Influence of the Arctic oscillation on central United States summer rainfall

Qi Hu[1] and Song Feng[1]

Received 26 January 2009; revised 9 September 2009; accepted 16 September 2009; published 5 January 2010.

[1] Effects of the Arctic oscillation (AO) on summer rainfall variability in the central United States are examined in order to improve understanding and prediction of the interannual variation in the summer rainfall. Major results show persistent AO effects that resulted in less summer rainfall in the central United States during the positive phase of the AO and more rainfall during the negative AO phase. These effects are most prominent at the interannual time scale. The key physical processes connecting the AO with the rainfall variations are shown in changes in latitudinal location of the midlatitude upper tropospheric westerly jet over North America, the transverse circulation around the jet, and the low-level moisture flux divergence in the central United States. Diagnostic analyses show that the change of the jet stream location may have resulted from the AO-induced eddy heat and momentum forcing on the mean zonal flow in the upper troposphere. The eddy-mean flow interactions caused a northward shift of the jet in the positive phase of the AO. The associated anomalies of downward motion and moisture divergence in the lower troposphere over the central United States suppressed rainfall development. A set of reversed anomalies developed in the negative phase of the AO, encouraging summer rainfall.

Citation: Hu, Q., and S. Feng (2010), Influence of the Arctic oscillation on central United States summer rainfall, *J. Geophys. Res.*, *115*, D01102, doi:10.1029/2009JD011805.

1. Introduction

[2] Interannual summer rainfall variations in the central United States are influenced by several major factors. The most recognized and studied one is the El Niño–Southern Oscillation (ENSO). During El Niño, or La Niña, extensive anomalies in distributions of the sea surface temperature (SST) and atmospheric convection in the tropical Pacific Ocean region disturb the midlatitude atmospheric circulation in North America [e.g., *Horel and Wallace*, 1981; *Held et al.*, 1989]. Alternations of the nearly opposite effect of El Niño and La Niña contribute to the interannual variations in circulation and precipitation in North America. Some detailed ENSO influences on the precipitation variations have been examined [e.g., *Ropelewski and Halpert*, 1986; *Ting and Wang*, 1997; *Hu and Feng*, 2001a, 2001b; *Weaver and Nigam*, 2008]. A particularly important circulation anomaly for increasing precipitation was found to be a strengthened northerly flow over the central United States. and an extended trough from the central to south central United States during El Niño [*Ting and Wang*, 1997; *Hu and Feng*, 2001a]. This trough interacted with the southerly low level jet (LLJ) from the Gulf of Mexico to the central United States, enhancing moisture flow and precipitation [e.g., *Higgins et al.*, 1997; *Hu and Feng*, 2001b].

[3] The LLJ, as another major factor, also strongly affects the summer rainfall variation in the central United States by its own variability [*Helfand and Schubert*, 1995; *Mo et al.*, 1995]. Intense LLJ and strengthened moisture flow, when aided with persistent local soil moisture anomalies, have resulted in extended and intense summer rainfall in the central and south central United States, causing severe floods such as the 1993 flood [*Helfand and Schubert*, 1995; *Mo et al.*, 1995; *Paegle et al.*, 1996; *Trenberth and Guillemot*, 1996; *Mo et al.*, 1997].

[4] It is intriguing that the influences of these factors on the central United States summer rainfall also have been varying in time. The ENSO effect was strong, explaining about 31% of the variance of the summer rainfall in the central United States during the time period of 1871–1916 and 1948–1978, but virtually diminished in 1917–1947 and from 1979 through recent years [*Hu and Feng*, 2001a]. Although the LLJ showed enhanced influence on the central U.S. summer rainfall in the period when ENSO effect languished, the influence was not persistent but episodic [*Hu and Feng*, 2001b]. These variations in the effects of ENSO and LLJ lead to the question of what could have been the factor or mechanism that may have sustained the interannual summer rainfall variation in the central United States when none of the LLJ and the ENSO effects was at work. Such a mechanism could explain the observed

[1]School of Natural Resources and Department of Geosciences, University of Nebraska at Lincoln, Lincoln, Nebraska, USA.

0148-0227/10/2009JD011805$09.00

persistent interannual variation in summer rainfall in the central United States.

[5] One such factor is the Arctic oscillation (AO) [*Thompson and Wallace*, 1998], also known as the "North's El Niño" [*Kerr*, 2001], which has a strong interannual component. The background and importance of the AO effect have been elaborated by *Thompson and Wallace* [1998, 2000], who show that in the positive phase of the AO the circulation in the midlatitude and high-latitude regions has an anomalously strong polar vortex and a northward shift of the midlatitude westerly jet stream. In the negative phase, the AO is characterized with a strong blocking over Greenland and/or Alaska and a trough in midlatitude North America. These anomalous atmospheric circulations affect the storm track, frequency and intensity of synoptic weather events, and precipitation in midlatitude and high-latitude North America [*Thompson and Wallace*, 2001; *Wettstein and Mearns*, 2002; *Higgins et al.*, 2002; *McAfee and Russell*, 2008; *Archambault et al.*, 2008; *Klingaman et al.*, 2008; *Knight et al.*, 2008]. In addition, *Higgins et al.* [2000] show that during the period from 1964 to 1993 atmospheric circulation anomalies associated with the AO and ENSO accounted for nearly all the observed fluctuations in boreal winter precipitation in the contiguous United States. These results prompted them to suggest that the U.S. cold season precipitation prediction can be considerably improved if, in addition to the tropical Pacific SST effect, the AO effects can be understood and accurately described in the prediction models.

[6] While the AO effect is strong in boreal winter *Thompson and Wallace* [2000] show that the AO also explains 16% of the total variance of the warm season atmospheric circulation in the midlatitude and high-latitude regions. This number is comparable to the 21% of the total variance of the atmospheric circulation that AO can explain during the northern winter. Thus, although the AO is weaker in northern summer than in winter the anomalous circulation associated with the AO may still have played significant roles in variations of the summer atmospheric circulation. Such roles of the AO in weather and climate in the boreal summer have been shown in several recent studies. For example, *Gong and Ho* [2003] show that the AO influences the summer monsoon rainfall in East Asia. *Ogi et al.* [2005] use a revised AO index and show that the AO played an important role in configuring the circulation anomalies responsible for the heat waves in Europe and Russia in 2003. Because the AO describes anomalies in circumpolar circulation, these effects of the AO on summer precipitation and temperature in East Asia and Europe suggest similar influences of the AO on the warm season circulation and precipitation in North America.

[7] While these results are showing the AO effects on warm season precipitation and temperature variations, the warm season AO itself has little predictability. In other words, the AO index of a future month has little predictability from the AO index of the previous month(s). Thus, the spring AO index may not be a predictor of the summer AO index and summer precipitation and temperature. Although this predictability is unavailable it remains important to examine the simultaneous AO effects on regional precipitation and temperature in the warm season. These effects and their underlying processes can help improve our understanding of development of warm season precipitation and temperature anomalies (such as the 2003 heat waves in Europe). This understanding can be used to improve predictions of summer rainfall when the AO becomes predictable, through other processes or variables.

[8] In this paper we present the analysis results showing the AO effects on the summer rainfall variation in the central United States. The data and methods used in the analyses are described in section 2. Major results of the analyses are presented and discussed in section 3. Possible underlying mechanisms that facilitate the AO influence on the warm season precipitation are proposed and evaluated in section 4. Section 5 contains a summary and discussion. Findings of this study help set a framework that will guide our further investigation and understanding of the processes and their interactions contributing/causing the interannual as well as longer time scale variations in warm season precipitation in the central United States.

2. Data and Methods

[9] Because of the skewed distribution of precipitation data and the requirement for normal (Gaussian) distribution of data in most of the statistical analyses we use the standardized precipitation index (SPI) to describe precipitation variations/anomalies in this study. As shown by *McKee et al.* [1993], the SPI time series is a transformed precipitation time series but has a normal distribution. The SPI is calculated at a given location using observed precipitation records for a desired period, e.g., a month or a season. This SPI time series is fitted to a probability distribution and then transformed into a time series with a normal distribution [*Edwards and McKee*, 1997]. Positive (negative) SPI indicates greater (less) than median precipitation. In this study, monthly SPI data series at 344 climate divisions in the contiguous United States from 1895 to 2007 were obtained from the U.S. National Climatic Data Center (NCDC) (http://www1.ncdc.noaa.gov/pub/data/cirs/drd964x.sp01.txt). These SPI data were developed at NCDC using stations' monthly precipitation data which were evaluated and quality controlled. Details of the quality control procedures were described by *Karl and Biebsame* [1984]. In addition to the SPI data, we also acquired the quality controlled divisional monthly precipitation data in the central United States. Using these precipitation data, we calculated the monthly SPI for each state using area weighted monthly precipitation from the climate divisions in a state, and also computed the monthly SPI for the central United States using area weighted monthly precipitation from the 43 climate divisions in Illinois, Iowa, Kansas, Missouri, and Nebraska.

