Multidecadal Variability of North China Aridity and Its Relationship to PDO during 1900–2010

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ABSTRACT

North China has undergone a severe drying trend since the 1950s, but whether this trend is natural variability or anthropogenic change remains unknown due to the short data length. This study extends the analysis of dry–wet changes in north China to 1900–2010 on the basis of self-calibrated Palmer drought severity index (PDSI) data. The ensemble empirical mode decomposition method is used to detect multidecadal variability. A transition from significant wetting to significant drying is detected around 1959/60. Approximately 70% of the drying trend during 1960–90 originates from 50–70-yr multidecadal variability related to Pacific decadal oscillation (PDO) phase changes. The PDSI in north China is significantly negatively correlated with the PDO index, particularly at the 50–70-yr time scale, and is also stable during 1900–2010. Composite differences between two positive PDO phases (1922–45 and 1977–2002) and one negative PDO phase (1946–76) for summer exhibit an anomalous Pacific–Japan/East Asian–Pacific patternlike teleconnection, which may develop locally in response to the PDO-associated warm sea surface temperature anomalies in the tropical Indo-Pacific Ocean and meridionally extends from the tropical western Pacific to north China along the East Asian coast. North China is dominated by an anomalous high pressure system at mid–low levels and an anticyclone at 850 hPa, which are favorable for dry conditions. In addition, a weakened land–sea thermal contrast in East Asia from a negative to a positive PDO phase also plays a role in the dry conditions in north China by weakening the East Asian summer monsoon.

1. Introduction

Drought, of the most damaging extreme climate events and natural disasters, causes tens of billions of dollars in damages and affects millions of people worldwide each year (Wilhite 2000). North China is one of the main industrial and agricultural production bases in the country. In recent decades, this region has been facing increased drought that has caused enormous economic losses each year (Fu and An 2002; Fu and Ma 2008). To alleviate the water shortage problem in north China, the central government has implemented the South–North Water Diversion Project. Therefore, understanding the reasons for the recent drought in north China and predicting its changes in the future 10–30 years is of vital importance to policymakers.

Previous studies have focused mainly on decadal-to-interdecadal time scale drought in north China since the 1950s (Hu et al. 2003; Gong et al. 2004; Dai et al. 2005; Yang et al. 2005; Ju et al. 2006; Ma 2007; Li et al. 2010). This period is associated with a weakening of the East Asian summer monsoon circulation, which leads to deficient summer rainfall in north China but excessive rainfall in central east China along the Yangtze River valley. This rainfall pattern is usually termed as “southern flooding and northern drought” in China (e.g., Yatagai and Yasunari 1994; Hu et al. 2003; Yu et al. 2004; Yu and Zhou 2007; Zhou et al. 2009a; Li et al. 2010). Many previous studies have suggested that the weakening tendency of the East Asian summer monsoon is forced by either a phase transition of the Pacific decadal oscillation
(PDO) (Yang et al. 2005; Ma 2007; Li et al. 2010) or recent warming of the tropical Pacific and Indian Oceans (Hu 1997; Chang et al. 2000; Yang and Lau 2004; Zhou et al. 2006; Zhou et al. 2009b; Li et al. 2010). Atmospheric general circulation model simulations forced by historical sea surface temperature (SST) during 1950–2000 demonstrated that the warming over tropical oceans, which is the tropical branch of PDO, has played a major role in the weakening of East Asian summer monsoon circulation during recent decades (Li et al. 2010). Other studies have linked the rainfall changes to human activities such as increased aerosols from pollution (e.g., Menon et al. 2002) and human-induced land cover changes (e.g., Fu 2003). However, no consensus on the mechanisms of the East Asian summer monsoon weakening during 1950–2000 has been reached thus far [see Zhou et al. (2009a) for a review]. The argument on the mechanisms of interdecadal-scale changes of drying trends in north China is partly related to the limitation of data length because most studies are based on climate data obtained after 1950. There are only a limited number of studies that have extended analysis based on instrumental data in the entire twentieth century. For example, Ma and Shao (2006) discussed the statistical relationship between the annual dry–wet evolution in north China and PDO by using a surface humidity index for data recorded during 1901–2002. Zhang and Zhou (2011) compared the change of summer rainfall in China during 1901–2001 to that in other monsoon regions. However, the physical mechanisms were not investigated in these studies. Furthermore, the role of multidecadal climate variability in periods of 60–80 yr (Schlesinger and Ramankutty 1994; Wu et al. 2011) or 50–70 yr (Minobe 1997, 1999) in dry–wet evolution, particularly for the recent drying trend in north China, remains unclear. The main motivation of the present study is to examine the evolution of dryness–wetness in north China during 1900–2010 and to investigate its relationship with the PDO phase transition at the multidecadal time scale. Self-calibrated Palmer drought severity index (PDSI) data (Dai 2011a,b) is used, which is the most prominent index of meteorological drought and considers both the cumulative effect of precipitation and temperature change on the surface water balance between atmospheric water supply (precipitation) and demand (potential evapotranspiration). The multidecadal variability is identified by applying an adaptive and temporally local data analysis tool known as the ensemble empirical mode decomposition (EEMD) method (Wu et al. 2009; Huang and Wu 2008). The results show that approximately 70% of the drying trend during 1960–90 in north China originates from 50–70-yr multidecadal climate variability, which is related to the PDO variability mode. The negative correlation of north China PDSI with the PDO index is stable at the multidecadal time scale during the entire twentieth century. Our result is a useful implication for 10–30-yr decadal prediction (Meehl et al. 2009).

