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Winter temperature extremes in China and their possible causes

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ABSTRACT: Cold and warm temperature extremes predominantly occurring in winter gained much more attention than mean temperatures. On the basis of daily maximum and minimum surface air temperature records at 303 meteorological stations in China, the spatial and temporal distributions of five indices for winter (DJF: December, subsequent January and February) temperature extremes are analysed during 1961–2003. For the majority of stations, the frequency of cold days/nights decreases by −1.33/−2.98 and warm days/nights increases by 0.92/2.35 d/decade, respectively. Cold days/nights are significantly negatively correlated with the Arctic Oscillation (AO) index, while warm days/nights are positively correlated with the AO. The diurnal temperature range (DTR) has a declining trend with rate of −0.25 °C/decade and positive correlation with the AO index. Compared with other regions in China, stations in the northern China have larger trend magnitudes and stronger correlations with the AO index, and the AO can explain more than 50% of winter temperature extreme change in China. Compared with the annual basis, the winter temperature extremes have larger trend magnitudes, which reflect the rapid warming. During strongly positive AO index years, enhanced contrast tropospheric temperature (defined as the average of air temperature vertically integrated between 200 hPa and 1000 hPa based on the National Centers for Environmental Prediction/National Center for Atmospheric Research reanalysis) between the north of China and the southern China weakens the East Asian winter monsoon which in turn reduces cold outbreaks in the northern and eastern China. The composites of large-scale atmospheric circulation are consistent with the asymmetrical changes of the geopotential height, zonal and meridional winds at high and mid latitudes at troposphere. Meanwhile, the linkage between the AO and solar activity also modulates the winter temperature extremes, while the mechanism needs to be investigated. Copyright © 2012 Royal Meteorological Society

KEY WORDS temperature extremes; Arctic Oscillation; atmospheric circulation; China

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1. Introduction

Extreme climate events can cause property damage, injury and loss of life and understanding their occurrence is very important to natural and human systems (Katz and Brown, 1992; Easterling *et al.*, 2000; Aguilar *et al.*, 2009). As a consequence, the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) paid more attention to climate extremes change (IPCC, 2007). Recent studies on global, regional and national scales have significantly improved the understanding of temperature and precipitation extremes (Peterson *et al.*, 2002; Aguilar *et al.*, 2005, 2009; Alexander *et al.*, 2006; Klein Tank *et al.*, 2006; New *et al.*, 2006; Brown *et al.*, 2008; Peterson and Manton, 2008; Peterson *et al.*, 2008; You *et al.*, 2008a,b; Choi *et al.*, 2009; Caesar *et al.*, 2011; You *et al.*, 2011a). Most of these studies have been fostered by the World Meteorological Organization (WMO) Joint Expert Team on Climate Change Detection and Indices (ETCCDI) (Peterson and Manton, 2008). They have revealed that cold extremes are generally changing more rapidly than warm extremes, but the exact reasons have not been explored in detail.

The Arctic Oscillation (AO), currently known as Northern Annular Mode, is one of the dominant patterns of Northern Hemisphere climate variability, and it is most prevalent in winter and in the mid and high latitudes. It strongly influences surface air temperatures over the Eurasian continent, especially Europe (Hurrell, 1995; Thompson and Wallace, 1998; 2001; Hurrell *et al.*, 2001; Hurrell and Deser, 2010). AO is a major controlling factor in basic meteorological variables such as surface wind, temperature and precipitation (Bojariu and Gimeno, 2003). AO is defined as a hemispheric mode

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whose dipole has suffered a displacement to the West during the last decades (Ramos *et al.*, 2010). The AO index has been used to describe the variability of AO in this study.

Recent studies have shown that the winter AO index has a strong positive correlation with temperatures in northern China (Gong and Wang, 2003) and is also correlated with the strength of the East Asian winter monsoon and Siberian Higher pressure system (Gong *et al.*, 2001; Wu and Wang, 2002). Since the 1980s, China has experienced significant temperature increases (Wang and Gong, 2000; Ding *et al.*, 2007), and warming is projected to continue. Although trends in temperature extremes on the annual basis have been studied (Zhai *et al.*, 1999; Zhai and Pan, 2003; Ren *et al.*, 2011; You *et al.*, 2011a), there have been little investigations of how the AO influences winter temperature extremes. Thus we quantify changes in winter temperature extremes during 1961–2003 throughout China, based on indices designed by the Commission for Climatology/Climate Variability and Predictability/Joint WMO Intergovernmental Oceanographic Commission Technical Commission for Oceanography and Marine Meteorology ETCCDI. The relationships between the AO index and winter temperature extremes are also examined.

2. Data and methods

Daily maximum and minimum temperatures for 303 stations in China are provided by the National Meteorological Information Center, China Meteorological Administration. Both the spatial density of stations and the quality of observational data in China meet the World Meteorological Organization's standards. Stations were selected according to procedures described in our recent papers (You *et al.*, 2011a). The selected stations should have the long-term data records and good data quality. The calculation of indices is facilitated using the information provided by ETCCDI (see http://cccma.seos.uvic.ca/ETCCDI for available calculated station-level indices) (Peterson and Manton, 2008). We concentrate on the winter (DJF) variation of five temperature indices (Table I), which have been shown to be most sensitive to climate change in previous studies (You *et al.*, 2008a, 2011a). The winter temperature extremes have the same definition as in previous studies (Aguilar *et al.*, 2005, 2009; Alexander *et al.*, 2006; Klein Tank *et al.*, 2006; New *et al.*, 2006; You *et al.*, 2008a,b; Caesar *et al.*, 2011; You *et al.*, 2011a). RClimDex software was used to perform data quality control and calculate the indices, and RHtest was used to assess homogeneity. Details about data quality control and homogeneity tests are described in our previous papers (You *et al.*, 2008a, 2011a).

The AO index is defined as the difference in the normalized monthly zonal-mean sea level pressure (SLP) between 35 and 65°N (Li and Wang, 2003), derived from http://web.lasg.ac.cn/staff/ljp/data-NAM-SAM-NAO/NAM-AO.htm. Monthly mean geopotential height, air temperature, zonal and meridional wind were obtained from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis (available from their website at http://www.cdc.noaa.gov/) (Kalnay *et al.*, 1996). The relationships between solar activity and winter temperature extremes are studied in the study and the solar activity is derived from the studies in Kodera (2002) and Ogi *et al*(2003).

