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Variations in North American Summer Precipitation Driven by the Atlantic Multidecadal Oscillation

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ABSTRACT

Understanding the development and variation of the atmospheric circulation regimes driven by the Atlantic multidecadal oscillation (AMO) is essential because these circulations interact with other forcings on decadal and interannual time scales. Collectively, they determine the summer (June, July, and August) precipitation variations for North America. In this study, a general circulation model (GCM) is used to obtain such understanding, with a focus on physical processes connecting the AMO and the summertime precipitation regime change in North America. Two experimental runs are conducted with sea surface temperature (SST) anomalies imposed in the North Atlantic Ocean that represent the warm and cold phases of the AMO. Climatological SSTs are used elsewhere in the oceans. Model results yield summertime precipitation anomalies in North America closely matching the observed anomaly patterns in North America, suggesting that the AMO provides a fundamental control on summertime precipitation in North America at decadal time scales. The impacts of the AMO are achieved by a chain of events arising from different circulation anomalies during warm and cold phases of the AMO. During the warm phase, the North Atlantic subtropical high pressure system (NASH) weakens, and the North American continent is much less influenced by it. A massive body of warm air develops over the heated land in North America from June–August, associated with high temperature and low pressure anomalies in the lower troposphere and high pressure anomalies in the upper troposphere. In contrast, during the cold phase of the AMO, the North American continent, particularly to the west, is much more influenced by an enhanced NASH. Cooler temperatures and high pressure anomalies prevail in the lower troposphere, and a frontal zone forms in the upper troposphere. These different circulation anomalies further induce a three-cell circulation anomaly pattern over North America in the warm and cold phases of the AMO. In particular, during the cold phase, the three-cell circulation anomaly pattern features a broad region of anomalous low-level southerly flow from the Gulf of Mexico into the U.S. Great Plains. Superimposed with an upper-troposphere front, more frequent summertime storms develop and excess precipitation occurs over most of North America. A nearly reversed condition occurs during the warm phase of the AMO, yielding drier conditions in North America. This new understanding provides a foundation for further study and better prediction of the variations of North American summer precipitation, especially when modulated by other multidecadal variations—for example, the Pacific decadal oscillation and interannual variations associated with the ENSO and the Arctic Oscillation.

1. Introduction

Summer precipitation in North America provides most of the water needed for both natural and managed ecosystems in the region. These ecosystems suffer, however, as do their services to the communities and societies when large, anomalous fluctuations occur in summer precipitation. Several of these anomalies were observed in the last 10 years, such as the consecutive

summer droughts from 2000–05 in the central and western United States, the 2008 summer flooding in the Ohio River valley, and the May–June 2010 flooding in the central and southern United States. The impacts of these anomalies in summer rainfall can be mitigated if they can be predicted with sufficient lead time and accuracy. Gaining this predictive capability requires causal understanding of summer precipitation variations at a wide range of time scales as interactions among processes across these time scales ultimately result in the observed interannual fluctuations in the precipitation.

Prior studies have examined several aspects of variations in summertime precipitation at different time

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scales and attributed possible causes. At the *seasonal* time scale, transient eddy processes rising from atmospheric circulation anomalies due to topography and surface heterogeneity in North America cause changes in the westerly jet stream intensity and position. This has implications for severe summer precipitation anomalies, including droughts and floods (e.g., Trenberth and Guillemot 1996; Mo et al. 1995). In addition, anomalies in soil moisture caused by antecedent precipitation anomalies may affect surface fluxes of water and energy. These in turn have been shown to contribute to summer precipitation fluctuations severe enough to either prolong drought or cause regional flooding (Oglesby and Erickson 1989; Oglesby et al. 2002; Hong and Kalnay 2000; Koster et al. 2009). These large fluctuations in summertime precipitation presumably owe their initial development to anomalies in the large-scale atmospheric circulation over North America (e.g., Hu and Feng 2004a,b, 2008). These circulation anomalies evolve from intricate interactions between physical processes arising from interannual and multidecadal time scale forcing that can also be of global scale.

Among variations on *interannual* time scales that strongly influence North American precipitation are the El Niño–Southern Oscillation (ENSO) (e.g., Ropelewski and Halpert 1986; Trenberth and Guillemot 1996; Ting and Wang 1997; Hu and Feng 2001a), the Arctic Oscillations (AO; Thompson and Wallace 2000), and the North Atlantic Oscillation (NAO, Hurrell 1995). These sources of interannual forcing primarily perturb the latitudinal position and strength of the upper-troposphere westerly jet stream and the associated transverse or meridional circulation (e.g., Trenberth and Guillemot 1996; Mo et al. 1995; Hu and Feng 2010). A recent study by Hu and Feng (2010) showed that in the central United States, the westerly jet shifts to higher latitudes during the positive phase of the AO as a result of eddy heat and momentum forcing on the mean zonal flow in the upper troposphere. The associated transverse circulation anomaly has a downward motion over the central United States, encouraging low-level divergence and suppressing summer precipitation development. Meanwhile, influences of the AO are different from that of ENSO, having a competing effect on the jet stream and regional circulation, complicating precipitation variations in the central United States (Hu and Feng 2010).