The remainder of the paper is organized as follows. In section 2, we describe the data and methods used in this study. The results are presented in section 3, and a summary with further discussion is given in section 4.

2. Data and methods

a. Data description

PDSI data used in this study are monthly self-calibrated PDSI with potential evapotranspiration estimated by using the more sophisticated Penman–Monteith equation (sc_PDSI_pm) based on historical data (Dai 2011a). We use an updated version of these data for the period 1850–2010 (which is archived at http://www.cgd.ucar.edu/cas/catalog/climind/pdsi.html). The data are available on a 2.5° × 2.5° grid and cover global land area between 60°S and 77.5°N. These data are calculated from observed monthly precipitation and surface air temperature data where available for the period 1850–2010 and are self-calibrated, that is, calibrated by using local climate conditions instead of using the (fixed) coefficients used by Palmer (1965) based on data from the central United States. Thus, the self-calibrated PDSI is more spatially comparable than the original PDSI (Dai 2011a).

The monthly PDO index, defined as the leading standardized principal component of monthly SST anomalies in the North Pacific Ocean, poleward of 20°N, is obtained from the University of Washington (http://jisao.washington.edu/pdo/PDO.latest). The monthly mean global average SST anomalies have been removed to separate this pattern of variability from any global warming signal that may be present in the data. Details of this index can be found in Zhang et al. (1997) and Mantua et al. (1997). In addition, monthly SST data during 1900–2010 from the Hadley Centre Sea Ice and Sea Surface Temperature dataset version 1.1 (HadISST1.1) (Rayner et al. 2003) are used with a grid resolution of 1.0° × 1.0°.

To examine the roles of precipitation and surface air temperature in the multidecadal variability of the PDSI, the following datasets are used.

1) Updated precipitation data originally merged by Dai (2011a) and the Climate Research Unit (CRU) temperature product version 3 (CRUTEM3) temperature anomalies (Brohan et al. 2006) (archived at http://www.cru.uea.ac.uk/cru/data/temperature/) for
the period 1850–2010, which are used to calculate the sc_PDSI_pm (Dai 2011a). In these precipitation datasets, the monthly anomaly data are adopted from Dai et al. (1997) for the period 1850–1947 and from Chen et al. (2002) for 1948–78, in addition to an updated version from Huffman et al. (2009) for 1979–2010 based on Global Precipitation Climatology Project version 2.2 (GPCP v2.2). Details on these merged datasets, hereafter referred to as Dai dataset, are discussed in Dai (2011a).

2) Precipitation and surface air temperature data from the CRU TS3.1 datasets for the period 1901–2009 (Harris et al. 2014), which are available on high-resolution 0.5° × 0.5° grids. For precipitation, we use the new data files from version 3.10.01 because a systematic error was discovered in the CRU TS v3.10 generation of precipitation, wet days, and frost frequency data files (http://badc.nerc.ac.uk/view/badc.nerc.ac.uk__ATOM__dataent_1256223773328276).

3) Global Historical Climatology Network version 2 (GHCN v2) monthly precipitation anomaly dataset on a 5° by 5° basis for the period 1900–2010, which is provided by the National Oceanic and Atmospheric Administration/National Climatic Data Center (NOAA/NCDC) (http://www.ncdc.noaa.gov/temp-and-precip/ghcn-gridded-products.php).

4) Precipitation data developed by Global Precipitation Climatology Centre full data reanalysis version 6.0 (GPCC v6) with a spatial resolution of 0.5° × 0.5° for the period 1901–2010 (Schneider et al. 2011) (which are archived at ftp://ftp.dwd.de/pub/data/gpcc/html/fulldata_v6_doi_download.html).