The Mann–Kendall test for trends and Sen's slope estimates are used to detect and quantify trends in winter temperature extremes (Sen, 1968), with magnitudes of trends and slopes assessed at the 0.05 significance level ($p < 0.05$).

3. Results

3.1. Winter temperature extremes (TX10, TN10, TX90, TN90 and DTR)

Figure 1 shows the spatial patterns of trend for five winter temperature indices (for 303 meteorological stations) along with the time series of the entire country. Aggregated regional trends of winter temperature extremes are listed in Table II (third column), calculated as the arithmetic mean of all station. The number of stations with negative, no trend and positive trends, as well as the number of stations passing the significant level for each index is also shown in Table II.

For cold days (TX10) and cold nights (TN10), about 97 and 98% of stations have decreasing trends, whereas 31 and 84% of stations are statistically decreasing trends. For TX10, stations in the northern China (such as Gansu province) show larger trend magnitudes, and significant

Table I. Definitions of five winter temperature indices used in this study. All indices are calculated by RClimDeX software.

Index	Descriptive name	Definition	Units
Temperature			
TX10	Cold day frequency	Percentage of days when TX < 10th percentile of 1961–1990	%
TN10	Cold night frequency	Percentage of days when TN < 10th percentile of 1961–1990	%
TX90	Warm day frequency	Percentage of days when TX > 90th percentile of 1961–1990	%
TN90	Warm night frequency	Percentage of days when TN > 90th percentile of 1961–1990	%
DTR	Diurnal temperature range	Annual mean difference between TX and TN	°C

TX is the daily maximum temperature; TN is the daily minimum temperature.

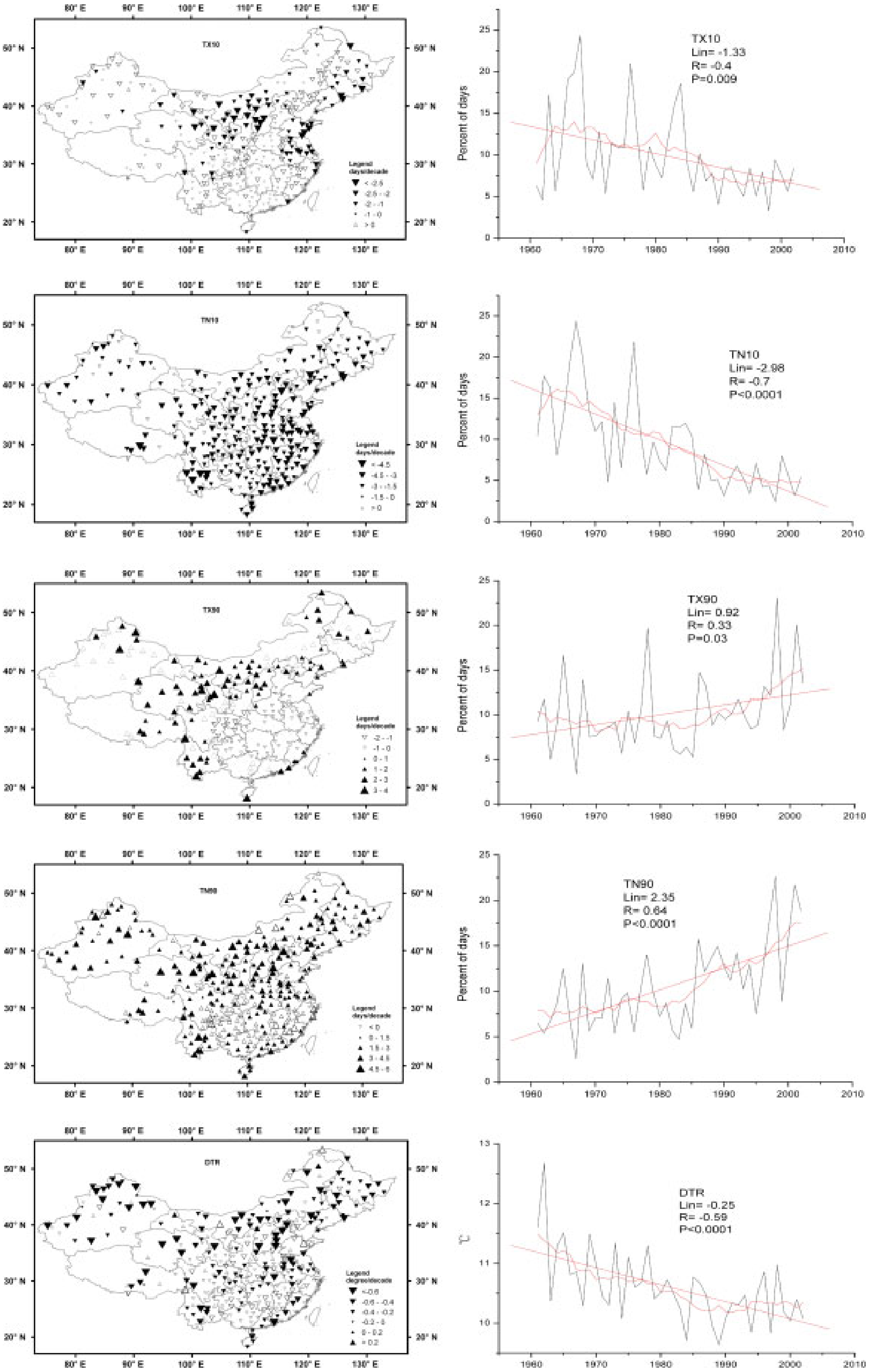

Figure 1. Spatial patterns of trends per decade and series of winter temperature indices (TX10, TN10, TX90, TN90 and DTR) during 1961–2003 in China. Positive/negative trends are shown as up/down triangles, and the filled symbols represent statistically significant trends (significant at the 0.05 level). The size of the triangles is proportional to the magnitude of the trends. The smoother line is the 9 year smoothing average. This figure is available in colour online at wileyonlinelibrary.com/journal/joc

decreasing trends are shown over most regions in China for TN10 (Figure 1). TX10 has shown some slight increasing changes since 1990, but the decrease in TN10 has been much more consistent before the 1990s, with only a slight levelling off after that. The countrywide trend (in % of days) for these two indices are −1.33 and −2.98 d/decade, respectively ($p < 0.05$).

For the percentage of days exceeding the 90th percentiles (TX90 and TN90), about 79 and 98% of stations have increasing trends, and about 30 and 70% of stations show statistically significant and increasing trends. Stations in the northern China show larger trend magnitudes for both TX90 and TN90 (Figure 1). Some stations in the southern China have decreasing

Table II. Trends per decade (with 95% confidence intervals in parentheses), and the number of stations with positive (significant at the 0.05 level), non-trend, and negative (significant at the 0.05 level) trends for winter temperature indices in the entire country.