[10] Two AO indices are used in this study, covering different periods from 1899 to 2007. The first index is the leading mode of the empirical orthogonal function (EOF) of the monthly mean sea level pressure (SLP) anomalies north of the 20°N parallel [*Thompson and Wallace*, 1998, 2000]. Data of this index from 1899 to 2001 are obtained from the online source at http://jisao.washington.edu/data/aots/. The second index is developed at the NOAA Climate Prediction Center (CPC). This index is constructed by projecting the daily 1000 mb height anomalies poleward of 20°N onto the

leading EOF of the monthly mean 1000 mb height during 1979–2000. Data of this index from 1950 to 2007 are obtained from NOAA CPC web site http://www.cpc.noaa.gov/products/precip/CWlink/daily_ao_index/ao_index.html.

[11] These two AO indices are highly correlated and their correlation coefficient is 0.98 for their common period of 1950–2001. This high correlation warrants us to extend the AO index based on the SLP from 2001 to 2007 using the index from the NOAA CPC, using a linear regression method. High AO index values characterize lower than average polar cap sea level pressure, stronger lower-stratospheric polar vortex, and stronger zonal winds along the 35°N–55°N latitudes [*Thompson and Wallace*, 1998, 2000].

[12] The National Center for Environmental Prediction-National Center for Atmospheric Research (NCEP-NCAR) reanalysis data [*Kalnay et al.*, 1996] are used to examine the atmospheric circulation and moisture flux associated with the AO-induced anomalies in North America. The moisture flux is integrated between the sigma levels 19–28 (surface to roughly 700 hPa) in the data. The monthly mean zonal and meridional wind (u, v) and vertical motion (ω) in the pressure coordinates are used to analyze variations in the midlatitude westerly jet stream and vertical motion in the study region.

[13] To assist our understanding of how the AO may affect the summer rainfall in the central United States, we examine the eddy heat and momentum forcing on the mean zonal flow [*Eliassen and Palm*, 1961; *Lindzen and Holton*, 1968], testing the hypothesis that the eddy flux anomalies developed in different phase of the AO interact with the mean zonal flow in the midlatitude and modify the jet stream which further influences the warm season weather and rainfall in the central United States. The eddy forcing is calculated using the localized Eliassen-Palm (E-P) flux theory [*Trenberth*, 1986, 1991], based on the geostrophic E-P diagnostics [*Eliassen and Palm*, 1961; *Edmon et al.*, 1980] with the advantage of including the ageostrophic effect on local mean flows. The NCEP-NCAR reanalysis daily data are used in calculations of the eddy heat and momentum fluxes. Before calculating these fluxes a 2–8 d band-pass filter was applied to the reanalysis data, similar to that used by *Trenberth* [1991]. The filtered data contain the mesoscale to synoptic-scale weather disturbances. Thus, eddy fluxes computed from these filtered data can assist us to examine how the synoptic disturbances, particularly those associated with the AO, may have interacted with the midlatitude jet stream in different AO phases and how the resulting circulation anomalies may have affected summer rainfall in the central United States.

[14] The localized E-P flux vector is defined as [*Trenberth*, 1986]

$$\mathbf{E}_u = \cos\phi\left\{\frac{1}{2}\left(\overline{v'^2} - \overline{u'^2}\right), -\overline{u'v'}, -\left(\frac{f}{\sigma}\overline{v'T'} + \overline{u'\omega'}\right)\right\} \qquad (1)$$

where, T is the air temperature and φ the latitude. The overbar in (1) indicates the time mean, the prime is departure from the mean, f is the Coriolis' parameter, and $\sigma = -(\partial T/\partial p) + \kappa T/p$ is the static stability. Because $\overline{u'\omega'}$ is much smaller than the other terms in (1), it has often been neglected [*Trenberth*, 1991]. The effect of eddies on the mean zonal flow is described in

$$\frac{\partial \bar{u}}{\partial t} = \frac{1}{\cos\phi}\nabla\cdot\boldsymbol{E}_{\mathbf{u}}, \qquad (2)$$

where

$$\nabla = \left[\frac{1}{a\cos\phi}\frac{\partial}{\partial\lambda}, \frac{1}{a\cos\phi}\frac{\partial}{\partial\phi}\cos\phi, \frac{\partial}{\partial p}\right].$$

Equation (2) states that the E-P flux divergence (convergence) causes acceleration (deceleration) of the westerly mean zonal flow. Changes in such divergence pattern could cause variations in the intensity and position of the westerly jet stream and influence synoptic processes and development of precipitation.

[15] To aid our understanding of the eddy-mean zonal flow interactions and their effects on synoptic processes and precipitation we also acquired the surface cyclone data from *Serreze et al.* [1997]. These data are composed of 6 hourly observations from the NCEP-NCAR reanalysis SLP and cover the period from 1948 to 2001. The selected cyclones had life span of at least 12 h. For these long-lived cyclones, their occurrences in a grid area of 5.0° × 5.0° latitude and longitude are counted and used to measure the cyclone frequency in the central United States. This cyclone frequency is used to describe the influence of the AO on the synoptic weather and precipitation anomalies in the central United States.

[16] The statistical methods used in this study are correlations and regression. Because the time series of the AO and SPI may be autocorrelated, the autocorrelation must be removed before the cross correlation can be calculated and its significance tested. We follow the approach of *Livezey* [1995] and *Janowiak et al.* [1998] to remove the autocorrelation. In this procedure, the "effective time" of autocorrelation between two sample series, $a(t)$ and $b(t)$ from the SPI time series, say, is computed from

$$TE = 1 + 2\sum_{i=1}^{N}\left(1 - \frac{i}{n}\right)(\rho_a)_i(\rho_b)_i. \qquad (3)$$

In (3), TE is the effective number of years (summer seasons) of autocorrelation of the two sample series, $a(t)$ and $b(t)$, ρ_a and ρ_b are the i-th year lagged autocorrelations for $a(t)$ and $b(t)$, n is the sample length of the original time series, and N is the total number of lags ($N = n/2$, i.e., one half of the sample length). For the AO and SPI time series, we use $n = 108$ and $N = 54$. According to the value of TE the SPI time series is "trimmed" by removing the correlated data and eventually contains $m(= n/\mathrm{TE})$ independent samples. The same procedure is applied to the AO index time series. The correlation coefficient of the trimmed SPI and AO time series is then computed. The significance of