In addition to observational data, monthly Twentieth Century Reanalysis version 2 data (20CR) (Compo et al. 2011) for the period 1900–2010 with a grid resolution of 2.0° × 2.0° are used to investigate the atmospheric circulation changes controlling the dry–wet conditions in north China. 20CR is the output of the atmospheric component of the National Centers for Environmental Prediction (NCEP) operational Climate Forecast System model driven by the time-evolving SST and sea ice concentration fields of the HadISST1.1 dataset (Rayner et al. 2003). The historical record of sea level pressure (SLP) was assimilated (Compo et al. 2011). The SLP, 850-hPa winds (UV850), 500-hPa geopotential height, 350-hPa meridional wind (V350), 200-hPa zonal wind (U200), air temperature from 1000 to 500 hPa, and surface pressure data derived from 20CR are used. Gaussian gridded (192 × 94) data of the precipitation rate at the surface forecasted by 20CR are also used. In addition, the data from the 40-yr European Centre for Medium-Range Weather Forecasts Re-Analysis (ERA-40) (Uppala et al. 2005) are used for comparison with 20CR for the summer climatology of atmospheric circulation.

b. Analysis method

As shown in Fig. 1a, we calculated the area weighted average of the sc_PDSI_pm data over the north China region (35°–42.5°N, 110°–117.5°E) to obtain an index time series, hereafter referred to as NC-PDSI. This region is consistent with Ma and Shao (2006) and Ma (2007). For consistency with the PDO index, which starts in 1900, the analysis period for NC-PDSI is 1900–2010 (Fig. 1b). The classes for wet and dry periods are defined on the basis of PDSI values (Palmer 1965) in Table 1. In addition, the precipitation and surface air temperature anomalies averaged over north China are calculated and hereafter referred to as NC-P and NC-T. All anomalies in this study are calculated relative to the 1961–90 base period, consistent with CRUTEM3.

The ensemble empirical mode decomposition (EEMD) method (Wu and Huang 2009; Huang and Wu 2008), an adaptive and temporally local filter (e.g., Wu et al. 2011;
Qian et al. (2011b) developed in recent years, is used to decompose the monthly NC-PDSI and PDO index for the period 1900–2010 into various time-scale components (Table 2). According to the mean periods of these components, we obtain five major time-scale components by summing C1 and C2 as intraannual variability, C3, C4, and C5 as interannual variability, and C6 and C7 as interdecadal variability; taking C8 as multidecadal variability (MDV); and summing the last two components as an adaptive nonlinear trend (Wu et al. 2007). EEMD is also used to decompose annual-mean NC-P and NC-T time series during 1900–2010 to obtain their MDVs and nonlinear trends, which are the sixth and last components of the annual-mean series, respectively.

Moreover, least squares linear trend analysis is applied when a tendency transition is revealed by the EEMD method to estimate the linear trends present before and after the turning point. The statistical significance of a linear trend is assessed by using the rank-based and nonparametric Mann–Kendall statistical test (Mann 1945; Kendall 1955). Since this test requires sample data to be serially independent, a prewhitening approach (von Storch and Navarra 1995) is used to remove the influence of serial correlation from the data prior to its application. The prewhitening is achieved by calculating

$$X_P = X_t - rX_{t-1},$$

where \(X_P\), is the prewhitened value at time \(t\), \(X_t\) is the original value of the time series at time \(t\), and \(r\) is the estimated lag-1 autocorrelation.

A 132-month (11 yr) running mean is used to determine the timings of the phase transitions in the NC-PDSI and PDO index. Because we focus on whether drought does or does not occur, a threshold of \(-1\) is chosen for NC-PDSI according to information listed in Table 1 (Fig. 1b). On the basis of the 132-month running mean, we analyze positive phases of PDO in 1922–45 and 1977–2002 minus the negative phase in 1946–76 (hereafter referred to as PDO+ and PDO−, respectively) composites for SLP, UV\(_{850}\), mid-lower troposphere (from the surface to 500 hPa) vertically integrated air temperature, \(V_{350}\), SST, and precipitation. Data before 1922 are not used in our composite analysis as SST observations over the tropical Pacific are sparse before around 1920 (Dai 2013). The statistical significances of the composite differences are determined by using the two-sided \(t\) test.