Index	Units	Trends	Positive	Non-trend	Negative
TX10	d/decade	−1.33 (−2.70 to −0.25)	4 (1)	5	294 (93)
TN10	d/decade	**−2.98** (−3.96 to −1.90)	4 (1)	1	298 (254)
TX90	d/decade	0.92 (0.05 to 1.85)	238 (90)	2	63 (4)
TN90	d/decade	**2.35** (1.30 to 3.27)	298 (213)	0	5 (1)
DTR	°C/decade	**−0.25** (−0.39 to −0.14)	40 (4)	0	263 (141)

Values for trends significant at the 5% level (t-test) are set in bold.

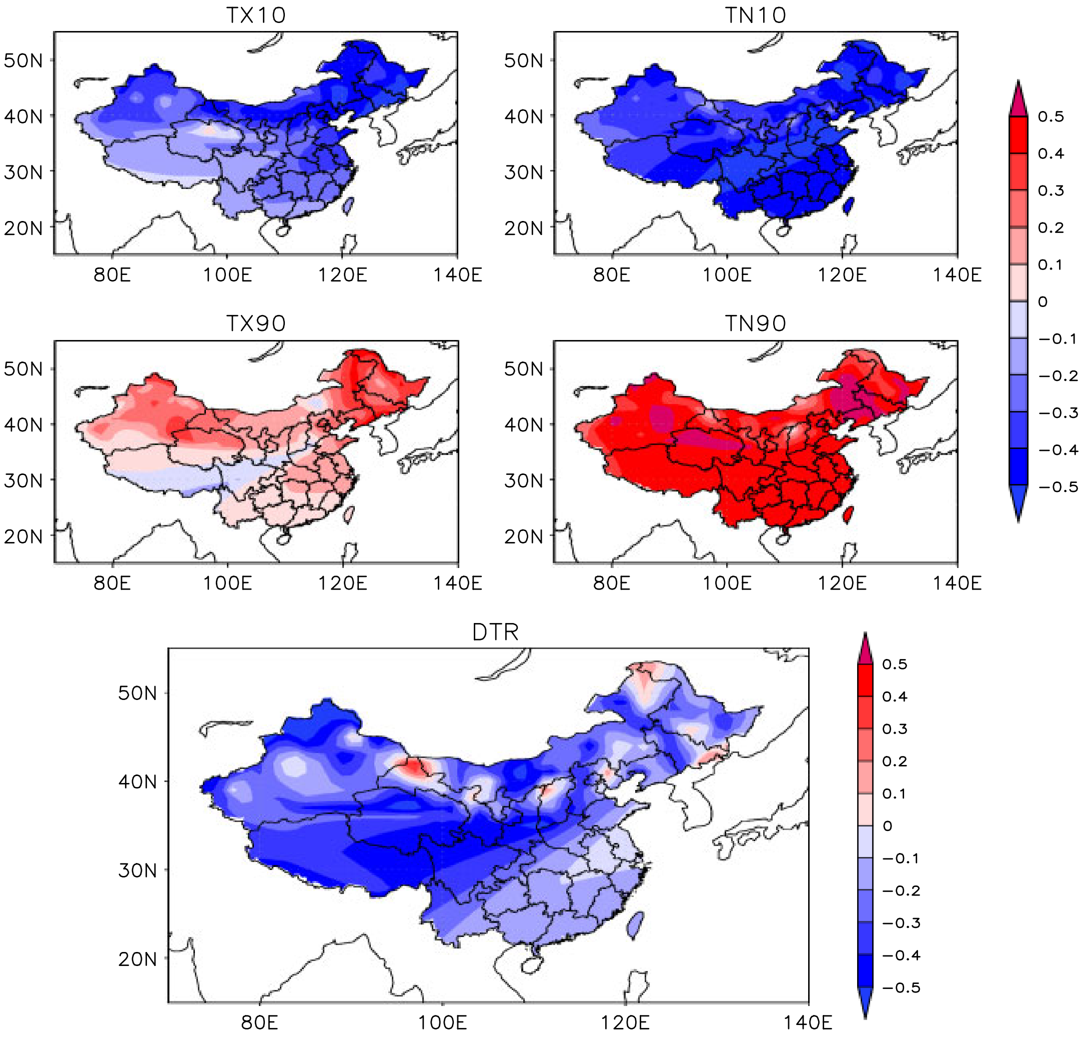

Figure 2. Spatial correlation coefficients between winter temperature indices (TX10, TN10, TX90, TN90 and DTR) and winter AO index during 1961–2003 in China. This figure is available in colour online at wileyonlinelibrary.com/journal/joc

trends for TX90 and non-significant increasing trends for TN90. Before the mid-1980s, both TX90 and TN90 have fluctuant (decreasing and increasing) changes, and show statistically increasing trends after that. The trends in the entire country for these two indices are 0.92 and 2.35 d/decade, respectively ($p < 0.05$).

For diurnal temperature range (DTR), about 87% of stations show decreasing trends, while 47% of stations decrease significantly. Similarly, stations in the northern China between 40° and 50°N show larger trend magnitudes, where have more pronounced warming. This illustrates that more warming leads to larger decreases for DTR (You *et al.*, 2008a, 2011a). DTR has shown a significant decreasing trend before the 1990s, with only a slight levelling off after that 1990. The overall trend in the entire country for DTR is −0.25 °C/decade ($p < 0.05$), which is larger than the annual DTR trend in the Tibetan Plateau (−0.20 °C/decade) during 1961–2005 (You *et al.*, 2008a) and entire China (−0.18 °C/decade) during 1961–2003 (You *et al.*, 2011a).

