlantic forcing did not appear to be as robust as the

Fig. 10 Same as Fig. 9 but for southwest North America (25°–40°N, 95°–120°W)



Pacific, whereas Feng et al. (2008) and Conroy et al. (2009) argued that both the North Atlantic and Pacific oceans are probably necessary to produce the intensity and longevity of the persistent drought in the North America.

Schubert et al. (2008) and Wang et al. (2006) argued that the tropical Pacific SST anomalies associated with ENSO typically cause changes in the planetary-scale waves and Pacific storm tracks (primarily during the winter and early spring), while the Atlantic SST anomalies cause changes in the subtropical High and associated low-level moisture fluxes transport to the North America (primary during the summer and fall). Therefore, the influences of cold tropical Pacific and warm Atlantic Ocean on the droughts in North America are expected to differ regionally and seasonally. This hypothesis is supported by previous studies (e.g. Cook et al. 2004, 2007) and the results

of our study. For example, the influence of a cool tropical Pacific is mainly identified over the western North America (Cook et al. 2004, 2007), where most of the annual precipitation occurs in the cold season. On the contrary, our results clearly showed that the influences of warm North Atlantic Ocean are mainly in the Great Plains and the southwest North America, where most of the annual precipitation occurs in the warm season (Figs. 6, 8). How the tropical Pacific and the North Atlantic interact, the relative roles of each ocean on the drought, and the seasonality of their effects are still not clear, and certainly deserve further investigation. Additionally, the role of land surface interactions and feedbacks with the regional atmospheric circulation (e.g. Oglesby and Erickson 1989; Oglesby et al. 2001) must also be included to fully understand and hence lead to better prediction of prolonged North American droughts.

6 Conclusions

During the past decade, there is increasing evidence that multidecadal variations in North Atlantic SST (i.e., the AMO) play an important role on severe and persistent droughts in North America. Previous studies are mainly based on observations that just cover one (or two) cycles of the AMO. This study provides a comprehensive investigation of the relationship between the Atlantic SST and persistent drought in North America throughout the Holocene using modern observations, proxy data and model simulations. Our results show that the observed dry/wet fluctuations in the Great Plains and the southwest North America are closely related to AMO. When the North Atlantic is warm (AMO warm phase), the two regions are dry, and vice versa. Such a relationship is also found using proxy SST and the tree ring reconstructed PDSI extended back to 1567 AD, suggesting a robust relationship between the SST in North Atlantic and the drought in the North America.

The close relationship between North Atlantic SST and North American drought are also identified on centennial timescales during the Holocene. The warm North Atlantic during the MWP (AMO-like warm SST anomalies) is closely associated with the severe and persistent mega droughts in the North America. Our results further suggest that the medieval drought is just one of the recent droughts on centennial timescales that affected the North America. Proxy SST records in the North Atlantic and the proxy drought records in the North America showed that the major AMO-like warm SST periods in the last 7.0 ka all correspond to dry conditions in the Great Plains. The cold periods in the North Atlantic also align with wet conditions in the region. These relationships are less pronounced prior to 7.0 ka BP, possibly related to the presence of Laurentide Ice Sheet. In summary, our results suggest that the AMO and AMO-like SST variations in the North Atlantic Ocean are likely a major and consistent driver of the decadal and centennial timescale circulation and drought in the North America during the last 7.0 ka.

The influence of the North Atlantic SST on the North American drought is also examined using the simulations made by five global climate models, all forced by the same sets of SST anomalies associated with AMO. The results show considerable large-scale similarity among the models when forced by warm SST anomalies in the North Atlantic (a pattern similar to AMO warm phase). Despite some regional differences among the models, all models capture the major dry features over the North America and show 0.1–0.3 mm/day precipitation deficit (about 10–20% of the annual average) over the Great Plains and the southwest North America. When precipitation is averaged over each region, all models simulated a significant reduction in

annual precipitation. The average of the five models showed a widespread dry condition over most of the North America, consistent with the statistical relationship using observational and proxy data. Compared to warm North Atlantic, the model responses to cold North Atlantic Ocean (a pattern similar to AMO cold phase) are much weaker and contain greater disagreement among the models, especially over the Great Plains. Though the average of the five models showed slightly wetter conditions over the Great Plains and the southwest North America, there is also considerable model bias on simulating the influence of the cold North Atlantic SST anomalies. Compared to warm North Atlantic, the responses of precipitation to cold North Atlantic are usually small and insignificant in the Great Plains.

The precipitation responses to warm (cold) North Atlantic Ocean also contain strong seasonal variations. There is general agreement among the models in simulating the warm season precipitation when forced by warm North Atlantic SST. The precipitation in the Great Plains and the southwest North America is significantly drier for all the five models during the warm seasons when the North Atlantic is warm. However, there is greater disagreement among the models in other seasons. Nonetheless, when the precipitation anomalies of the five models are averaged, the Great Plains and the southwest North America are dry (wet) on all seasons when the North Atlantic is warm (cold), consistent with the statistical relationship using observational and proxy data.

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