### 3. Results

#### a. Characteristics of NC-PDSI

Figure 1b shows that NC-PDSI ranges from \(-7.4\) to 4.7 during 1900–2010. According to the categories listed in Table 1, extreme drought occurred mainly in the early 1900s and near 2000, whereas extremely wet occurred in only 1964 for two months. EEMD results show that NC-PDSI has oscillation periods of 3.4 and 8 months, and 1.4, 2.8, 5.8, 15.9, 27.8, and 55.5 yr (Table 2). The 55.5-yr MDV accounts for 13.1% of the total variance. The peak of this MDV is near 1960, whereas two minima are evident near the early 1930s and 1990s (Fig. 1b). Between these two minima, the timing of the phase transition from negative (positive) to positive (negative) is near 1945–46 (1976–77). This MDV is superimposed on a nonlinear trend, which contributes to 15.0% of the total variance and includes a transition from a wetting to drying trend near the late 1950s, resulting in a drying tendency present since the 1960s. The turning point at 1959/60 is determined from the timing when the first-order derivative equals zero (Qian et al. 2011a) in the combination time series of the MDV and nonlinear trend. Linear trends for 1900–59 and 1960–2010 in the monthly NC-PDSI are 0.44 and \(-0.81\) decade\(^{-1}\), respectively, both of which are statistically significant at the 0.05 level when autocorrelation is considered.

<table>
<thead>
<tr>
<th>Value</th>
<th>Classes</th>
<th>Value</th>
<th>Classes</th>
</tr>
</thead>
<tbody>
<tr>
<td>≥4.00</td>
<td>Extremely wet</td>
<td>-1.00 to -1.99</td>
<td>Mild drought</td>
</tr>
<tr>
<td>3.00 to 3.99</td>
<td>Very wet</td>
<td>-2.00 to -2.99</td>
<td>Moderate drought</td>
</tr>
<tr>
<td>2.00 to 2.99</td>
<td>Moderately wet</td>
<td>-3.00 to -3.99</td>
<td>Severe drought</td>
</tr>
<tr>
<td>1.00 to 1.99</td>
<td>Slightly wet</td>
<td>≤ -4.00</td>
<td>Extreme drought</td>
</tr>
<tr>
<td>0.99 to -0.99</td>
<td>Normal</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

### Table 1. Classes for wet and dry periods based on PDSI value.

<table>
<thead>
<tr>
<th>NC-PDSI</th>
<th>PDO</th>
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<tbody>
<tr>
<td>C1</td>
<td>3.4 months (3.8%)</td>
</tr>
<tr>
<td>C2</td>
<td>8 months (7.1%)</td>
</tr>
<tr>
<td>C3</td>
<td>1.4 yr (13.3%)</td>
</tr>
<tr>
<td>C4</td>
<td>2.8 yr (13.6%)</td>
</tr>
<tr>
<td>C5</td>
<td>5.8 yr (13.6%)</td>
</tr>
<tr>
<td>C6</td>
<td>15.9 yr (12.1%)</td>
</tr>
<tr>
<td>C7</td>
<td>27.8 yr (8.5%)</td>
</tr>
<tr>
<td>C8</td>
<td>55.5 yr (13.1%)</td>
</tr>
</tbody>
</table>

| Trend | (15.0%) | (2.8%) |

### Table 2. The mean periods and their variance contributions (in parentheses) of various time-scale components for monthly NC-PDSI and for PDO during 1900–2010 obtained by the EEMD method, respectively.
drying trend during 1960–2010 is a well-known phenomenon referred to as northern drought. During 1960–90, the drying rate of approximately $1.07 \text{ decade}^{-1}$ is relatively faster owing to reinforcement of the MDV component to the nonlinear trend. From January 1960 to December 1990, the combination of the MDV and nonlinear trend monotonically decreases by 2.57, whereas the MDV component monotonically decreases by 1.8. Thus, during this period, the MDV component accounts for approximately 70% of the drying tendency represented by the combination of the MDV and nonlinear trend.

Note that the sc_PDSI_pm is calibrated by using local climate conditions during 1950–79 (Dai 2011a). During this period, the mean of NC-PDSI is $-0.15$, and the shape of the distribution is Gaussian at a 95% confidence level when the Jarque–Bera test (Steinskog et al. 2007) and quantile–quantile plotting are applied to test normality (Figs. 2a,b). However, the long-term mean of NC-PDSI during 1900–2010 is $-1.5$, indicating that the long-term moisture demand in north China, which is located in a semiarid zone, is greater than the moisture supply. Although the monthly NC-PDSI for the period 1900–2010 is positively skewed slightly at 0.204, the skewness of the summer, June–August (JJA), NC-PDSI is nearly zero at 0.039. The results of the Jarque–Bera test and quantile–quantile plotting (Figs. 2c–d) also show a normal distribution of the JJA NC-PDSI.