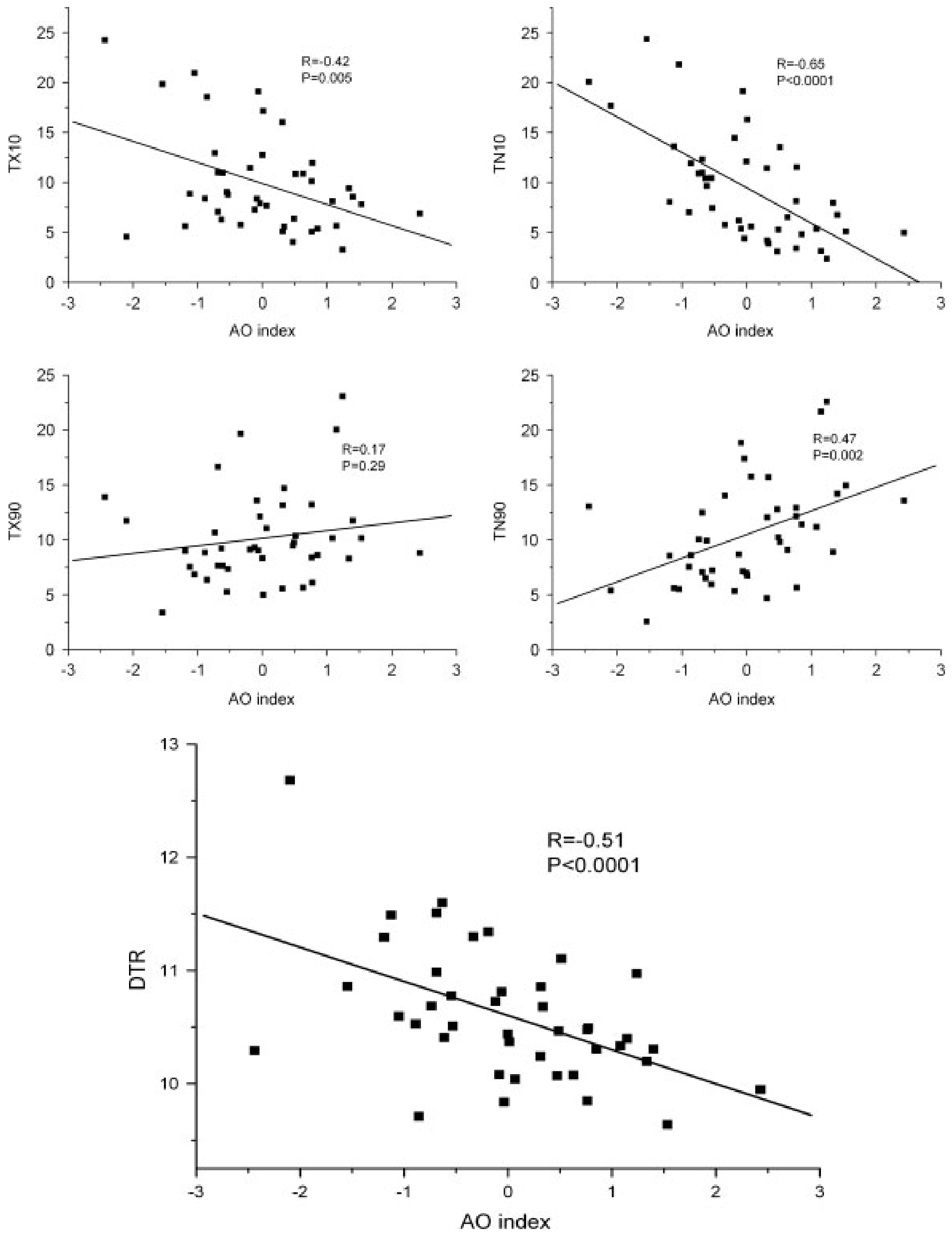

Figure 3. Correlations between winter temperature indices (TX10, TN10, TX90, TN90 and DTR) and the AO index during 1961–2003 in China. All dots in each panel are the mean values for the entire China. The straight lines, R and P are linear fits, correlation coefficients and statistical significance, respectively.

3.2. Comparison with the annual temperature extremes

In the previous study, the spatial and temporal distributions of temperature extremes on the annual basis have been analysed using the same datasets (You *et al.*, 2011a). Countrywide, the annual trends for TX10, TN10, TX90, TN90 and DTR are −0.47 d/decade, −2.06 d/decade, 0.62 d/decade, 1.75 d/decade, −0.18 °C/decade, respectively. For TX10, TN10 and DTR, about 77, 97, 80% of stations have decreasing trends, and about 83 and 94% of stations have increasing trends for TX90 and TN90, respectively. Compared with the results at the annual scale, the absolute trend magnitudes of winter temperature extremes are higher, and the proportions of stations with positive/negative trends are larger with the exception of TX90. Thus, the spatial and temporal patterns of winter temperature extremes are broadly similar to those on the annual basis, but the trends of temperature extremes in winter are generally higher, indicating pronounced climate warming in winter.

3.3. Correlation with the AO

The correlation between winter temperature indices and the AO in China during 1961–2003 are shown in Figure 2. National linear correlations and coefficients are listed in Figure 3. Strongest correlations occur in the northern China for TX10 and TN10 (some values lower than −0.5), and the correlations are slight in the southeastern part of the Tibetan Plateau for TX10 and TN10 (Figure 2). Taking China as a whole, the AO