While these interannual variations affect the regional circulation and precipitation in North America, their specific effects are realized only in particular atmospheric circulation environments. Such large-scale circulation environments are heavily influenced by *multidecadal* time-scale variations in the oceans and their interactions with the atmosphere. The Pacific decadal oscillation

(PDO) in the Pacific Ocean (Zhang et al. 1997), for example, has been found to strongly influence ENSO effects on North America. During the warm phase of the PDO, the ENSO effects on North American summer precipitation weaken and are essentially eliminated in some regions. Conversely, the effects of ENSO strengthen during the PDO cold phase (Gershunov and Barnett 1998; Hu and Feng 2001a). By regulating the effects due to ENSO and other factors, such as the persistence of the SST in the North Pacific (Hu and Feng 2004a), the PDO can explain about 24% of the variance in summer droughts in the United States for the past 100 years (McCabe et al. 2004). Another 28% of the drought variance in the region can be explained by the Atlantic multidecadal oscillation (AMO) (McCabe et al. 2004).

The AMO is defined as the basin-scale (from 0°–60°N) SST anomalies in the North Atlantic (Mestas-Nunez and Enfield 1999; Enfield et al. 2001). Recent observational studies of the effects due to the AMO (e.g., Hu and Feng 2008) show that during different phases of the AMO, different regional circulation regimes prevail in North America, affecting especially the North American summer monsoon (which is defined by rainfall and its variations in the southwestern United States and northern Mexico). Prevailing winds and persistent pressure anomalies in North America during the warm or cold phase of the AMO favor specific rainfall anomaly patterns across the region. For example, during the cold phase of the AMO, more frequent northwesterly wind anomalies in the North American monsoon region confine the monsoon rainfall to the south of the southwestern United States (Hu and Feng 2008). Meanwhile, a strong southerly low-level flow from the Gulf of Mexico, associated with the sea surface temperature (SST) anomalies in the western tropical Atlantic Ocean, enhanced moisture transport and circulation anomalies that encourage above-average summer precipitation in the central United States (also see Wang et al. 2006, 2008; Feng et al. 2008, 2011). This anomaly pattern changes during the warm phase of the AMO, resulting in below-average precipitation for most of central North America. These changes effectively describe an alternating monsoon regime that follows the phases of the AMO (Hu and Feng 2008).

These observational findings of the direct influence of the AMO on the North American monsoon regime indicate that differing and persistent large-scale circulation regimes exist at the multidecadal time scales in association with multidecadal variations of SST in both the North Pacific and the North Atlantic Ocean. These regimes favor specific precipitation anomaly patterns in North America and also condition particular circulation environments for shorter-term interannual time-scale

forcings, for example, ENSO and AO, to further affect seasonal precipitation. These interannual scale forcings could in turn feedback to the multidecadal effect. Collectively, they, along with local surface effects and interactions, determine the precipitation distributions in North America. Hence, understanding the development and persistence of these circulation regimes during different phases of the AMO and how these regimes and associated summer rainfall patterns are modulated by the higher-frequency forcings is essential to improving our understanding and hence prediction of summer precipitation variations in North America.

Among the efforts toward understanding the AMO effects, Sutton and Hodson (2005, 2007) used the U.K. Hadley Center Atmospheric Model version 3 (HadAM3) and simulated global circulation and precipitation responses to SST forcing in the North Atlantic. Their results show that the basin-scale SST anomalies during the AMO have the most significant influence on North American precipitation during boreal summer (June–August), with more (less) precipitation during the cold (warm) phase of the AMO. The general agreement of these model results with observations has led to the suggestion that changes in North Atlantic SST associated with the AMO may have played a “central role” in forcing the observed multidecadal changes in summertime circulation in Europe and North America (Sutton and Hodson 2007).

In recognition of the important role of the AMO in persistent summertime circulation and precipitation anomalies in North America, the U.S. Climate Variability and Predictability (CLIVAR) drought working group recently organized an effort to facilitate understanding of the SST forcing associated with the AMO, and PDO, on development of prolonged droughts in North America (Schubert et al. 2009). While the current focus of this effort remains on the Pacific SST effect, the initial results have shown consistent AMO effects on decadal-scale variations in North American summertime precipitation (Schubert et al. 2009; Mo et al. 2009; Kushnir et al. 2010; Feng et al. 2011). All five general circulation models (GCM) used by Schubert et al. (2009) simulated less (more) annual mean precipitation in North America during the warm (cold) phase of the AMO, albeit the magnitudes of the anomalies vary among the models (see their Fig. 8). While these and the results from previous modeling studies of climate responses to the AMO are consistent with observations (e.g., Enfield et al. 2001; Hu and Feng 2010; Feng et al. 2011) the *physical processes behind these results remain to be understood*.

In this study, we use a general circulation model to evaluate the North American summer season circulation and precipitation anomalies forced by SST anomalies associated with the AMO, focusing on understanding the

physical processes connecting the AMO forcing with the precipitation anomalies in North America. This study extends previous studies that focused on the atmospheric response to interannual time-scale SST variations in the North Atlantic (e.g., Kushnir et al. 2010), and provides specific and detailed understanding of the role of the AMO on multidecadal time-scale variations in summertime circulation and precipitation in North America. By doing so, this study also helps provide a foundation for further investigation and understanding of the variations in North American summertime precipitation modulated by other multidecadal-scale forcing, for example, the PDO, as well as interactions with interannual variations associated with the ENSO and the AO. Details of the model are described in the next section (section 2) along with the model experiments. Major results of these experiments are discussed in section 3. Based on these results a mechanism is proposed in section 4 to explain the circulation and precipitation anomalies in North America that occur during different phases of the AMO. The significance of the AMO effects in the context of multidecadal variations in North American summer precipitation and the importance of understanding these effects for improving the precipitation prediction at seasonal to interannual time scales are discussed in section 5.