According to the 132-month running mean (Fig. 3), the decades from the mid-1900s to late 1940s and those from the late 1970s to the mid-2000s includes dry conditions, whereas those from the late 1940s to the mid-1970s are normal, as can be seen from Fig. 3: the values of the running mean are all below 1. Since the early 1980s, north China has experienced a moderate drought, as indicated by the NC-PDSI value. However, the drying tendency is alleviated and tended to level off after the late 1990s. All of these characteristics are consistent with the results obtained from EEMD analysis (Fig. 1b), which further demonstrates that this leveling off is due to the reversal of multidecadal variability since the early 1990s.
b. Relationship between NC-PDSI and PDO

Two major internal variability modes at multidecadal time scales include the Atlantic multidecadal oscillation (AMO) with a period of 65–70 yr (e.g., Schlesinger and Ramankutty 1994) and the PDO with a period of 50–70 yr (e.g., Minobe 1997, 1999). The phase of AMO is quite different from that of NC-PDSI (figure not shown). However, an inverse relationship is evident between PDO and NC-PDSI, with PDO\(^1\) (PDO\(^2\)) corresponding to a dry (normal) period in north China, particularly after 1921 when the SST data over the tropical Pacific are more reliable (Fig. 3). The correlation coefficient between monthly (annual) original unfiltered NC-PDSI and PDO during 1900–2010 is \(-0.23\) \((-0.31)\), which is statistically significant at the 0.05 level by the Student’s \(t\) test. For monthly data, the significance is determined by an effective degree of freedom of 237 when considering the autocorrelations. This effective degree of freedom is calculated as

\[
N_{\text{edof}} = N \frac{1 - r_1 r_2}{1 + r_1 r_2},
\]

where \(N\) is the original sample size of 1332, and \(r_1\) (\(r_2\)) are the lag-1 autocorrelation of the first (second) time series (Bretherton et al. 1999). In addition, when the monthly NC-PDSI and PDO are both linearly detrended, their correlation coefficient \((-0.24)\) is slightly larger and more significant than using the original data. Although the monthly NC-PDSI for the period 1900–2010 is slightly skewed, the significant correlation between NC-PDSI and PDO is robust. The coefficient of the non-parametric Spearman correlation between the monthly unfiltered NC-PDSI and PDO for the period 1900–2010 is \(-0.22\), which is significant at the 0.05 level with an effective degree of freedom of 237. The commonly used Pearson correlation coefficient between JJA NC-PDSI, which is normally distributed, and PDO for the period 1900–2010 is \(-0.25\) and is also statistically significant at the 0.05 level with an effective degree of freedom of 85. The statistically significant inverse relationship between the dry–wet conditions in north China and PDO are consistent with that reported by Ma and Shao (2006), who analyzed a humidity index defined as the difference between the observed annual precipitation and the sum of potential evapotranspiration in each month.

The negative correlation between NC-PDSI and PDO is particularly prominent at the 50–70-yr time scale with a mean period of approximately 56 yr, as revealed by the EEMD result (Fig. 4a). The timing of the phase transition in the 50–70-yr oscillation of PDO, which occurred near 1945/46 and 1976/77, agree well with those of NC-PDSI. The 50–70-yr oscillation in PDO revealed by EEMD is consistent with that reported by Minobe (1997, 1999), who used the multitaper method and wavelet analysis to detect the 50–70-yr oscillation over the North Pacific in addition to a 15–25-yr oscillation. As indicated in Table 2, this 50–70-yr oscillation accounts for 16% of the total PDO variance, comparable to that of the 15–25-yr oscillation as represented by the combination of C6 and C7 of PDO but is significantly larger than the contribution from the trend, which is only 2.8%. This result is because the monthly-mean global average SST anomalies have been removed in calculating this PDO index to filter out “global warming” signals (Mantua et al. 1997). The 660-month (55 yr) running correlations show that the inverse relationship between monthly NC-PDSI and PDO over the entire analysis period is quite stable (Fig. 4b), which further confirms the EEMD result shown in Fig. 4a.

c. Possible mechanisms

1) ROLE OF PRECIPITATION AND TEMPERATURE

Because the PDSI is calculated on the basis of precipitation and temperature data, these two variables over the north China region are examined separately to determine their relative roles. Figure 5a shows that the annual-mean NC-P from the Dai dataset agrees well with those from GHCN v2, GPCC v6, and CRU TS3.10.01. The correlation coefficients between the annual NC-P from the Dai dataset and those from the other datasets
are 0.92 (1900–2010), 0.97 (1901–2010), and 0.87 (1901–2009). Although the correlation coefficient between the annual NC-P from the Dai dataset and that from CRU TS3.10.01 over the entire analysis period is relatively small: the value increases to 0.92 after 1916. The relatively large difference between these two datasets during the early years occurred because interpolation is used by CRU TS3.10.01 even in the absence of nearby station data, whereas the Dai dataset uses only stations available before 1948 (Dai et al. 1997). The similarities among the Dai dataset, GHCN v2, and GPCC v6 during the early years add reliability to the Dai dataset. For the years after 1997, there is a decline in the rain gauge number in the CRU datasets due to its heavy dependence on the GHCN station data set; however, the GPCP v2.2 dataset, which is used in the Dai dataset, uses a nearly fixed station network over land from GPCC for the period 1979–2010 (Dai 2011a). Therefore, the Dai dataset is reliable over the entire 1900–2010 period.