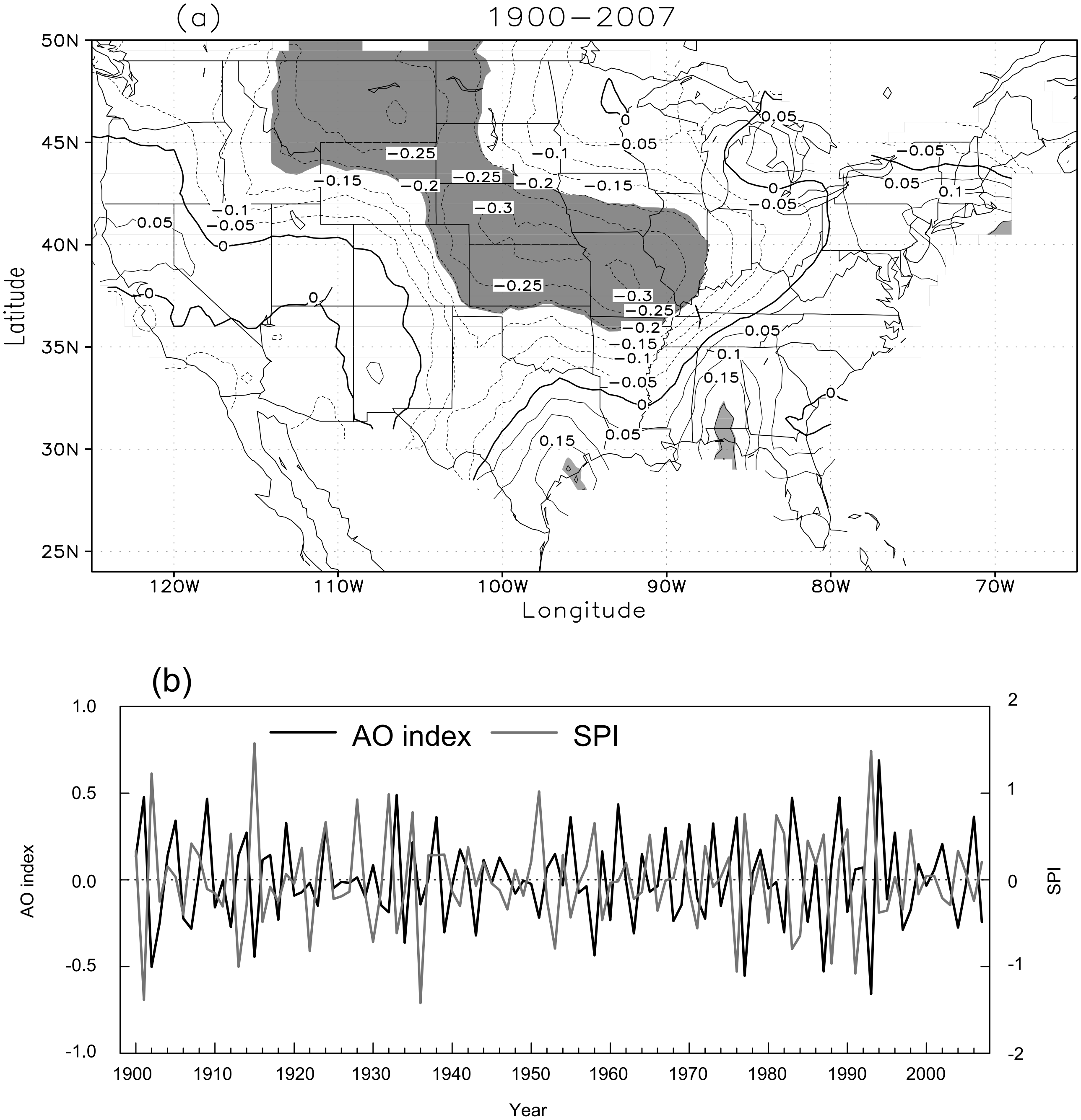

Figure 1. (a) Correlation between the summer Arctic oscillation (AO) and the summer standardized precipitation index (SPI) at the 344 climate divisions in the contiguous United States from 1900 to 2007. Shading indicates correlation significant at the 95% confidence level using a two-tailed t test (same significance test was applied to the rest of the results). (b) Interannual variations of the summer AO index and the SPI in the central United States during 1900–2007.

the correlation (selected at 95% confidence level) is evaluated using a two-tailed t test with $m - 2$ degrees of freedom (see *Livezey* [1995] and *Janowiak et al.* [1998] for further details of this method).

[17] To focus on the interannual variations, a 9 point (year) binomial high-pass filter is applied to the AO index, SPI, and the atmospheric circulation and moisture flux data. A 9 year filter will keep variations shorter than 10 years in the filtered data series.

3. Influence of the AO on Central U.S. Summer Rainfall

[18] Influence of the AO on summer rainfall variations in the central United States is indicated by the result in

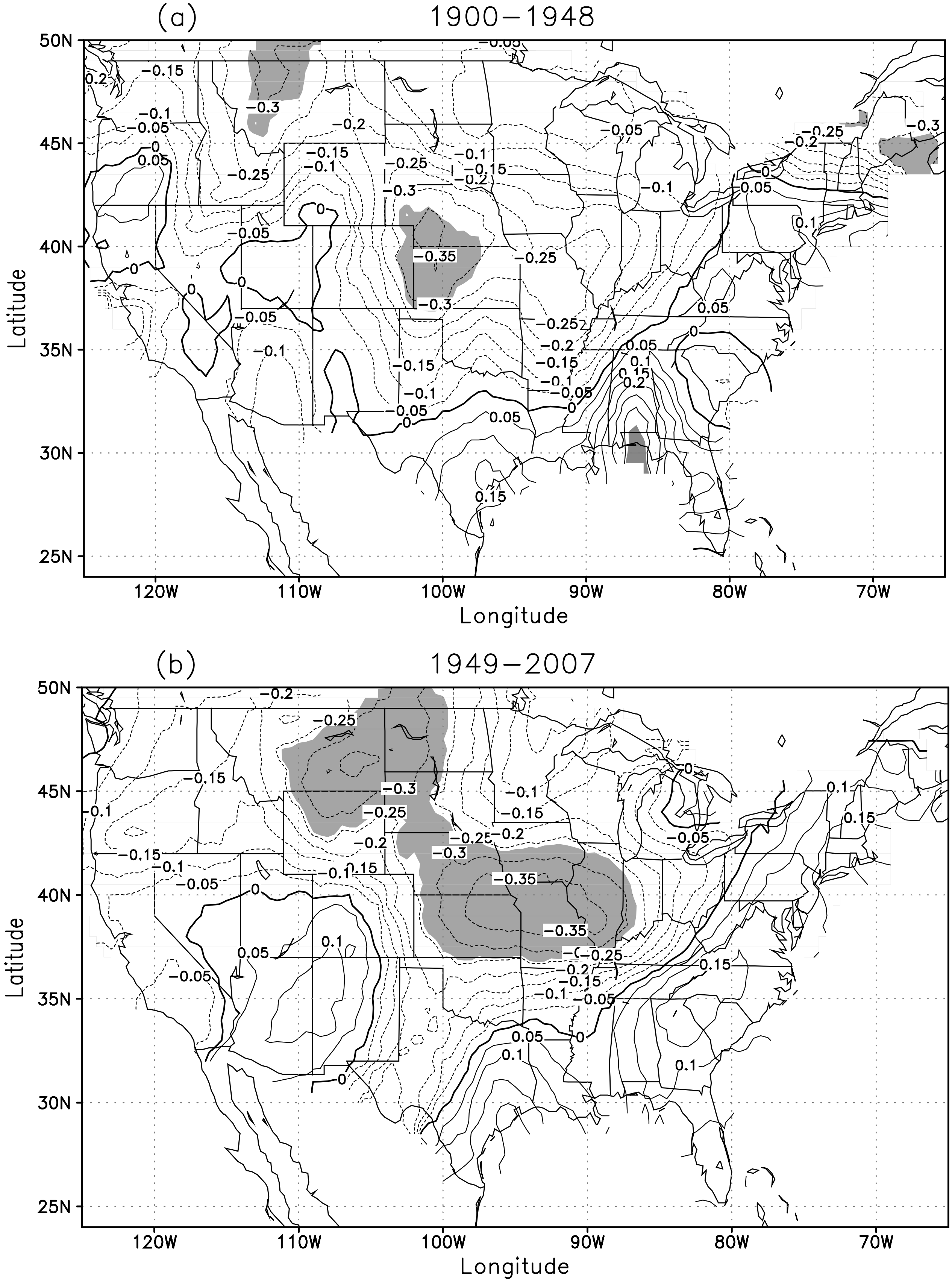

Figure 2. Spatial patterns of the correlation between the summer AO index and the summer SPI for (a) 1900−1948 and (b) 1949−2007. Shading indicates correlation significant at the 95% confidence level.

Table 1. Relationship Between the Summer Antarctic Oscillation Index and the State-Averaged Standardized Precipitation Index in the Five States of the Central United States for the Period From 1900 to 2007[a]

States	Unfiltered Data Result	Filtered Data Result	SPI Anomaly in Positive AO Phase	SPI Anomaly in Negative AO Phase	Difference in SPI Anomaly
Illinois	**−0.305**	**−0.420**	−0.153	0.327	**−0.439**
Iowa	**−0.233**	**−0.417**	−0.166	0.297	**−0.463**
Kansas	**−0.294**	**−0.321**	−0.263	0.336	**−0.599**
Missouri	**−0.363**	**−0.373**	−0.291	0.344	**−0.636**
Nebraska	**−0.375**	**−0.316**	−0.349	0.452	**−0.801**
Five-states mean	**−0.376**	**−0.457**	−0.311	0.422	**−0.733**

[a]Columns 2 and 3 show the correlation between the standardized precipitation index (SPI) and the Arctic oscillation (AO) from unfiltered and filtered data series, respectively. Columns 4 and 5 show state-averaged SPI anomaly during the positive and negative AO phase, respectively. Their differences are given in column 6. Bold type indicates that the correlation or the difference is significant at the 95% confidence level using a two-tailed *t* test.

Figure 1a, which shows the correlations between the June–July–August (JJA) SPI at the 344 climate divisions in the contiguous United States and the summer AO index from 1900 to 2007. Statistically significant negative correlations between the two are found in the area of Montana, Wyoming, western portions of North and South Dakota, Nebraska, Kansas, Missouri, and Illinois. These results indicate less summer rainfall in the central United States and the northern Great Plains during the positive phase of the AO and more rainfall during the negative phase of the AO.