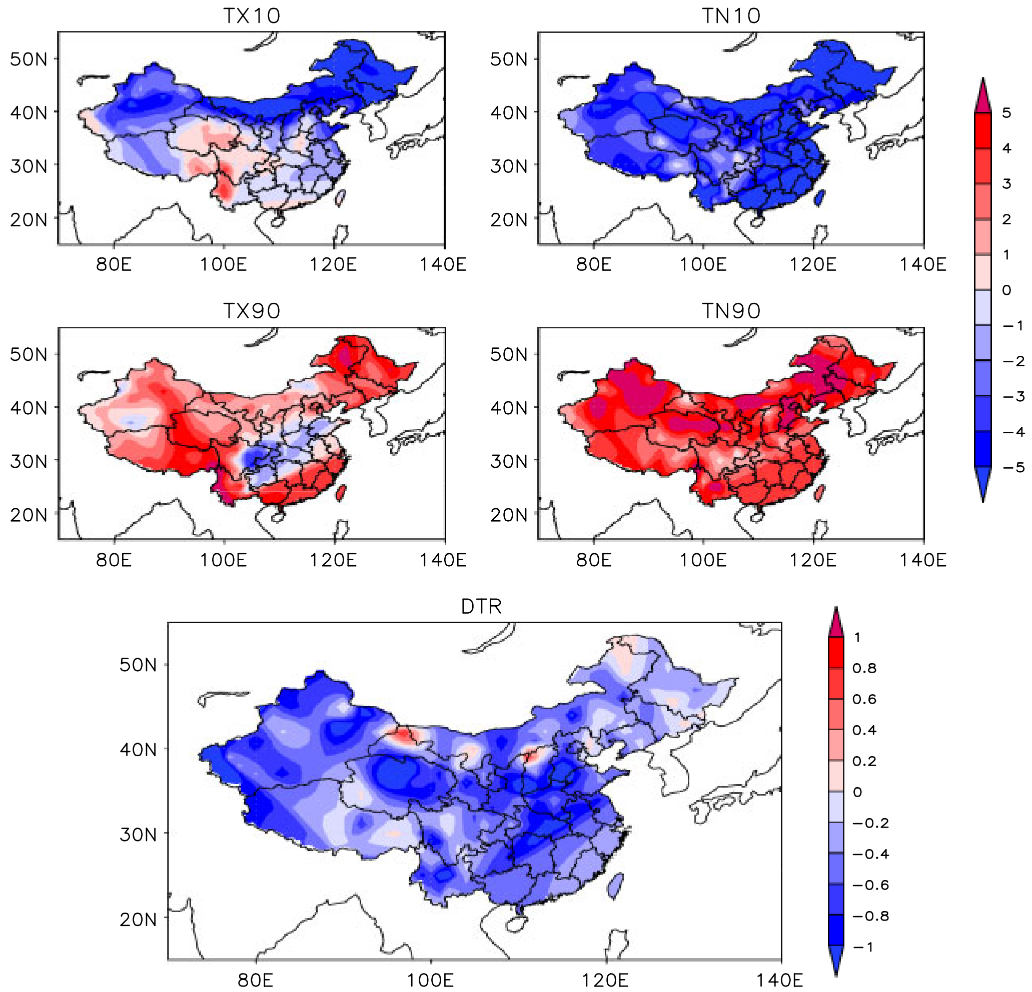

Figure 4. Differences of mean winter temperature extremes in positive and negative winter AO years during 1961–2003 are presented. The 23 positive (1963, 1966, 1971–1972, 1974–1975, 1982–1983, 1986–1994 and 1996–2001) and 19 negative (1961–1962, 1964–1965, 1967–1970, 1973, 1976–1981, 1984–1985, 1995 and 2002) winter AO years are selected whether the AO index is above or below the mean value. The unit is same as Table I. This figure is available in colour online at wileyonlinelibrary.com/journal/joc

index is significantly correlated with the winter cold temperature extremes (TX10 and TN10). It is negatively correlated with TX10 ($R = -0.42$, $p < 0.01$) and TN10 ($R = -0.65$, $p < 0.01$) during the studied period. For the winter warm temperature extremes, the AO index is positively correlated with TX90 and TN90, but only the correlations with TN90 ($R = 0.47$, $p < 0.01$) pass the significant level. In most cases, it is clear that the northern and northwestern China have larger correlation coefficients, and the southeastern China have lower values for the winter warm temperature extremes (TX90 and TN90, especially for TN90). Meanwhile, the AO index also has significantly negative correlation with DTR with the value of −0.51 (Figure 3), and correlation coefficients in most regions are more than −0.3. Thus winter temperature extremes are strongly connected with the AO index, especially in the northern China.

3.4. Atmospheric circulation composite analysis

In order to examine the influence of AO on climate extremes, the differences of mean winter temperature extremes in positive and negative winter AO years during 1961–2003 are presented (Figure 4). The 23 positive and 19 negative winter AO years are based on whether the AO index is above or below the mean value. The differences (positive minus negative AO years) show that the majority of stations have negative values for TX10, TN10 and DTR, and positive values for TX90 and TN90, while there has spatial variability (Figure 4). Thus winter temperature extremes are significantly different during winters with positive *versus* negative AO phases, which is consistent with that there are significant relationships between AO and temperature extremes (Figures 2 and 3).

To show the influence of atmosphere circulation on winter temperature extremes, Figure 5 shows the differences (positive minus negative AO years) of mean geopotential height and wind field (m s^{-1}) at 850 hPa during 1961–2003. The selected region covers the domain 10° – 70°N and 40° – 160°E. The largest negative differences in geopotential height are approximately 30 geopotential meter (gpm), with enhanced cyclonic circulation over the region (focused near 60°N and 60°E) (You *et al.*, 2011b). This generates an anomalous southwesterly flow in the Siberian region and northern China

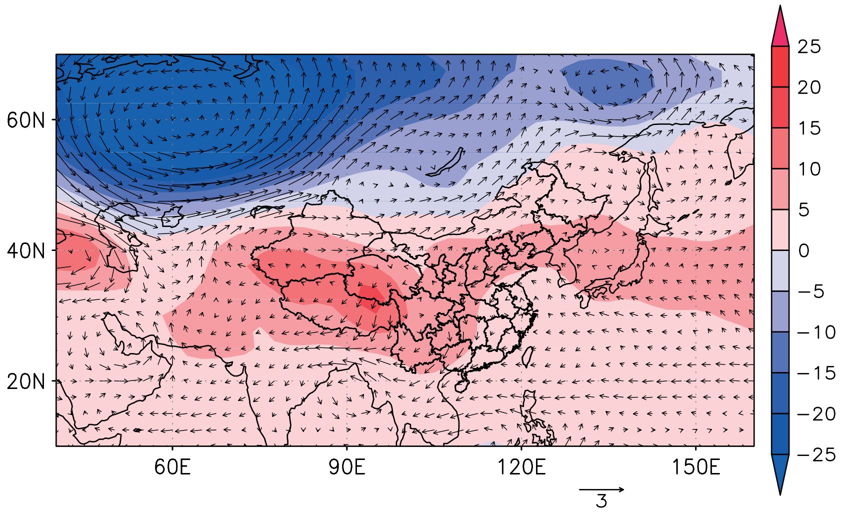

Figure 5. Differences (positive minus negative winter AO years) of mean geopotential height and wind field (m s^{-1}) at 850 hPa during 1961–2003. This figure is available in colour online at wileyonlinelibrary.com/journal/joc