The ensemble empirical mode decomposition analysis of the Dai dataset indicates a secular change from increasing to decreasing trend near 1959/1960. The linear trend in annual-mean NC-P is 1.7 mm decade$^{-1}$ during

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**Fig. 4.** (a) Decompositions of monthly NC-PDSI (red) and PDO index (blue) during 1900–2010 into five major time scales determined using the EEMD filter. In the last subpanel, the nonlinear trends of NC-PDSI and PDO have been both subtracted by their mean values to facilitate comparison. (b) The 55-yr (660 month) running correlations between monthly NC-PDSI and PDO index during 1900–2010 are plotted as the solid line. The dashed line indicates the 0.05 significant level with an effective degree of freedom of 117. A negative correlation coefficient below this line indicates significant correlation.

**Fig. 5.** (a) Annual-mean NC-P anomaly during 1900–2010 from the Dai dataset (red) and its MDV and Trend obtained by using the EEMD method. The MDV has been subtracted by a value of 15 for visual purposes. To validate this precipitation data, precipitation from GHCN version 2 (green), GPCC version 6 (orange), and CRU TS3.10.01 (blue) and their corresponding MDVs are also plotted (mm month$^{-1}$). (b) Annual-mean NC-T anomaly during 1901–2009 from CRU TS3.10 (red) and its MDV and Trend. The MDV has been subtracted by a value of 1 for visual purposes. The NC-T anomaly during 1900–2010 from the CRUTEEM3 dataset (blue) is also plotted for comparison, although bias occurs in the early 1900s ($^\circ$C).
1900–59 and $-1.3\,\text{mm}\,\text{decade}^{-1}$ during 1960–2010, both of which are statistically significant at the 0.05 level. In addition to the secular change, a prominent 50–70-yr MDV is identified with a mean period of approximately 56 yr. The peak of this MDV is near 1960, whereas two minimums are near the early 1930s and mid-1990s. This MDV is also evident in the other three precipitation datasets. The MDV of NC-P is consistent with that of the in-phase NC-PDSI.

The annual-mean temperature variability derived from the CRU TS3.10 and CRUTEM3 datasets is similar (Fig. 5b), and the correlation coefficient between them is 0.93 during 1901–2009. One exception is seen in the early 1900s when the CRUTEM3 data overestimate the annual-mean temperature. In our analysis, we use CRU TS3.10 data to examine the changes in NC-T. The linear trend in annual-mean NC-T during 1901–2009 is $0.13\,\text{°C}\,\text{decade}^{-1}$, which is statistically significant at the 0.05 level. EEMD analysis of NC-T indicates a secular nonlinear warming trend, which leads to a warming of $1.4\,\text{°C}$ during 1901–2009. Superimposed on this warming trend is an MDV with a discernible peak (valley) near the late 1930s (mid-1970s). The phase of the MDV in NC-T differs from that in NC-PDSI. A comparison of Figs. 1b and 5 indicates that the MDV in NC-PDSI is dominated by precipitation.

2) MECHANISM LINKING NC-P AND PDO

The Dai dataset indicates that for precipitation over the north China region, JJA precipitation contributes to nearly two-thirds of the annual precipitation, with 62% during 1900–2010 and 69% during 1951–2010. We therefore perform a composite analysis of PDO+ (1922–45 and 1977–2002) minus PDO− (1946–76) for JJA mean precipitation (Fig. 6). Both high-resolution datasets, GPCC v6 and CRU TS3.10.01, show less precipitation in north China (cf. Figs. 6a,b) and more precipitation in the mid and lower reaches of the Yangtze River from PDO− to PDO+. A further examination of separate periods for PDO+ and PDO−, shown in Figs. 6c–h, reveals that both PDO+ periods (PDO− period) correspond to anomalously less (more) precipitation in north China. These features are consistent between the GPCC and CRU datasets. Zhu et al. (2011) examined the recent decadal change in JJA precipitation in east China in the most recent PDO− phase after 2000 on the basis of 160 station observations and reported that the rainfall in the southern part of north China (Huang–Huai River region) increased again. Their study supports our result, such that anomalous JJA rainfall in north China is closely associated with PDO phase transition.