[19] Persistence of this negative correlation between the AO and the SPI in the central United States is indicated by the result in Figure 1b. The two time series plotted in Figure 1b show an out-of-phase relation of the two indices over the study period, although there are a few exceptions. These fairly consistent out-of-phase variations indicate that alternations of the AO have been associated with consistent fluctuations in the SPI (though the amplitude varied). These features in the variations support a strong linear relationship between the AO and the central U.S. summer SPI although the interacting processes resulting in these coherent variations may be highly nonlinear.

[20] Correlations between the summer SPI and the AO in various time periods during 1900–2007 also were calculated. In addition, the spatial pattern of the correlation of the summer SPI with the AO index was examined in those different time periods. Two of such results are shown in Figures 2a and 2b. The patterns of the correlation in these two periods are similar, and they also are very similar to that in Figure 1a. Although the areas of significant negative correlation in Figure 2a are smaller than those in Figures 2b and 1a, possibly because of the fewer number of stations and observations in the early part of the 20th century [*Karl and Biebsame*, 1984], this similarity indicates a strong and significant negative correlation of the summer AO and the SPI in the central United States. It also indicates that the influence of the AO on the summer rainfall has been fairly consistent over the past 108 years. An additional fact supporting the AO influence is that the severe floods in the central United States in 1902, 1915, 1951, and 1993 all occurred when the AO was in strong negative phase, and that the severe droughts in the region in 1953, 1976, 1983, and 1988 happened during the strong positive phase of the AO.

[21] For the five individual states in the central United States their summer SPI correlation with the AO index during 1900–2007 ranges from −0.23 to −0.37, all significant at the 95% confidence level (Table 1). Furthermore, the correlation is stronger at the interannual time scale, except for Nebraska (see the second column in Table 1). This strong covariance of the SPI and the AO at the interannual time scale could arise because the AO has the largest power at such time scale [*Thompson and Wallace*, 1998]. Moreover, because the AO-induced eddy forcing also is strong at the interannual time scale, as will be shown in section 4, the strong covariance of the AO and the SPI suggests a strong effect of the AO on interannual variations in the summer SPI and rainfall of the central United States.

[22] In columns 4 and 5 of Table 1, we also show the state averaged SPI anomaly in different phases of the AO. The difference of the averaged SPI anomaly in opposite phases of the AO is summarized and shown in column 6, demonstrating significant change in dryness and wetness in these states in accordance with the AO. In these comparisons, the positive (negative) phase of the AO is defined when the AO index is greater than or equal to 1.0 (less than or equal to −1.0) standard deviation of the AO variation. It is also interesting to note that the anomalies of the summer SPI in the negative phase of the AO are usually larger than the SPI anomalies in the positive phase. This asymmetry may suggest a stronger effect of the AO on the summer rainfall variation during the negative phase of the AO.

4. An Explanation of the AO Effect

[23] The results presented in section 3 show a persistent effect of the AO on interannual summer rainfall variations in the central United States. They also pose the question of what processes may have connected the AO and the circulation and rainfall anomalies in the study region. To address this question, we examined the 200 hPa zonal wind variation and its relation with the AO and the SPI anomalies. A major reason for this analysis is that the westerly jet stream, serving as a guide of the storm track, is strongest in the upper troposphere and its latitudinal location and structure influence the storm development and precipitation in North America [e.g., *Trenberth*, 1986; *Mo et al.*, 1995; *Trenberth and Guillemot*, 1996]. The interrelationships of the AO, 200 hPa zonal wind (U200), and the central U.S. SPI variations are shown in Figure 3. The climatology of

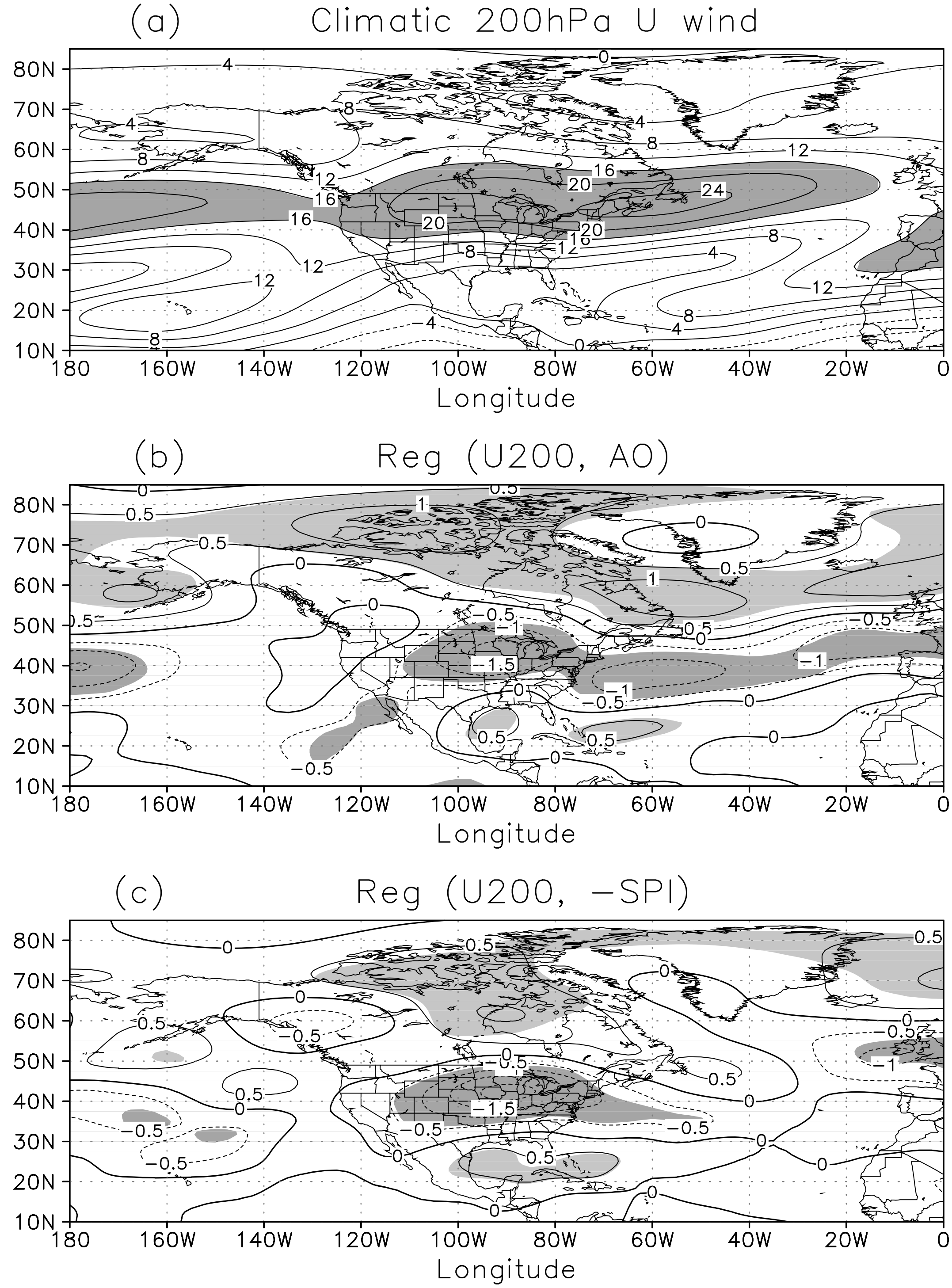

Figure 3. (a) Climatology of 200 hPa zonal wind for the period 1948–2007. (b) Changes in summer 200 hPa zonal wind corresponding to a unit deviation of the summer AO index at interannual time scale during 1948–2007. The AO index was normalized before regression. (c) Same as Figure 3b but −SPI in the central United States (SPI is multiplied by −1 so it has the same sign as precipitation). The contour interval is 5 m/s in Figure 3a and 0.5 m/s in Figures 3b and 3c. Shading in Figure 3a is the jet core and shading in Figures 3b and 3c indicates the regression significant at the 95% confidence level.

U200 is shown in Figure 3a for comparison purpose. The change in the U200 corresponding to a unit deviation of the AO index at interannual time scale is shown in Figure 3b, and the influence of changes in U200 on the warm season SPI in the central United States during 1948–2007 is illustrated in Figure 3c. The most noticeable feature in Figures 3b and 3c is that both the variations in the summer AO index and SPI have strong association with the U200.