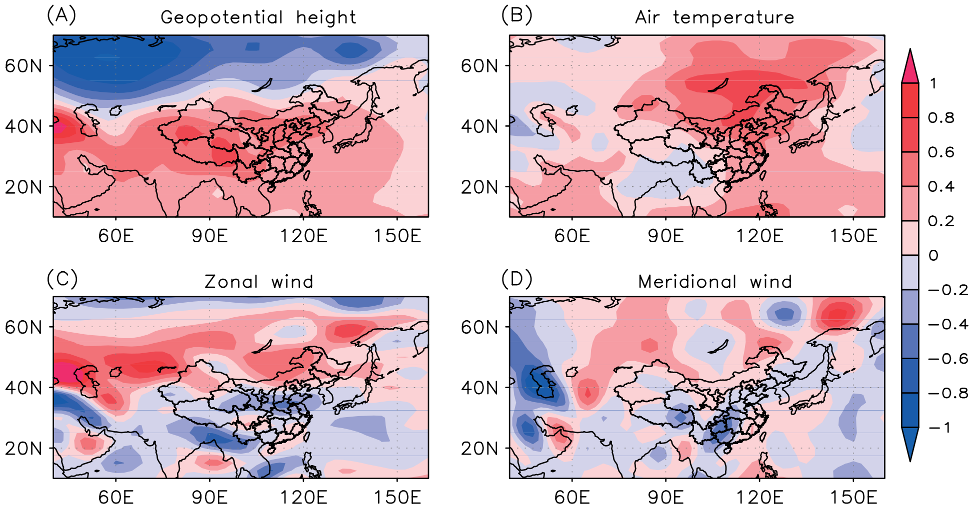

Figure 6. Spatial trends of geopotential height (A), air temperature (B), zonal wind (m s^{-1}) (C) and meridional wind (m s^{-1}) (D) at 850 hPa in winter during 1961–2003. The datasets come from NCEP/NCAR reanalysis. The units for geopotential height, air temperature, zonal wind and meridional wind are gpm/decade, 10× °C/decade, 10×m s^{-1}/decade and 10×m s^{-1}/decade, respectively. This figure is available in colour online at wileyonlinelibrary.com/journal/joc

which carries relative warm air into inland, reducing the intensity of Asian winter monsoon. These results suggest that decreasing trends for winter cold temperature extremes and increasing trends for winter warm temperature extremes are highly related to the circulation change (You *et al.*, 2011b).

To examine the atmosphere circulation, the spatial trends of geopotential height (A), air temperature (B), zonal wind (m s^{-1}) (C) and meridional wind (m s^{-1}) (D) at 850 hPa in winter during 1961–2003 are presented in Figure 6. The geopotential height at 850 hPa has decreasing trends at high latitude between 40° and 60°N and increasing trends at low latitude between 10° and 40°N (Figure 6(A)), suggesting that the asymmetrical changes between high latitude and mid latitude will begin to reduce the winter monsoon system, which is consistent with the asymmetrical global warming (IPCC, 2007). Although, the air temperature has increasing trends and more pronounced warming in the northeastern China (Figure 6(B)), the zonal and meridional wind increases significantly between 40° and 60°N (Figure 6(C) and (D)), revealing that the western and southern wind are increasing. The atmosphere conditions support the hypothesis that the increasing contrast between high and mid latitude will reduce the winter monsoon and influence the outbreak of winter temperature extremes.

The tropospheric temperature contrast between high and mid latitude is also of great importance to form the winter monsoon, supporting the atmospheric circulation for the change of temperature extremes. The tropospheric temperature (unit is °C) is defined as the average of air temperature vertically integrated between 200 and 1000 hPa based on the NCEP/NCAR reanalysis. The differences of composite tropospheric temperature

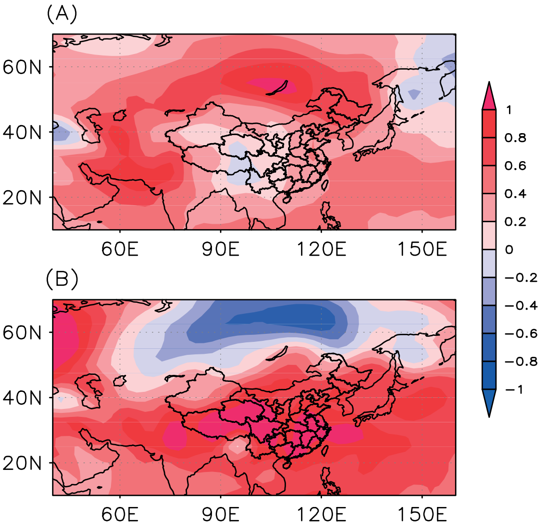

Figure 7. Differences of composite tropospheric temperature during the period of 1961–1982 and 1983–2003 [latter minus former, top plot (A)] and in strongly positive and strongly negative winter AO years during 1961–2003 [bottom plot (B)] are shown. Strongly positive (1982, 1988, 1989, 1991, 1994) and strongly negative (1962, 1964, 1968, 1976, 1978) winter AO years are those with index anomalies exceeding $\pm 1\sigma$. The study area is 40°W–160°E and 10° – 70°N. Tropospheric temperature (unit is °C) is defined the average of air temperature vertically integrated between 200 hPa and 1000 hPa based on the NCEP/NCAR reanalysis. This figure is available in colour online at wileyonlinelibrary.com/journal/joc

during the period of 1961–1982 and 1983–2003 (latter minus former) are show in Figure 7 (top plot). The tropospheric temperature has larger values at the higher latitude, especially near the 55°N and 100°E (1.2 °C). These atmospheric patterns will reduce the transport of energy through baroclinic waves, diminish the strength of troughs and ridges, and increase the occurrence of calm atmospheric conditions (Niu *et al*., 2010). Thus, the transport of cold air originating from high latitude around 70°N will become less powerful and influence the frequency of warm temperature extremes (Gong *et al*., 2001; Niu *et al*., 2010).

The AO may influence the warm temperature extremes in China through the contrast of atmosphere conditions. Differences of composite tropospheric temperature in strongly positive and strongly negative winter AO years during 1961–2003 are shown in Figure 7 (bottom plot). Strongly positive (1982, 1988, 1989, 1991 and 1994) and strongly negative (1962, 1964, 1968, 1976 and 1978) winter AO years are those with index anomalies exceeding $\pm 1\sigma$. During the positive AO years, the tropospheric temperature is positive in most southern China (near 1 °C), and there has negative anomaly in the north of China (almost 0.8 °C). Thus, the enlarging contrast of tropospheric temperature between high (around 60°N) and mid latitude (around 30°N) are helpful to bring more warm air flow from the ocean and prevent the cold air flow from the north, which will weaken the Asian winter monsoon and reduce the cold outbreaks. The patterns are similar to the correlation map of vertical-latitude from the 1000 to 10 hPa (Figure 8), which the AO has significant negative/positive correlations with atmospheric variables (geopotential height, temperature and wind) at high/mid latitude at both troposphere and stratosphere.