To determine the atmospheric circulation changes associated with PDO phase transition, we analyze the composites of PDO+ (1922–45 and 1977–2002) minus PDO− (1946–76) for four atmospheric variables of summer SLP, UV$_{850}$, mid–lower tropospheric mean air temperature, and $V_{350}$ from 20CR (Figs. 7a–d). The climatology of the East Asian summer monsoon circulation derived from 20CR is first compared with that of ERA-40 (figure not shown). The major circulation systems are consistent, including the Mascarene high and the Australian high in the Southern Hemisphere, the Indian low and the subtropical high over the North Pacific, and the cross-equatorial flow and southwesterly wind into mainland China at the 850-hPa level. The 20CR also well represents the 500-hPa geopotential height, particularly the western Pacific subtropical high, which is crucial in controlling the water vapor transport of the summer monsoon (Zhou and Yu 2005), and the East Asian subtropical westerly jet at 200 hPa (Zhang et al. 2006). This comparison indicates that the performance of 20CR is comparable to that of ERA-40 in describing East Asian summer monsoon circulation.

Composite differences (Fig. 7) clearly show a deep anomalous wave train from sea level to the upper troposphere with a meridional pattern along the eastern coastal area of China from PDO− to PDO+, resulting in anomalous high pressure (anticyclone circulation) over the Philippine Sea, anomalous low pressure (cyclone circulation) over the southeastern coastal area of mainland China near 20°N, 120°E, and anomalous high pressure (anticyclone circulation) over north China near 40°N, 120°E in SLP (UV$_{850}$) (Figs. 7a,b). This anomalous meridional wave train causes anomalous northeasterly winds over southern China, resulting in a weakened summer monsoonal flow over the region (Fig. 7b). This wave train is similar to the Pacific–Japan/East Asia–Pacific (PJ/EAP) pattern of teleconnection (Nitta 1987; Huang and Sun 1992). To trace the origin of this wave train, we investigate the PDO+ minus PDO− composite $V_{350}$, SST, and 20CR precipitation (Figs. 7d–f) because the upper-tropospheric meridional wind perturbations provide useful diagnostics for tracing the origins and regions of influence of Rossby waves (Jin and Hoskins 1995; Hu and Huang 2009). The high-latitude equilibrium response to low-latitude SST anomalies can be explained by Rossby wave propagation forced by convective heating anomalies that develop locally in response to the SST anomalies (Hoskins and Karoly 1981; Hu and Huang 2009). It is shown that the PJ/EAP patternlike wave train begins from the tropical western Pacific Ocean (Fig. 7d) and may be forced by convective heating anomalies (Fig. 7f) that develop locally in response to the PDO-associated warm SST anomalies in the tropical western Pacific Ocean (Fig. 7e). In addition, the tropical Indian Ocean basinwide
warm anomaly (Fig. 7e) may also play a role in the formation of the PJ/EAP patternlike wave train on the multidecadal time scale through two possible mechanisms. The first is through convective heating of the Indian Ocean (Fig. 7f), which forces a Kelvin wave response and suppresses the convection to create favorable anticyclone conditions over the Philippine Sea as proposed by Wu et al. (2009) at the interannual time scale. The other is through the increase of surface moisture over the Indian Ocean and the advection-strengthened convective heating over the Maritime Continent (Fig. 7f), which induces local Hadley circulation and causes anomalous subsidence near the Philippines (Fig. 7f). This mechanism was proposed by Hu (1997) to interpret the impact of the Indian Ocean on summertime interdecadal changes in the East Asian climate. Through either of these two mechanisms, the negative precipitation anomaly or anomalous convective activity near the Philippines plays an important role in the formation of the PJ/EAP patternlike wave train (Huang and Sun 1992; Hu et al. 2011). In summary, under the PDO+ phase, north China is dominated by an anomalous high pressure and anticyclone system in the mid–lower troposphere, which is part of the PJ/EAP patternlike teleconnection. Thus, deficient precipitation is seen in north China.