During the positive phase of the AO, U200 strengthened in the north side of the climatological position of the midlatitude jet stream and weakened on the south side of the jet. Meanwhile, U200 weakened considerably over the central and north central United States. The weakened U200 over the central United States is closely associated with the decrease of summer precipitation in the region (Figure 3c). On the other hand, during the negative phase of the AO, the midlatitude westerly jet is stronger and displaced south of its climatological summer position, and the central United States is wetter. These associations indicate a negative relationship between the central U.S. summer rainfall and the AO. This connection is through their interactions with the upper tropospheric zonal wind and the jet stream. It is important to further point out that the northward shift of the 200 hPa jet stream and subsequent weakening of the U200 over the central United States were found as the major factors contributing to the 1988 U.S. summer droughts. A reversed anomaly in the same region also was found in strong association with the 1993 floods in the central and south central United States [*Bell and Janowiak*, 1995; *Trenberth and Guillemot*, 1996]. These findings suggest that by influencing the upper tropospheric westerly jet stream and storm track the AO could have played important roles in development of extreme rainfall events and the interannual variation of the central U.S. summer rainfall.

[24] It is also reasonable to question if, instead of the northward shift of the westerly jet, a double-jet structure would emerge in response to the positive AO. In such a case, a new westerly jet emerges to the north of the existing midlatitude jet. An example of such double-jet structure was reported by *Ogi et al.* [2005], who suggested that the double jet strongly influenced the development of the synoptic pattern for the 2003 summer heat waves in Europe. A reexamination of the 2003 summer circulations in the Northern Hemisphere showed that the double-jet structure only appeared during the short period from 17 July to 6 August 2003, which was the period when heat waves developed in Europe [*Ogi et al.*, 2005]. A high-latitude jet also emerged in North America around 70°N–75°N and this jet was much weaker than the jet over Europe. This high-latitude jet disappeared, however, on monthly to seasonal time scales. This brief appearance of this high-latitude jet may suggest it as a synoptic-scale disturbance instead of seasonally persistent circulation feature associated with the AO and affecting seasonal rainfall in North America.

[25] The observed changes in the midlatitude jet stream during the AO (Figure 3) can be achieved by influence on the mean zonal flow from eddy heat and momentum forcing associated with the AO, in a way similar to the eddy effects on the mean zonal flow elaborated by *Edmon et al.* [1980] and *Trenberth* [1986]. To examine the effect of the eddy forcing on variations in the westerly jet stream during the AO we applied the E-P flux diagnostics described in section 2. According to the diagnostics, a divergence of the eddy flux would strengthen the westerly mean zonal flow and the jet, and a convergence would weaken the mean flow and the jet. The divergence and convergence of eddy fluxes for each summer during 1948–2007 were calculated. Before showing them in Figure 4, these divergence and convergence have been smoothed using a spherical harmonic filter [*Sardeshmukh and Hoskins*, 1984; *Mo et al.*, 1995] with a triangular truncation at wave number 16 to capture the synoptic-scale variations. Thus, the results will show the effects on the mean zonal flow from synoptic-scale eddy disturbances in different phases of the AO.

[26] Figure 4a shows the changes in horizontal component of the E-P flux divergence, which describes the eddy barotropic forcing on the mean zonal flow. Figure 4b shows the changes in the vertical component of the E-P flux divergence, which describes the eddy baroclinic forcing on the mean zonal flow. These changes all correspond to a unit change in the AO index during the positive AO phase. The result in Figure 4a shows that during the positive AO phase there is large convergence in horizontal E-P flux in most of North America except for the region from the central United States to northern Canada. Meanwhile, the vertical E-P flux (Figure 4b) has significant convergence in the central United States. This region of convergence extends to the eastern United States and has another center in the midlatitude North Atlantic. While these results indicate that the barotropic and baroclinic eddy forcing tend to balance each other in individual years, their net forcing after averaging over the period from 1948 to 2007 shows that the baroclinic eddy forcing has been dominant during the positive phase of the AO. This net effect could be an indication that there is stronger polarward transport of heat during this AO phase [*Trenberth*, 1991; *Mo et al.*, 1995]. Although the net E-P flux convergence shown in Figure 4c is not significant over the central United States, it is adequate in showing that the eddy forcing works to weaken the westerly jet stream in the positive phase of the AO. This area of E-P flux convergence also coincides with the area of the maximum negative zonal wind anomaly shown in Figure 3b.

[27] To the north of this area of eddy flux convergence and weakening of the mean zonal flow is a broad region of strong net divergence of E-P flux. Zonal wind in this region is strengthening (Figure 4c). The strengthened zonal flow in this area is consistent with the findings from prior studies showing the strengthened polar vortex during the positive phase of the AO [*Thompson and Wallace*, 1998, 2000]. The strengthened zonal flow in higher latitude and weakened zonal flow in midlatitude North America result in a shift of the westerly jet to a northern position from its climatological mean summer location.

[28] To understand how the northward shift of westerly jet stream and the underlying eddy processes may have contributed to decrease in summer rainfall in the central United States, we examined the circulation anomalies associated with the shift of the jet. The major results are summarized in Figure 5. As shown in Figure 5, when the westerly jet stream shifts to a northern location positive pressure anomaly develops in the central United States between 35°N–45°N. In this region subsidence motion develops in the troposphere, particularly strong in the mid and lower troposphere. The subsidence motion plays several roles in suppressing summer rainfall in the central United States. It discourages development of strong convection which is a major source of summer rainfall in the region. The low-level divergence driven by the subsidence also weakens the low-level southerly flow from the Gulf of Mexico to the central United States. In Figure 5, the suppressed southerly low level flow during the positive phase

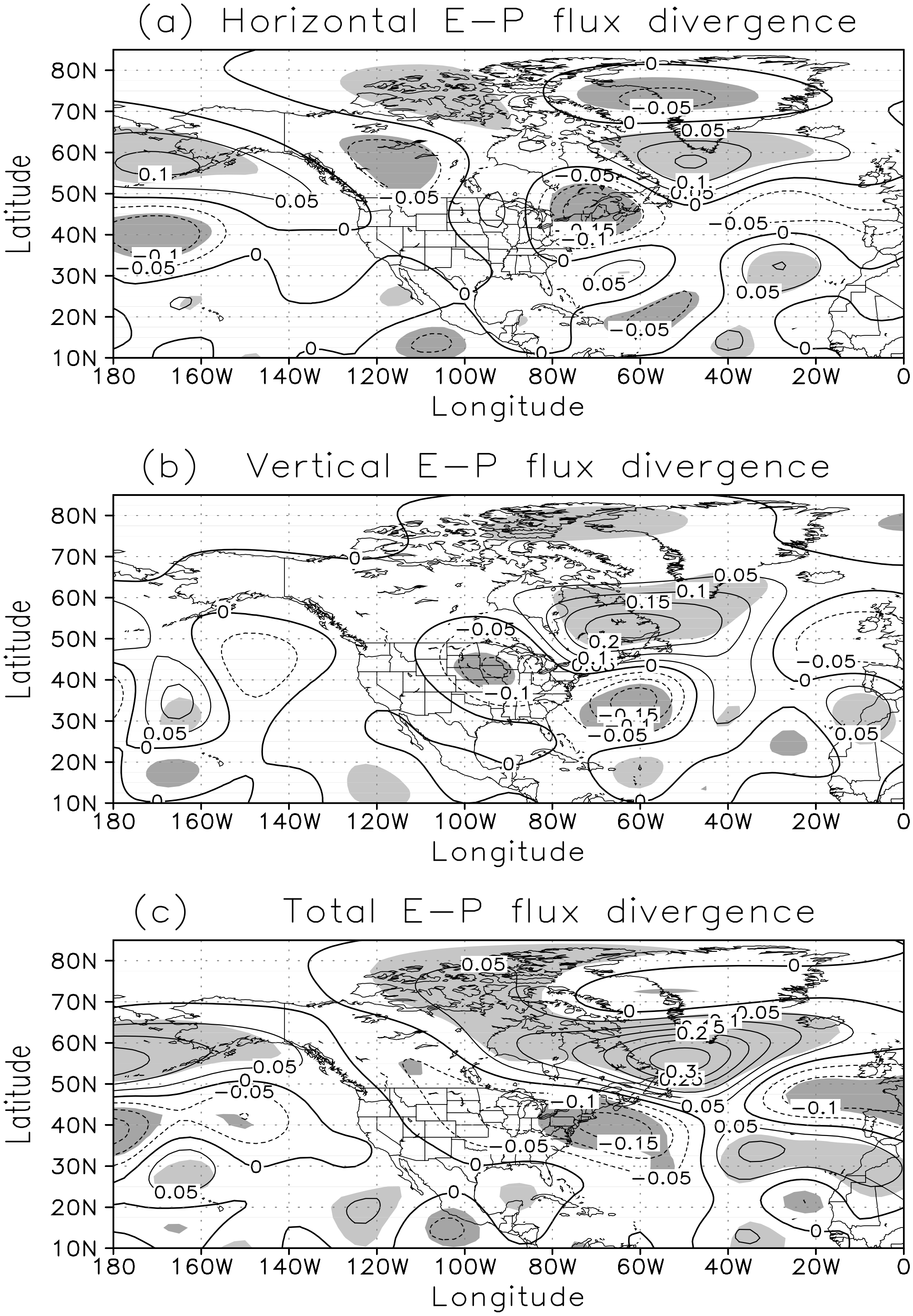

Figure 4. Changes in summer (a) horizontal, (b) vertical, and (c) total E-P flux divergence at 200 hPa corresponding to a unit deviation of the AO index on interannual time scale during 1948–2007. The contour interval is 0.05 m/s per day. Regions of changes significant above the 95% confidence level are shaded.

of the AO is shown by the anomalous northerly flow between 850 and 1000 hPa from 30°N–40°N. As a consequence, divergence of low-level moisture flux enhances as indicated in Figure 6, causing drying in the central United States.