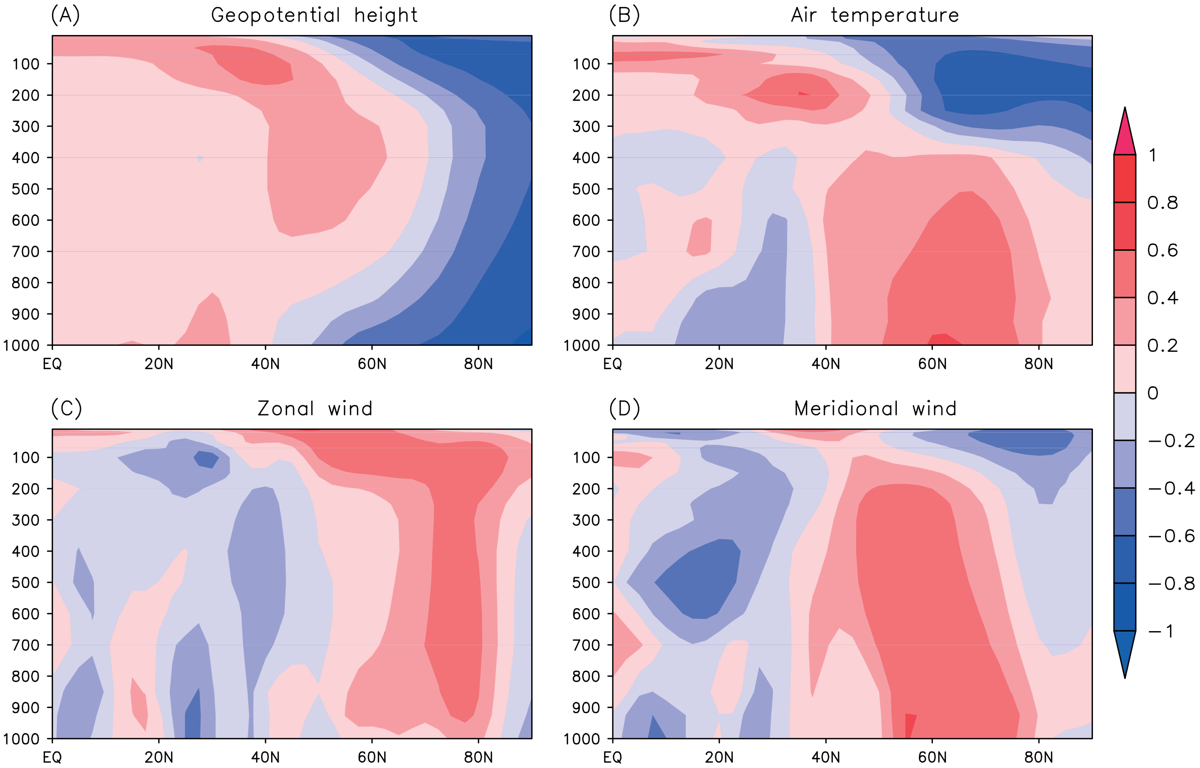

Figure 8. Vertical-latitude correlation coefficients between winter AO index and geopotential height (A), air temperature (B), zonal wind (m s^{-1}) (C) and meridional wind (m s^{-1}) (D) during 1961–2003. This figure is available in colour online at wileyonlinelibrary.com/journal/joc

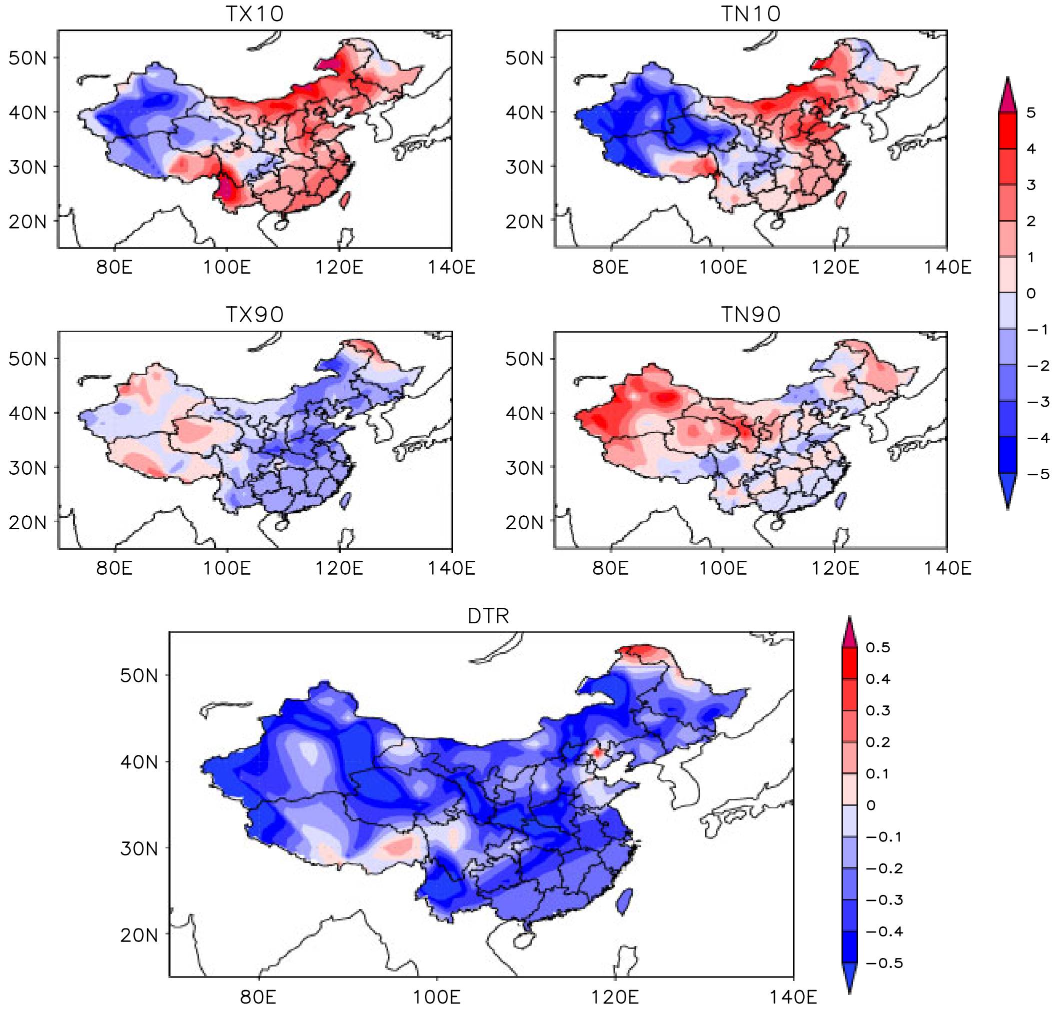

Figure 9. Differences of mean winter temperature extremes in solar maximum and minimum years are shown. The solar maximum years (1967–1971, 1979–1983, 1989–1992, and 1999–2002) and minimum years (1961–1966, 1972–1978, 1984–1988, and 1993–1998) are based on whether the solar fluxes are above or below the mean value. The unit is same as Table I. This figure is available in colour online at wileyonlinelibrary.com/journal/joc