2. Model and experiments

a. Model

We use the Community Atmospheric Model version 3.1 (CAM3.1; Collins et al. 2006), developed at the National Center for Atmospheric Research (NCAR). The CAM3.1 includes the Community Land Model version 3 (CLM3). The model configuration for this study has 26 levels in the vertical and a 42-wave triangular spectral truncation (equivalent to 2.8-degree resolution in latitude and longitude). The sea surface temperatures are prescribed in the model using observed climatological SST, to which we add specific anomalies to make up the model experiments. The land surface vegetation parameters/distributions, concentration of atmospheric greenhouse gases, and solar constant all remain at current conditions during the model runs.

b. Model experiments

The model experiments were designed to identify and quantify the forcing of North Atlantic SST anomalies on summer season atmospheric circulation and precipitation in North America. The “control run” uses the monthly global climatological SST field averaged over the period from 1871–2008. These SST data were obtained from the merged monthly mean U.K. Hadley Centre sea ice and SST dataset version 1 (HadISST1)

and version 2 of the U.S. National Oceanic and Atmospheric Administration (NOAA) weekly optimum interpolation SST analysis (Hurrell et al. 2008). We use this merged SST product, which has spatial resolution of 1.0×1.0 degrees in latitude and longitude, because it is updated monthly and readily available to the modeling community. The same monthly global climatological SST is used for every year in the 50-yr control run. Results of the control run are used as the reference by which we quantify the forcing of the North Atlantic SST anomalies on circulation and precipitation in North America.

Two experimental runs are conducted with contrasted SST anomalies imposed in the North Atlantic Ocean designed to represent the warm and cold phases of the AMO. To develop the SST anomalies representing the warm (cold) phases of the AMO, we first identified those years that had the annual AMO index (averaged SST over 0° – 60° N, 7.5° – 75° W, see also Enfield et al. 2001) in the warmest (coldest) 25% (quartile) for the period 1871–2008. Corresponding SST anomalies during those warmest (coldest) 25% years from the mean of 1871–2008 at each model grid point in the North Atlantic Ocean were calculated. The magnitudes of these SST anomalies were then inflated by a factor of 2 to amplify signal to noise ratios and hence allow for a clearer dissection of causal mechanisms. [A similar scaling method also was used in Schubert et al. (2009) and others, for a similar purpose.] The derived SST anomaly fields in the North Atlantic Ocean for the AMO warm and cold phases are shown in Fig. 1.

In both experiments, the (invariant) SST anomalies in the North Atlantic Ocean are imposed in every month and year, and the CAM3.1 is integrated for 20 years. Climatological SSTs were imposed elsewhere in the oceans. As the other model parameters in the experiments are the same as in the control run, differences between the experiments and the control run will show summertime circulation and precipitation driven by the AMO SST anomalies.

In this study, the differences between the model years 7–20 in the experimental runs, after removing the first six years for spin up, and the 50-yr control run, are analyzed. Comparisons between the 7–20 and 50-yr control run results showed little differences in both the mean and variance of seasonal circulation (geopotential height) and precipitation; statistical tests identified no significant differences between these results. These test results warrant the use of the 20-yr simulations in the experiments.

Because the use of specified invariant SST anomalies in these experiments prevents feedbacks/interactions of atmospheric responses to the SST variations, model results

FIG. 1. (a) North Atlantic Ocean surface temperature anomalies in the warm phase of the AMO, and (b) as in (a), but for cold phase (unit: $^{\circ}$ C). The SST anomalies are inflated by 2 times to amplify the signal to noise ratio and allow for clear dissection of causal mechanisms.

contain no transient effects from the lower boundary and will describe only the forcing effects of the specified boundary conditions (Bretherton and Battisti 2000). A coupled ocean–atmosphere model would be required to evaluate the more complete effect of the boundary forcing evolved from transient interactions between the atmosphere and the ocean. Another limitation of these experiments is that they are from a single atmosphere model, and model biases could have potential impact on the results of these experiments. Prior studies comparing multiple GCMs (e.g., Schubert et al. 2009; Mo et al. 2009) have shown results of the CAM3 model that were consistent with other models in simulations of forcing effects from the North Atlantic and the Pacific, lending support to the notion that the CAM3 model results can provide indications of effects and processes that are important for the AMO to influence the summertime circulation and precipitation in North America (Sutton and Hodson 2007). Nevertheless, we caution that more detailed studies are necessary to further validate these findings using a single model.

FIG. 2. Observed and simulated mean state of circulation in the lower and upper troposphere for warm and cold phases of the AMO.

3. Results

Before discussion of the results from model experiments it is necessary to demonstrate that the model adequately simulates the mean state of the atmospheric circulation for the period of the study (1949–2000). Figure 2 shows the simulated geopotential heights at 850 and 300 hPa and their comparisons with the National Centers for Environmental Prediction (NCEP)–NCAR reanalysis data. These comparisons show similar geopotential distributions in both the lower and upper troposphere, indicating similar wind and temperature configurations and thus a consistent mean state circulation between the reanalysis data and the model simulation. Such a consistent mean state in the model is sufficient for its use in this investigation, which focuses on anomalies in North America rising solely from SST forcing of the AMO.