[29] In addition, along with the northward shift of the westerly jet during the positive AO is the change in the storm track position. The departing storm track and the anomalous subsidence motion in the central United States (Figure 5) suppress the potential for development of synoptic weather events, which are the other major source of the region's summer precipitation in addition to convective storms. The decrease in synoptic weather events in the central United States during the positive AO phase is

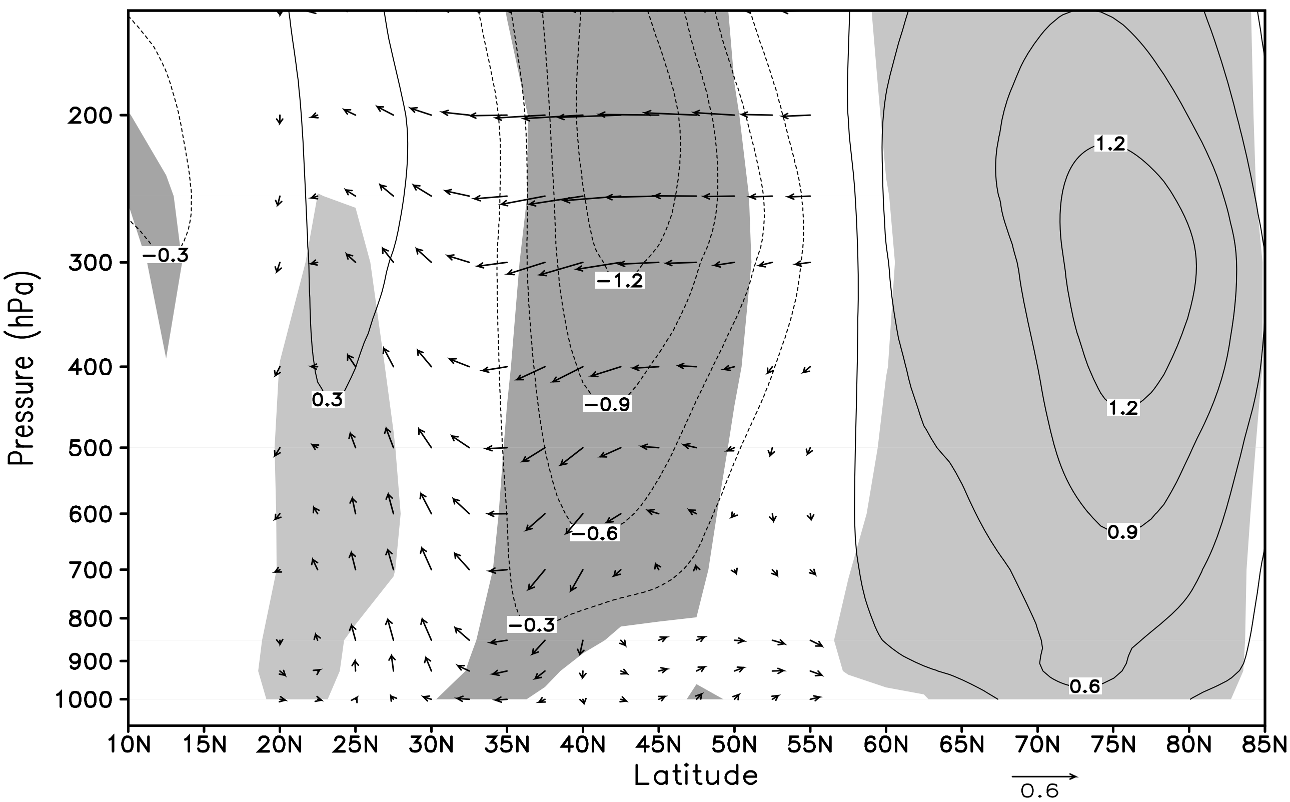

Figure 5. Cross-section of changes in zonal mean zonal wind (u), meridional wind (v), and vertical motion (ω) averaged between 85°W–105°W corresponding to a unit deviation of the summer AO index on interannual time scale during 1948–2007. The u is shown as contours with an interval of 0.3 m/s and its regression with the AO index at above the 95% confidence level is shaded. The covariance of v and ω are shown by vectors. The ω was multiplied by 100 before plotting. The covariance of v and ω north of the 55°N and south of the 20°N are not shown for display clarity.

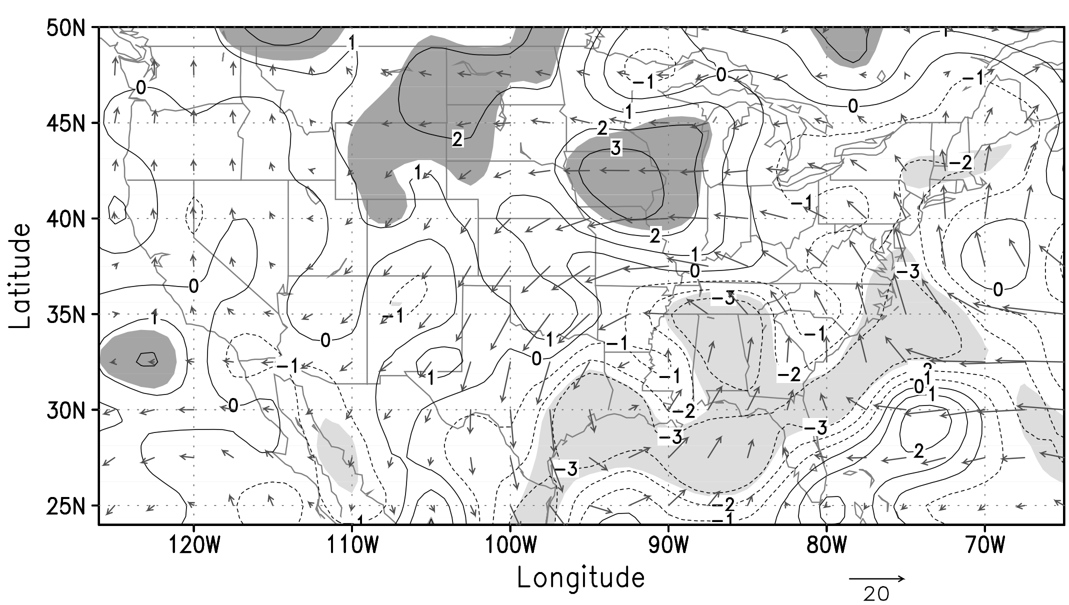

Figure 6. Changes in summer moisture divergence (contour) and moisture flux (vector) corresponding to a unit deviation of the summer AO index on interannual time scale during 1948–2007. The units of moisture flux vector are kg/(ms) and the contour interval for moisture divergence is 1.0×10^{-6} kg/(m^{2}s). Regions of above the 95% confidence level are shaded.