Fig. 6. (left) Composite of JJA mean precipitation anomaly (mm month$^{-1}$) from the GPCC dataset for (a) PDO+ (1922–45 and 1977–2002) minus PDO− (1946–76), (c) 1922–45, (e) 1946–76, and (g) 1977–2002; (right) composite from the CRU dataset. Values within the contours in (a) and (b) are significant at the 0.05 level.
Beside the wave train, Fig. 7c also indicates that during the PDO+ phase, the mid–lower tropospheric mean air temperature over the tropical western Pacific Ocean and the Maritime Continent is anomalously warm, whereas that over the mid–high latitude land, generally at 40°–60°N, is anomalously cold. This difference indicates a weakened land–sea thermal contrast, which favors weakening of the East Asian summer monsoon. Composite analysis of the PDO+ minus PDO− SST (Fig. 7e) shows that during the PDO+ phase, warm anomalies are seen over nearly all of the tropical equatorial oceans, whereas cold anomalies exist over the North and South...

Fig. 7. The PDO+ (1922–45 and 1977–2002) minus PDO− (1946–76) composite difference in JJA mean (a) SLP (hPa), (b) 850-hPa winds (m s⁻¹), (c) mid–lower troposphere integrated air temperature from the surface to 500 hPa (1000°C kg m⁻²), (d) 350-hPa meridional wind (m s⁻¹), (e) SST from the HadISST1.1 dataset (°C), and (f) precipitation rate from the 20CR dataset (kg m⁻² month⁻¹). In (a), (b), (c), and (d) solid (dashed) lines represent positive (negative) values, and shaded areas are statistically significant at the 0.05 level. In (e) and (f) values above the contour line are significant at the 0.05 level.
Pacific. This pattern is similar to that reported by Li et al. (2010), who performed a numerical study to show that two atmospheric general circulation models driven by the tropical lobe of PDO-related SST anomalies are able to reproduce most of the observed weakening of the East Asian summer monsoon circulation during 1979–1999 through decreasing the land–sea thermal contrast.

4. Summary and discussion

To investigate why north China has undergone a drying trend since the 1950s, we use PDSI data to examine the multidecadal variability and changes in dry–wet conditions in the region during 1900–2010. In addition, the forcing of PDO to multidecadal variability in dry–wet conditions in north China is analyzed. The main findings are summarized as follows:

1) A secular transition from significant wetting to significant drying trend is detected in both the NC-PDSI and NC-P near 1959/60. The aridification in north China since 1960 is a consequence of the evolution of the 50–70-yr MDV from a positive to negative phase superimposed on a drying nonlinear trend. The MDV component of NC-PDSI, which is related to PDO, accounts for approximately 70% of the northern drought tendency during 1960–90. The 50–70-yr MDV in NC-PDSI is dominated by precipitation.

2) Significant negative correlations are detected between NC-PDSI and PDO. These negative correlations are most robust at time scales longer than 10 yr, particularly at the 50–70-yr time scale, with PDO+ (PDO−) corresponding to drier (wetter) conditions in north China. Their inverse relationship at the multidecadal time scale is stable.

3) The composite differences between two positive PDO phases (1922–45 and 1977–2002) and one negative PDO phase (1946–76) for summer exhibits an anomalous PJ/EAP patternlike meridional teleconnection along the East Asian coast, which may develop locally in response to the PDO-associated warm SST anomalies in the tropical Indo-Pacific Ocean. Following the PJ/EAP teleconnection pattern, north China is dominated by an anomalous high pressure and anticyclone system at mid–low levels, which are favorable for dry conditions in north China. Moreover, a weakened land–sea thermal contrast in East Asia from PDO− to PDO+ also plays a role in the dry conditions in north China by weakening the East Asian summer monsoon.

Limitations of the current study should be acknowledged. First, there are only approximately 1.5 cycles on the multidecadal time scale based on the instrumental data used in this study. Whether the strong correlation between north China aridity and PDO existed before 1900 deserves further study. Future research that uses coupled model long-term control runs or long-term reconstruction data may be helpful for establishing a more robust correlation between north China aridity and PDO. Second, only six stations in the north China region have recorded both observational monthly precipitation and temperature data since 1920, according to the two long-term instrumental climatic databases of the People’s Republic of China (Tao et al. 1997), which was compiled in accordance with a joint research agreement signed by the U.S. Department of Energy and the Chinese Academy of Sciences, China. These six stations are Huhehaote (40.80°N, 111.63°E), Taiyuan (37.78°N, 112.55°E), Beijing (39.93°N, 116.28°E), Tianjin (39.10°N, 117.17°E), Baoding (38.83°N, 115.57°E), and Jinan (36.68°N, 116.98°E). Although these stations are almost evenly distributed in the north China region, the results of the north China area mean before 1949 are not as reliable as those after that time. Finally, the relationship between PDO and AMO on the multidecadal time scale is rather complex. The mismatching phases between NC-PDSI and AMO and between NC-P and AMO cannot guarantee that they are uncorrelated. In this study, we confine our objective to the relationship between NC-PDSI and PDO only.

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