Results of the June–August (JJA) precipitation for North America and surrounding oceanic areas are shown in Fig. 3. Figure 3a shows the JJA rainfall (in mm day^{-1}) in the 50-yr control run. More precipitation is shown in the eastern and southeastern United States and less precipitation in the west, particularly the southwestern United States. The central Great Plains receives more precipitation than the northern plains and the Great Lakes area, partially because of converging low-level moisture flows from the Gulf of Mexico as suggested by

the surface 700-hPa moisture fluxes (shown by the arrows in Fig. 3). This simulated climatological distribution of JJA precipitation captures the major features of the observed JJA precipitation (e.g., Feng et al. 2008), a result further supporting the ability of the model to describe summertime precipitation in North America.

Anomalies in JJA precipitation from experimental runs for the AMO warm and cool phases relative to the 50-yr control run are shown in Figs. 3b and 3c, and their differences (from each other) are shown in Fig. 3d. These anomalies were averaged over years 7–20 in the experimental runs, after removing the first six years for spinup. In the warm phase (Fig. 3b) most of the United States shows reduced rainfall. A large dry region covers the U.S. Midwest and central and southern Great Plains. The driest area is in southern Texas and Mexico along the eastern slope of the Sierra Madre Oriental. Large positive anomalies, representing increases in summertime precipitation, are shown in the North American monsoon region and over the low-latitude oceans.

A different pattern of summer rainfall anomalies is found for the cold phase of the AMO (Fig. 3c), not quite a mirror image of that for the warm phase (Fig. 3b) however. In Fig. 3c, increased precipitation is seen in most areas of the United States and in northern Mexico, except for a strip of negative anomalies from the south-central to the northeastern United States sandwiched between

FIG. 3. Modeled JJA precipitation (contours, units: mm day^{-1}) and the surface 700-hPa moisture flux (arrows, units: $\text{kg m}^{-1} \text{s}^{-1}$) for (a) control run, (b) anomalies in warm phase of the AMO, (c) anomalies in cold phase of the AMO [negative anomalies are shaded in (b),(c)], and (d) difference of warm and cold phases. Shading in Fig. 3d indicates differences in precipitation significant at the 95% confidence level in a two-tailed Student's t test.

positive anomalies to the east and the west. Below-average rainfall occurs in central Canada, stretching down to the western United States. Low-latitude regions, south of 30°N , are generally drier. It is interesting to note that the North American monsoon region has near-average rainfall during the cold phase of the AMO. The asymmetry of this response between the AMO cold and warm phases, also discussed in Sutton and Hodson (2007) and Schubert et al. (2009), suggests that the strong wet summer monsoon during the warm phase of the AMO may be a primary reason for the multidecadal time-scale variation in the North American summer monsoon regime, as shown in Hu and Feng (2008).

A comparison of the JJA rainfall anomalies in North America between the warm and cold phases of the AMO is presented in Fig. 3d. Regions with negative values in Fig. 3d show reduced summer rainfall in the warm phase of the AMO and enhanced rainfall in its cold phase. Regions with positive values indicate drier conditions in the cold phase than in the warm phase of the AMO. A scrutiny of these differences indicates that except for central Canada and the North American monsoon region, the North American continent receives more summer rainfall during the cold phase of the AMO than in the

warm phase. The comparison in Fig. 3d and precipitation anomalies in Figs. 3b and 3c also show that the largest precipitation anomalies in both warm and cold phases of the AMO are in the low-latitude and tropical oceans, consistent with Sutton and Hodson (2007).

A key question is how these summertime precipitation anomalies driven solely by the AMO forcing may describe the actual, observed AMO effects on precipitation anomalies. To address this, we first evaluate the correlation of the observed summer rainfall anomaly with the SST anomalies in the North Atlantic Ocean and then compare and contrast the observed correlation pattern with that obtained from the model results (Fig. 3). The observed correlation will show the portion of the North American summer rainfall variations that is (statistically) associated with the AMO forcing in the North Atlantic. This observed correlation pattern is shown in Fig. 4a, obtained from analysis of the observational data for the period 1901–2006. An inspection suggests that this correlation pattern is very similar to that shown in Fig. 8b of Hu and Feng (2008). Negative correlations occur in most of North America, except for southeastern United States, the North American monsoon region, and a region stretching from the U.S. northern Great Plains to the

FIG. 4. Regressions between JJA AMO and JJA daily rainfall in North America for (a) observation and (b) model simulation. Shading indicates significant correlations at the 95% confidence level. Observational data are from Climatic Research Unit at University of East Anglia (New et al. 2000).

Great Lakes area. These negative correlations indicate below (above) average summer rainfall in most of the United States in accordance to warmer (cooler) SST in the North Atlantic Ocean. The largest negative correlations are in southern Texas and eastern Mexico and also the U.S. central Great Plains.