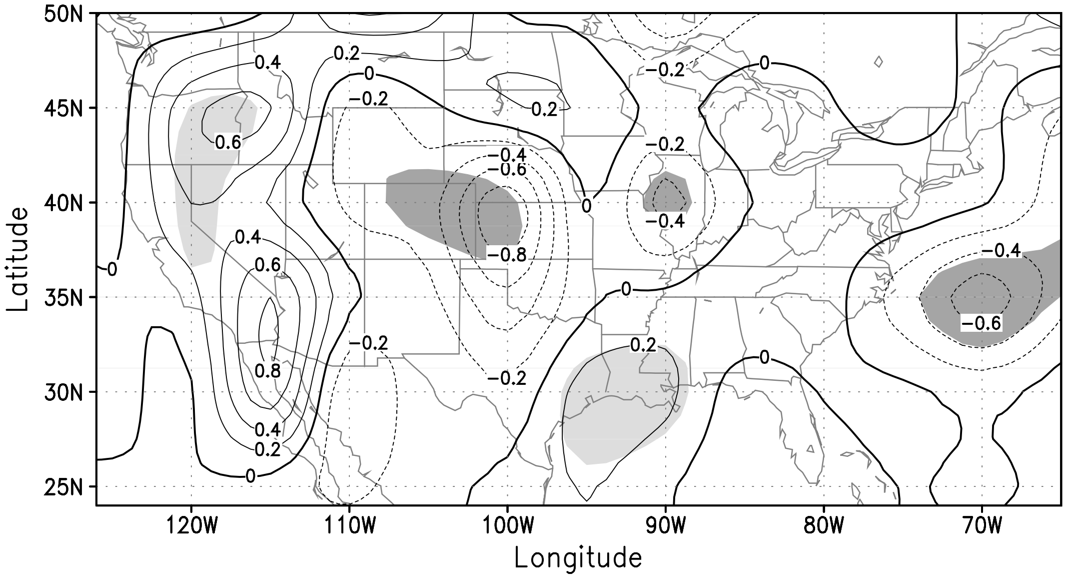

Figure 7. Changes in summer surface cyclone frequency corresponding to a unit deviation of the summer AO index on interannual time scale. Contour interval is 0.2 cyclones. Regions of the above 95% confidence level are shaded.

shown by the result in Figure 7. The reduced number of cyclonic events further reduces summer rainfall in the region.

[30] These anomalies in eddy processes and resulting anomalous circulation reverse during the negative phase of the AO. The reversed anomalies configure a circulation environment that favors development of synoptic events, strong moisture convergence, more convective storms and summer rainfall in the central United States. These variations associated with the AO in its different phases contribute to the interannual summer rainfall variation in the central United States.

[31] Finally, it is necessary to point out that these correlation/regression analyses showed little asymmetry as shown in the results in Table 1, primarily because correlations only show an averaged strength of association between two variations over a specific period (time series). It is likely that a similar asymmetry exists between the effects of the AO during its positive and negative phases, because the pattern and strength of the circulation and jet anomalies in the midlatitude and high-latitude North America were unlikely mirror images between the opposite phases of the AO [e.g., *Thompson and Wallace*, 1998, 2000]. The subtle asymmetric effects of the AO on the circulation and precipitation anomalies between the different AO phases are not examined in this study because such effects may only add information when making quantitative predictions of summer precipitation.

5. Discussions and Concluding Remarks

[32] The results presented in the previous sections show that, in addition to the ENSO and the LLJ, the AO and its associated circulation anomalies are another significant factor influencing the interannual variation in the summer rainfall of the central United States. Different from the fluctuating influence of the ENSO at the multidecadal timescale and the influence of the LLJ, the AO has a persistent effect on the rainfall in the past 108 years. The key process in the different phases of the AO that contributes to the precipitation variation is the eddy forcing on the mean zonal flow in midlatitude and high-latitude North America. Both barotropic and baroclinic eddies developed during the AO. In the positive phase of the AO, the anomalies of these eddy fluxes caused weakening of the upper tropospheric zonal mean flow over the central United States and strengthening of the zonal flow in the higher latitudes north of the region, resulting in a northward shift of the westerly jet stream. As a result of this shift of the jet there were subsidence motion and enhanced divergence in the lower troposphere over the central United States. These changes in the jet stream and associated secondary circulation reduced the number of summer cyclones and also suppressed the moisture supply and the potential for convective storm development. Because precipitation from these cyclones and convection activities constitutes nearly the entire summer rainfall, fewer cyclones and weaker convection resulted in deficit of summer precipitation and dry conditions in the central United States. Reversed anomalies in the negative phase of the AO contributed to an increase in the summer precipitation and wet conditions.

[33] Intriguingly, some of these AO-induced circulation and precipitation anomalies in the central United States share similarities with those from the ENSO and LLJ. For example, both synoptic analyses and modeling studies have shown that the midlatitude upper tropospheric jet shifted northward during major La Niña events and this anomaly contributed to drier conditions and sometimes severe drought in the central United States [e.g., *Trenberth and Guillemot*, 1996]. On the other hand, in strong El Niño years, the westerly jet displaced to a southern position from its climatological mean, and the associated circulation anomalies favored wet conditions in the central United States [e.g., *Hu and Feng*, 2001a; *Higgins et al.*, 2002].

These common features in jet stream anomalies arising from the different forcings suggest that the midlatitude upper tropospheric jet and associated transverse circulation anomalies are the primary cause of the interannual summer rainfall variation in the central United States. By altering the jet and the transverse circulation the different forcing processes have affected the region's interannual rainfall variation.

[34] Because these key features in jet stream anomaly can arise from these different forcings (ENSO and AO), they complicate the prediction of the interannual variation of summer rainfall in the central United States. This complexity also roots from the fact that the AO and the ENSO do not vary coherently over time [e.g., *Higgins et al.*, 2000]. Thus, their forcing on the jet and regional circulation and rainfall can be either in- or out-of-phase in any particular year. In an El Niño year when the AO also is in its negative phase forcing from El Niño and AO can mutually enhance their effects on the midlatitude upper tropospheric jet and the transverse circulation and amplify the net effect on the summer rainfall in the central United States. On the other extreme, in a La Niña year when the AO is in its positive phase the anomalies in the jet and associated circulation arising from these forcings could interact to severely suppress rainfall development in the central United States. Because ENSO has a time scale of 2–9 years and the AO is fairly regular at an interannual time scale, El Niño or La Niña can develop at different phases of the AO, and their interactions can result in rather different and complex influences on rainfall. These different influences could be an explanation, among others, of why some ENSO events had strong effects on the rainfall while the others produced trivial responses in rainfall variation. Some examples of such differences include the 1993 floods and the 1983 drought in the central United States. A strong El Niño peaked in early 1993 and maintained through the summer of that year when the AO was in its negative phase. Both the AO and the El Niño should have enhanced their effects on the jet and circulation anomalies, contributing to the severe flooding in the central United States. The 1983 was different, however. The El Niño in 1982–1983 tended to cause wet conditions in the central U.S, but the strong positive AO in the summer of 1983 favored a dry condition. As a net result from these effects, a dry summer emerged and prevailed in the central United States.

[35] Identifying these effects of the AO on the central U.S. summer rainfall variation helps improve our understanding of not only the rainfall variation but also the AO interactions with the other forcing processes and their collective influence on the rainfall. Knowing these sources of influences can improve our capacity to better understand and predict the interannual variation in summer rainfall of the central United States.

[36] **Acknowledgments.** We thank three anonymous reviewers, Editor Steve Ghan, and Robert Diffendal for their inputs that have led to improving this manuscript. This work was supported by NOAA grant NA09OAR4310188 to the University of Nebraska at Lincoln and by the USDA Cooperative Research Project NEB-40-040.

References

Archambault, H. M., L. F. Bosart, D. Keyser, and A. R. Aiyyer (2008), Influence of large-scale flow regimes on cool-season precipitation in the northeastern United States, *Mon. Weather Rev.*, *136*, 2945–2963, doi:10.1175/2007MWR2308.1.

Bell, G. D., and J. E. Janowiak (1995), Atmospheric circulation associated with the Midwest floods of 1993, *Bull. Am. Meteorol. Soc.*, *76*, 681–695, doi:10.1175/1520-0477(1995)076<0681:ACAWTM>2.0.CO;2.

Edmon, H. J., Jr., B. J. Hoskins, and M. E. McIntyre (1980), Eliassen-Palm cross sections for the troposphere, *J. Atmos. Sci.*, *37*, 2600–2616, doi:10.1175/1520-0469(1980)037<2600:EPCSFT>2.0.CO;2.

Edwards, D. C., and T. B. McKee (1997), Characteristics of 20th century drought in the United States at multiple time scales, *Climatol. Rep. 97-2*, 155 pp., Dep. of Atmos. Sci., Colo. State Univ., Fort Collins.

Eliassen, A., and E. Palm (1961), On the transfer of energy in stationary mountain waves, *Geofys. Publ.*, *22*(3), 23 pp., I Kommisjon hos Aschehoug, Oslo.

Gong, D.-Y., and C.-H. Ho (2003), Arctic oscillation signals in the East Asian summer monsoon, *J. Geophys. Res.*, *108*(D2), 4066, doi:10.1029/2002JD002193.

Held, I. M., S. W. Lyons, and S. Nigam (1989), Transients and the extratropical response to El Niño, *J. Atmos. Sci.*, *46*, 163–174, doi:10.1175/1520-0469(1989)046<0163:TATERT>2.0.CO;2.

Helfand, H. M., and S. D. Schubert (1995), Climatology of the simulated Great Plains low-level jet and its contribution to the continental moisture budget of the United States, *J. Clim.*, *8*, 784–806, doi:10.1175/1520-0442(1995)008<0784:COTSGP>2.0.CO;2.