The correlation pattern in Fig. 4a is fairly comparable with the modeled relationship in Fig. 4b (a duplicate of Fig. 3d for easy comparison). Figure 4b shows negative precipitation anomalies in most of North America associated with warmer SST in the North Atlantic Ocean, and positive precipitation anomalies in North America corresponding to cooler SST in the North Atlantic. In detail, large negative anomalies occur during the AMO warm phase along the eastern slope of the Rockies and most of the Great Plains, with the largest anomalies in southern Texas and northeastern Mexico. Large negative anomalies

also are observed from the Great Plains eastward through most of eastern United States and Canada, except for the southeastern United States. Negative anomalies are seen in the northwestern United States and western Canada. On the other hand, during the cold phase of the AMO, positive anomalies are observed in the North American monsoon region and the central region of North America along the U.S.–Canadian border. Interestingly, even the observed small region of weak negative anomalies in the strip from eastern Nebraska to eastern Oklahoma is found in the model results. A pattern correlation analysis between Figs. 4a and 4b for the region of 65°–125°W and 24°–50°N yielded a score of 0.2, which is statistically significant at the 99.99% confidence level, indicating that the simulated summer rainfall anomalies in North America forced by the AMO in Fig. 4b (Fig. 3d) resemble well the observed pattern of associations between the summertime precipitation in North

America and the SST anomalies in the North Atlantic at a multidecadal time scale.¹

Comparisons of Figs. 4a and 4b also indicate some important differences between the observed and simulated summertime precipitation forced by the AMO. Besides the difference in south-central Canada, strong differences are shown in the low-latitude regions, although no suitable observations are available for tropical ocean areas. The simulated summer rainfall anomalies in low-latitude Mexico are much larger than the regressed rainfall and the sign also is opposite to that in the tropical Atlantic and the Caribbean. This discrepancy suggests that other forcings may have played important roles in affecting the tropical precipitation variations or simply that the multidecadal variations may be a less important component in the variations of tropical precipitation during boreal summer.

While discrepancies exist in the tropical regions, it is intriguing that the simulated summer rainfall anomaly pattern in North America captures the general features of observed distribution of the rainfall anomalies associated with the SST anomalies of the AMO, for the model only has SST anomalies in the North Atlantic Ocean. This similarity therefore shows a deterministic and potentially the “central role” for North Atlantic SST associated with the AMO on North American summertime circulation and precipitation variations (Sutton and Hodson 2005, 2007). The North Atlantic SST anomalies associated with the AMO force large-scale summer circulation anomalies in North America that favor specific precipitation patterns. Accordingly, warm SST anomalies in the North Atlantic during warm phase of the AMO would result in less summer rainfall in most of North America. SST anomalies during the cold phase of the AMO would induce circulation anomalies featuring more summer rainfall for most of North America. In other words, drier conditions would be more frequent in summers in North America during the decades of the warm phase of the AMO while more rainfall could be expected in most summers when the AMO is in a cold phase. Exceptions to this relationship are found in the southeastern United States, the North

American monsoon region, and the region stretching from the northern Great Plains to the Great Lakes. In these regions either little change occurs with AMO anomalies, or the response is the opposite of that described above.

4. A mechanism for the AMO forcing on North American summertime precipitation

To further understand how the AMO affects North American precipitation, and identify the physical processes involved, we examine the mass and wind fields from the CAM3.1 model experiments. Figure 5 shows the sea level pressure (SLP) fields from the control and experimental runs. Comparing the SLP anomalies in the warm and cold phases of the AMO (Fig. 5b versus Fig. 5c), we find the largest differences in the North Atlantic subtropical high pressure system (NASH). Compared to the control run (Fig. 5a) the NASH enhanced and shifted its center westward by nearly 20° in the cold phase of the AMO (Fig. 5c). Meanwhile, high pressure anomalies extended westward into the eastern subtropical and part of the northern Pacific. It is important to note that the western portion of the NASH is extended considerably in the meridional direction. All of North America is effectively under the influence of the western part of this enhanced high pressure system.

On the other hand, during the warm phase of the AMO (Fig. 5b), the NASH contracts substantially, as its center shifts eastward by about 20° and also northward by 20°. Strong negative SLP anomalies occur over the subtropical North Atlantic, extending to the eastern subtropical Pacific. Negative SLP anomalies also occur over most of North America.

An interpretation of these changes in terms of the low-level mass field is that when the SST in the North Atlantic Ocean are warmer than average, the (warmer) air mass would expand and spread primarily downstream to northeastern portions of the North Atlantic region. The resultant decrease in air mass would result in lowering the SLP and contracting the size of the high pressure system. This contraction in the NASH would leave the land areas in North America, especially mid-latitude regions, to more freely develop their own pressure anomalies. Because heated land areas in summer favor low pressure, low pressure anomalies tend to develop and prevail in North America following the contraction of the NASH. This notion is supported by the much warmer surface air temperatures, shown by the temperature anomalies in Fig. 6a and lower SLP in Fig. 5b in summers during the warm phase of the AMO.

The negative SLP anomalies in North America and North Atlantic during the warm phase of the AMO

¹ In calculating this score, we first interpolated the observed pattern in Fig. 4a, which is on a $0.5^\circ \times 0.5^\circ$ mesh, and the model result in Fig. 4b, which has a T42 resolution (2.8125° grid system), to a common $1.0^\circ \times 1.0^\circ$ grid system, which has 1132 grid points over the study domain. Over these grids we calculated the pattern correlation. While this score is low in value it is highly significant in a statistical sense because of the correlation over a large number of samples. When focusing on a smaller region in the Great Plains (90° – 110° W, 30° – 50° N) the pattern correlation score is 0.615, again significant at the 99.99% confidence level.

FIG. 5. (a) SLP for control run, (b),(c) SLP anomalies averaged for warm and cold phase of the AMO, respectively (negative anomalies are shaded), and (d) difference of (b),(c). The arrows show the surface wind speed (m s^{-1}). Shading in (d) indicates differences in SLP significant at the 95% confidence level in two-tailed Student's *t* test.

appear further modified by regional changes and contrasts in topography and land surface conditions. As a result, a three-cell SLP anomaly pattern is found, comprising anomalous low pressure over North America, a low pressure anomaly in the subtropical North Atlantic, particularly the warm pool region in the western subtropical North Atlantic and the Caribbean (Wang et al. 2006, 2008), and a low pressure anomaly centered in the eastern subtropical Pacific. These anomaly cells also extend into the lower troposphere, as shown in the mass and wind anomalies in the lower troposphere (850 and 700 hPa) during the warm phase of the AMO (similar to that shown in Fig. 5 for the SLP anomalies).

During the cold phase of the AMO, a different three-cell pattern is observed in the lower troposphere (as indicated in Fig. 5c). An anomalous low pressure cell remains over North America though with a more north-south orientation from that in the AMO warm phase. The two cells in the subtropical North Atlantic and the eastern subtropical Pacific are reversed from that in the warm phase, with high pressure anomalies in the SLP and in the lower troposphere. A plausible explanation for the persistent relative low pressure anomaly in SLP and the lower troposphere in the western United States in both phases of the AMO is the strong “heat island”

effect of the Rockies (Fig. 6c), which provides an elevated summer heat source in the lower troposphere. This orography and surface heterogeneity clearly plays an important role in the regional circulation during boreal summer.

A schematic summarizing these three-cell structures, associated circulation anomalies, and their changes between the cold and warm phases of the AMO is shown in Figs. 7a and 7b. The circulation anomalies summarize the responses of the mass and wind fields in the lower troposphere to the SST forcing in the North Atlantic during the AMO. Particularly, during the AMO warm phase (Fig. 7a) the cyclonic anomaly over the North Atlantic warm-pool region weakens the clockwise rotation of low-level winds around the NASH. This also induces or enhances easterly onshore flows into the southeastern United States, favoring summertime precipitation in that region during the warm phase of the AMO (hatched area in Fig. 7a), as also observed in Hu and Feng (2008).

The anomalous cyclonic cell over the eastern subtropical Pacific during the AMO warm phase has anomalous southeasterly flows along western Mexico and Baja California. These anomalous flows weaken the northerly and northwesterly flow along the coast of southwestern North America that otherwise occurs

FIG. 6. Anomalies of temperature at 2 m above the surface in response to (a) warm SST and (b) cold SST during the AMO (unit: $^{\circ}\text{C}$; negative anomalies are shaded).

during “average” climate conditions, allowing warmer and moister air to be advected/transported to the North American monsoon region. This helps establish strong positive anomalies in monsoon rainfall (hatched area in Fig. 7a, also see Fig. 3b, and discussions in Hu and Feng 2007, 2008).

North of this cell, the low pressure anomaly in the western and northern U.S. Great Plains induces anomalous southerly flow through the central and north-central United States. This anomalous low-level southerly flow, in conjunction with the northerly anomaly flow in the cell in western subtropical Atlantic, allows strong low-level divergence anomalies, which would favor less rainfall for the central and south-central United States.

During the cold phase of the AMO, the entire western Northern Hemisphere is under the influence of a strengthened NASH (Fig. 5c), and North America is embraced within the western part of this enhanced high pressure

system. As summarized in Fig. 7b, the stronger easterly flow along the southern flank of the enhanced NASH created two anticyclonic anomaly cells (separated by the Sierra Madre Mountains)—one in the western subtropical North Atlantic and the other in the eastern subtropical Pacific. North America overall has positive pressure anomalies in the surface and lower troposphere. The warm surface temperatures of the elevated terrain of the Rockies, however, induce a relative low pressure area (thermal or lee trough) with a north–south orientation over the Rocky Mountains and the eastern foothills. Along the eastern fringe of this relative low pressure region (still within the high pressure anomalies in North America, see Fig. 5c), southerly wind anomalies occur. These anomalies help extend the enhanced southerly flow from the Gulf of Mexico northward (along the western branch of the cell extending from the subtropical western Atlantic), opening a channel for low-level moisture flow from the Gulf of

Mexico to the central and northern United States. The enhanced moisture flow and convergence into the central and north United States provide uplift mechanisms that favor storm development and encourage positive rainfall anomalies in the central United States during the cold phase of the AMO. In the North American monsoon region, the anticyclonic anomaly in the lower troposphere of the eastern subtropical Pacific (Fig. 7b) enhances westerly and northwesterly flow into the region, helping suppress rainfall development. Overall, these mass and wind anomalies in the cold phase of the AMO lead to precipitation anomalies matching well with the observed spatial distribution of summer rainfall anomalies in North America (Hu and Feng 2008).