Higgins, R. W., Y. Yao, E. S. Yarosh, J. E. Janowiak, and K. C. Mo (1997), Influence of the Great Plains low-level jet on summertime precipitation and moisture transport over the central United States, *J. Clim.*, *10*, 481–507, doi:10.1175/1520-0442(1997)010<0481:IOTGPL>2.0.CO;2.

Higgins, R. W., A. Leetmaa, Y. Xue, and A. Barnston (2000), Dominant factors influencing the seasonal predictability of U.S. precipitation and surface air temperature, *J. Clim.*, *13*, 3994–4017, doi:10.1175/1520-0442(2000)013<3994:DFITSP>2.0.CO;2.

Higgins, R. W., A. Leetmaa, and V. E. Kousky (2002), Relationships between climate variability and winter temperature extremes in the United States, *J. Clim.*, *15*, 1555–1572, doi:10.1175/1520-0442(2002)015<1555:RBCVAW>2.0.CO;2.

Horel, J. D., and J. M. Wallace (1981), Planetary-scale atmospheric phenomena associated with the Southern Oscillation, *Mon. Weather Rev.*, *109*, 813–829, doi:10.1175/1520-0493(1981)109<0813:PSAPAW>2.0. CO;2.

Hu, Q., and S. Feng (2001a), Variations of teleconnection of ENSO and interannual variation in summer rainfall in the central United States, *J. Clim.*, *14*, 2469–2480, doi:10.1175/1520-0442(2001)014<2469:VOTOEA>2.0.CO;2.

Hu, Q., and S. Feng (2001b), Climatic role of the southerly flow from the Gulf of Mexico in interannual variations in summer rainfall in the central United States, *J. Clim.*, *14*, 3156–3170, doi:10.1175/1520-0442(2001)014<3156:CROTSF>2.0.CO;2.

Janowiak, J. E., A. Gruber, C. R. Kondragunta, R. E. Livezey, and G. J. Huffman (1998), A comparison of the NCEP–NCAR reanalysis precipitation and the GPCP raingauge–satellite combined dataset with observational error considerations, *J. Clim.*, *11*, 2960–2979, doi:10.1175/1520-0442(1998)011<2960:ACOTNN>2.0.CO;2.

Kalnay, E., et al. (1996), The NCEP/NCAR 40-year reanalysis project, *Bull. Am. Meteorol. Soc.*, *77*, 437–471, doi:10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2.

Karl, T. R., and W. E. Biebsame (1984), The identification of 10- to 20-year temperature and precipitation fluctuations in the contiguous United States, *J. Appl. Meteorol.*, *23*, 950–966, doi:10.1175/1520-0450(1984)023<0950:TIOTYT>2.0.CO;2.

Kerr, R. K. (2001), Getting a handle on the north's 'El Niño', *Science*, *294*, 494–495, doi:10.1126/science.294.5542.494b.

Klingaman, N. P., B. Hanson, and D. J. Leathers (2008), A teleconnection between forced great plains snow cover and European winter climate, *J. Clim.*, *21*, 2503–2518.

Knight, D. B., et al. (2008), Increasing frequencies of warm and humid air masses over the conterminous United States from 1948 to 2005, *Geophys. Res. Lett.*, *35*, L10702, doi:10.1029/2008GL033697.

Lindzen, R. S., and J. R. Holton (1968), A theory of the quasi-biennial oscillation, *J. Atmos. Sci.*, *25*, 1095–1107, doi:10.1175/1520-0469(1968)025<1095:ATOTQB>2.0.CO;2.

Livezey, R. E. (1995), Field intercomparison, in *An Analysis of Climate Variability Application of Statistical Techniques*, pp. 159–176, edited by H. Von Storch and A. Navara, Springer, New York.

McAfee, S. A., and J. L. Russell (2008), Northern annular mode impact on spring climate in the western United States, *Geophys. Res. Lett.*, *35*, L17701, doi:10.1029/2008GL034828.

McKee, T. B., N. J. Doesken, and J. Kleist (1993), The relationship of drought frequency and duration to time scales, in *Eighth Conference on Applied Climatology*, pp. 179–184, Am. Meteorol. Soc., Anaheim, Calif.

Mo, K. C., J. N. Paegle, and J. Paegle (1995), Physical mechanisms of the 1993 summer floods, *J. Atmos. Sci.*, *52*, 879–895, doi:10.1175/1520-0469(1995)052<0879:PMOTSF>2.0.CO;2.

Mo, K. C., J. N. Paegle, and R. W. Higgins (1997), Atmospheric processes associated with summer floods and droughts in the central United States, *J. Clim.*, *10*, 3028–3046, doi:10.1175/1520-0442(1997)010<3028:APAWSF>2.0.CO;2.

Ogi, M., K. Yamazaki, and Y. Tachibana (2005), The summer northern annular mode and abnormal summer weather in 2003, *Geophys. Res. Lett.*, *32*, L04706, doi:10.1029/2004GL021528.

Paegle, J., K. C. Mo, and J. N. Paegle (1996), Dependence of simulated precipitation on surface evaporation during the 1993 United States summer floods, *Mon. Weather Rev.*, *124*, 345–361, doi:10.1175/1520-0493(1996)124<0345:DOSPOS>2.0.CO;2.

Ropelewski, C. F., and M. S. Halpert (1986), North American precipitation and temperature patterns associated with the El Niño/Southern Oscillation (ENSO), *Mon. Weather Rev.*, *114*, 2352–2362, doi:10.1175/1520-0493(1986)114<2352:NAPATP>2.0.CO;2.

Sardeshmukh, P. D., and B. J. Hoskins (1984), Spatial smoothing on a sphere, *Mon. Weather Rev.*, *112*, 2524–2529, doi:10.1175/1520-0493(1984)112<2524:SSOTS>2.0.CO;2.

Serreze, M. C., F. Carse, R. G. Barry, and J. C. Rogers (1997), Icelandic low cyclone activity: Climatological features, linkages with the NAO, and relationships with recent changes in the Northern Hemisphere circulation, *J. Clim.*, *10*, 453–464, doi:10.1175/1520-0442(1997)010<0453:ILCACF>2.0.CO;2.

Thompson, D. W. J., and J. M. Wallace (1998), The Arctic oscillation signature in the wintertime geopotential height and temperature fields, *Geophys. Res. Lett.*, *25*, 1297–1300, doi:10.1029/98GL00950.

Thompson, D. W. J., and J. M. Wallace (2000), Annular modes in the extratropical circulation. Part I: Month-to-month variability, *J. Clim.*, *13*, 1000–1016, doi:10.1175/1520-0442(2000)013<1000:AMITEC>2.0.CO;2.

Thompson, D. W. J., and J. M. Wallace (2001), Regional climate impacts of the Northern Hemisphere annular mode, *Science*, *293*, 85–89, doi:10.1126/science.1058958.

Ting, M. F., and H. Wang (1997), Summertime United States precipitation variability and its relation to Pacific sea surface temperatures, *J. Clim.*, *10*, 1853–1873, doi:10.1175/1520-0442(1997)010<1853:SUSPVA>2.0.CO;2.

Trenberth, K. E. (1986), An assessment of the impact of transient eddies on the zonal flow during a blocking episode using localized Eliassen-Palm flux diagnostics, *J. Atmos. Sci.*, *43*, 2070–2087, doi:10.1175/1520-0469(1986)043<2070:AAOTIO>2.0.CO;2.

Trenberth, K. E. (1991), Storm tracks in the Southern Hemisphere, *J. Atmos. Sci.*, *48*, 2159–2178, doi:10.1175/1520-0469(1991)048<2159:STITSH>2.0.CO;2.

Trenberth, K. E., and C. J. Guillemot (1996), Physical processes involved in the 1988 drought and 1993 floods in North America, *J. Clim.*, *9*, 1288–1298, doi:10.1175/1520-0442(1996)009<1288:PPIITD>2.0.CO;2.

Weaver, S. J., and S. Nigam (2008), Variability of the Great Plains low-level jet: Large-scale circulation context and hydroclimate impacts, *J. Clim.*, *21*, 1532–1551, doi:10.1175/2007JCLI1586.1.

Wettstein, J. J., and L. O. Mearns (2002), The influence of the North Atlantic–Arctic oscillation on mean, variance, and extremes of temperature in the northeastern United States and Canada, *J. Clim.*, *15*(24), 3586–3600.

S. Feng and Q. Hu, School of Natural Resources and Department of Geosciences, University of Nebraska at Lincoln, 707 Hardin Hall, Lincoln, NE 68583-0987, USA. (qhu2@unl.edu)