Consistent with the circulation anomalies in the lower troposphere, the mass and wind fields in the upper troposphere also show a three-cell anomaly pattern over North America and adjacent subtropical oceanic regions. The three cells, however, have differing anomaly signs in the upper troposphere (Fig. 8), which indicate differences in the strength and profile of these systems in the vertical. Examining the pressure and wind anomalies at 300 hPa in the warm phase of the AMO (Fig. 8b), we find one cell of high pressure (anticyclonic) anomaly over most of North America in the upper troposphere. There are two, much weaker, low pressure anomaly cells, one in the subtropical North Atlantic, stretching to the Gulf of Mexico and the Caribbean, and one over the eastern subtropical Pacific and southwestern North America. Major circulation features from this three-cell anomaly pattern are strong easterly anomalies in midlatitude North America and southerly anomalies in the western United States.

The circulation anomalies in the cold phase of the AMO are nearly the reverse of those in the warm phase. As shown in Fig. 8c, the anomaly pattern at 300 hPa in the AMO cold phase has low pressure anomalies in high-latitude North America (north of 55°N), with a trough extending down to the Great Lakes area. In lower latitudes, a broad region of low pressure anomalies occurs in the eastern subtropical Atlantic and the Caribbean Sea. Between these two low pressure anomaly regions are strong high pressure anomalies in the midlatitudes across the central United States. A center of these high pressure anomalies is found in the western United States, indicating intense mass buildup over midlatitude North America during the cold phase of the AMO. These mid- and upper-troposphere mass and wind anomalies during the warm and cold phases of the AMO are summarized in the schematics in Figs. 7c and 7d.

These contrasting circulation anomalies in the mid and upper troposphere between the cold and warm phases of the AMO help account for the differing precipitation anomalies and their spatial variations over

North America. During the cold phase of the AMO, a zone of strong horizontal shear occurs across midlatitude North America. On the northern and southern sides of this shear zone the air masses are quite different, with an oceanic origin for the air on the south and more continental origin for the air to the north. These air masses interact actively across this upper-level frontal zone. A key aspect of this exchange is a strong northerly flow anomaly over the central United States (Figs. 8c and 7d). This northerly flow anomaly is a “classical” feature related to above average precipitation in the central United States (Ting and Wang 1997; Hu and Feng 2001a,b, 2008, 2010). This anomalous flow favors advection of air from high-latitude regions into the middle and low latitudes, creating an environment conducive for active and severe storms in those regions. Hence, more precipitation in the central and northeastern United States is favored during the cold phase of the AMO.

Such an upper-level frontal zone and associated air-mass interactions are absent during the warm phase of the AMO. As summarized in Fig. 7c, the flow anomalies in the warm phase have an anticyclone cell over most of North America, with easterly flow anomalies found in the 30°–40°N latitude for most of North America. The relatively stable air mass of continental origin over divergent flow anomalies in the lower troposphere (Figs. 7a and 5b) yields an environment conducive to fewer storms and less precipitation in most of North America.

These contrasting anomalous mass and circulation patterns in the upper troposphere between the warm and cold phases of the AMO also are apparent in the 300-hPa zonal wind variations shown in Fig. 9. By comparison to the 300-hPa mean zonal flow in the control run (Fig. 9a), the 300-hPa circulation anomaly during the cold phase (Fig. 9c) shows a northward shift of the midlatitude westerly jet over North America (also see Fig. 7d versus Fig. 7c). The shift is largest in the central part of the continent, consistent with the enhanced NASH system and its westward expansion. Along with this shift and enhanced NASH, the zone of strong shear in the zonal flow moves northward into midlatitude North America, extending from the west-central United States to the northeastern United States. The large northward shift of the shear zone in the central and eastern part of the continent also brings cyclonic curvature in the jet stream over the region. This combination of strong shear in the zonal wind and the cyclonic curvature marks an active upper-tropospheric front from the west-central and central United States to the northeastern United States, favoring development of storms in this region during the cold phase of the AMO.

a) Three-cell anomalous circulation in lower troposphere during warm phase

b) Three-cell anomalous circulation in lower troposphere during cold phase

FIG. 7. Schematic summary of pressure and flow anomalies (the three-cell anomalous circulation) in the lower troposphere during the (a) warm and (b) cold phase of the AMO and in the upper troposphere during the (c) warm and (d) cold phase of the AMO. The hatched areas have above average summer (JJA) precipitation and the dotted areas have below-average summer precipitation. The double line in (c),(d) indicates the upper-troposphere front.

In the warm phase of the AMO the upper-troposphere zonal wind anomalies over the North America weaken the shear in the westerly jet, without causing much change to the jet position (comparing Fig. 9c versus Fig. 9a). The strong shear zone stays across the southern United States north of the jet (from about 35°N), and weak anomalies in the zonal wind contribute to only weak activities in storm development in the region, which contrast to the active storms in the cold phase of the AMO.

5. Summary and concluding remarks

We made simulations using the NCAR CAM3.1 model with specified SST anomalies associated with the Atlantic multidecadal oscillation in the North Atlantic Ocean

and climatological SSTs imposed elsewhere in the oceans. The results yield summertime (JJA) precipitation anomalies in North America that closely match observed anomaly patterns (Fig. 4). This suggests that the model can adequately describe the effects of North Atlantic SST anomalies associated with the AMO on North American summertime precipitation at multidecadal time scales. Furthermore, it strengthens the emerging conclusion that the AMO provides a fundamental control on summertime precipitation for North America on decadal time scales (Sutton and Hodson 2007). With this result it becomes possible to investigate the processes and physical chain of events by which the AMO SST anomalies can induce specific summertime circulation and precipitation anomalies in North America.

FIG. 7. (Continued)

During the warm phase of the AMO, the summertime North Atlantic subtropical high pressure system (NASH) weakens. The NASH contracts and shifts northeastward to higher latitudes over the eastern North Atlantic Ocean. The weakened influence from the NASH on North America allows enhancement of thermal low pressure systems over the heated lands during these summer months. Meanwhile, a diminishing influence from the NASH on the eastern tropical and subtropical Pacific also favors regional development of low pressure anomalies.

Additional details further reveal a three-cell circulation low pressure anomaly pattern, with one cell centered in the western United States, one in the western subtropical North Atlantic, and the third in the eastern subtropical North Pacific. Around these cyclonic anomaly cells, wind anomalies include increased easterly and northeasterly onshore flow to the southeastern United States, more southerly flow along western Mexico and Baja California, and a divergent flow pattern in the

central and western United States. Accompanying these anomalies are increased summer rainfall in both the southeastern United States and in the North American monsoon region, but decreased precipitation in most of the rest of North America, particularly the central United States.

In the cold phase of the AMO, the NASH is enhanced and shifts 20° west of its climatological position. A key feature of this enhanced NASH is its westward expansion. This expansion is further accompanied by a substantial longitudinal spread of the strengthened NASH into the mid- and high-latitude western North America and the eastern North Pacific. These pressure anomalies from the enhanced NASH increase the surface and low-level pressure over the continent, diminishing effects of the processes that would otherwise have developed low pressure over the heated land during boreal summer. Associated with these pressure anomalies are anticyclonic flow anomalies around North America. Embedded in the anomalies are two regional, anomalous

FIG. 8. As in Figs. 5a–c, but for 300-hPa geopotential height ($\text{m}^2 \text{s}^{-2}$) and winds (m s^{-1}).

anticyclonic cells, one in the eastern subtropical Pacific and the North American monsoon region and the other in the western Atlantic warm-pool region. Both occur along the southern boundary of the enhanced NASH. Over North America, the elevated heating of the Rockies induces a north–south distortion in the pressure anomalies. These (three-cell) circulation anomalies feature a broad region of anomalous southerly flow from the Gulf of Mexico to the northern U.S. Great Plains (Fig. 7b). The anomalies in lower-tropospheric mass (pressure and temperature) and wind fields are consistent with observed anomalies in corresponding fields during the cold phases of the AMO (Hu and Feng 2008).

The physical processes that lead to these mass and circulation anomalies in the lower troposphere also are apparent in the middle and upper troposphere. In particular, the *key difference* between the anomalies during warm and cold phases of the AMO is that the North American continent is left out from the strong influence

FIG. 9. (a) Average 300-hPa zonal wind from the control run. (b),(c) Anomalies of 300-hPa zonal wind for warm and cold phases of the AMO, respectively (units: m s^{-1} , negative anomalies are shaded).

of the NASH in the warm phase of the AMO, whereas it is much more under the influence of the NASH during the cold phase of the AMO. Specifically, the strong contraction and northeast shift of the NASH during the warm phase of the AMO allows continental air masses to develop and influence North America. From this process evolves one massive body of warm air of continental origin over North America in June–August. In correspondence, high temperature and low pressure anomalies prevail in the lower troposphere, and high pressure anomalies form in the upper troposphere along the east coast of the continent.

In the cold phase of the AMO, a strengthened NASH, particularly its westward and meridional expansions into western North America and the eastern North Pacific, brings maritime air into mid- and high-latitude North America. As a consequence, a strong upper-troposphere front develops between this air mass and the more continental air pushed to the north. Across this front, northerly anomalies over northern and north-central

United States bring high-latitude upper-tropospheric air with presumptively high potential vorticity into mid-latitude North America. It is this northerly anomaly and its advection of air from high latitudes that may help initiate development of instability and associated storms in the central and portions of western United States during the cold phase of the AMO. This interpretation is supported by numerous previous observational studies showing that strong northerly anomalies in the mid and upper troposphere over central and northern U.S. Great Plains are a key circulation feature for development of above-average summer rainfall in the central United States (e.g., Ting and Wang 1997; Hu and Feng 2001a,b, 2008). In a continuing study we are applying a mesoscale model and examining the advection processes of higher PV by this northerly anomaly into the northern and central Great Plains and its role in precipitation anomalies in the central United States.

Finally, results of this study show that the atmospheric processes forced by North Atlantic SST anomalies associated with warm and cold phases of the AMO induce differing circulation anomalies over North America. The similarity of these anomalies to the observed variations associated with the AMO strongly implies a substantial role of the AMO in determining multidecadal time-scale variations in North American summertime circulation and precipitation. Results of this study enhance the notion that the AMO plays a “central role” in forcing the observed multidecadal changes in summertime circulation over North America. Furthermore, our understanding of these AMO effects provides a basis for future investigations of how other SST variations that operate on interannual time scales, for example, ENSO, may modify the AMO-forced circulation anomalies at interannual time scales (or vice-versa). Additionally, further understanding of the PDO influence on these interactions should lead to better predictability of interannual and multidecadal variations in summertime precipitation in North America.

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