3.5. Influenced by the solar activity

Previous studies have shown that the extension of the AO differs significantly for different phases of solar activity (Kodera, 2002; Gimeno *et al.*, 2003; Ogi *et al.*, 2003). In order to investigate the effect of solar activity on winter temperature, the studied period has been separated into two phases for maximum and minimum solar activity, depending on whether the solar fluxes are above or below the mean value (Kodera, 2002; Gimeno *et al.*, 2003; Ogi *et al.*, 2003). The 18 years (1967–1971, 1979–1983, 1989–1992 and 1999–2002) are classified as the solar maximum years, and the 24 years (1961–1966, 1972–1978, 1984–1988 and 1993–1998) as the solar minimum years. The classification is the same as the studies in Kodera (2002) and Ogi *et al*(2003).

During solar maximum years, about 70, 49, 17 and 47% of stations for TX10, TN10, TX90 and TN90, respectively, have larger values than that during solar minimum years. It is clear that the larger negative differences between solar maximum and minimum years for both TX10 and TN10 are shown in the western and northwestern China, while stations in the western and northwestern China show larger positive values for both TX90 and TN90 (Figure 9). In most regions, DTR is sensitive to the change of solar activity, and about 93% of stations have larger values during solar minimum years than that during solar maximum years, resulting the negative differences between them (Figure 9). This is probably because more solar activity will heat the surface and increase the winter surface temperature, thus influencing the winter temperature extremes. Moreover, solar activity can also influence the atmospheric circulations which are memorized in the snow-cover, ice and permafrost regions (Ogi *et al.*, 2003). The western China especially in the Tibetan Plateau is more sensitive to climate change due to the larger cryospheric area, and shows stronger signal. This probably suggests that solar activity can influence the winter temperature extremes to some extents and vary with the surface conditions, while the detailed mechanism needs to be investigated in future studies.

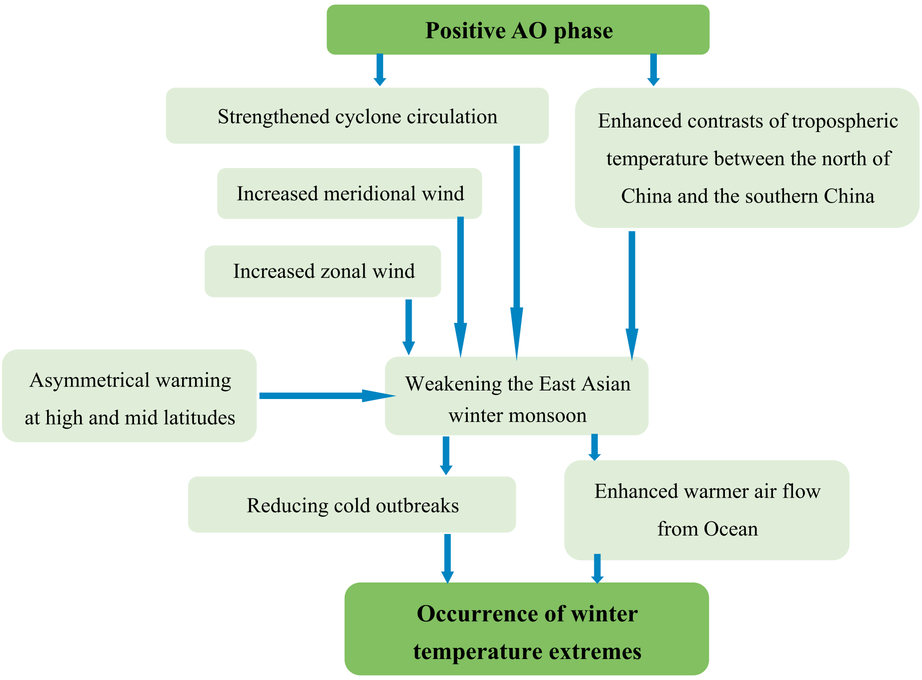

Figure 10. Possible causes of winter temperature extremes in China. This figure is available in colour online at wileyonlinelibrary.com/journal/joc

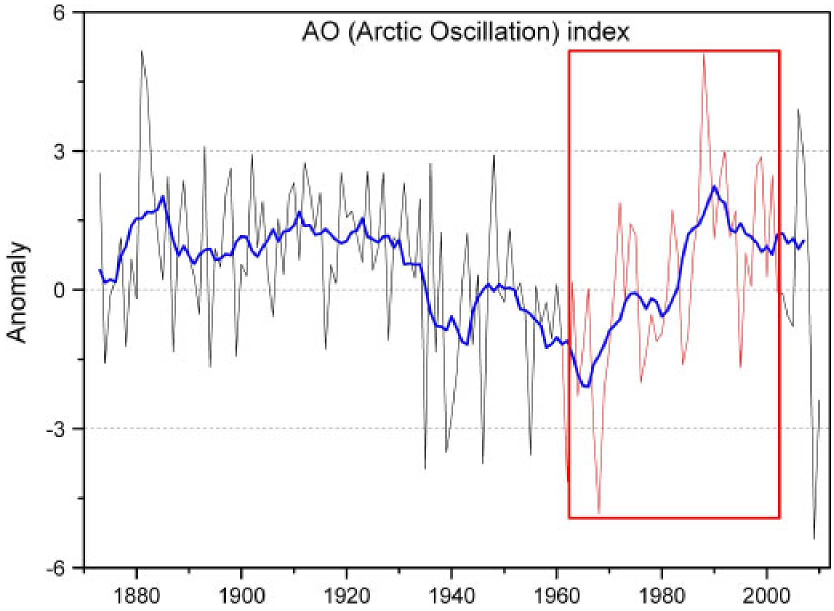

Figure 11. Time series of winter AO index during 1873–2010. The studied period of 1961–2003 is showed in the red rectangle. The smoother line is the 9 year smoothing average. The AO index is updated from Li and Wang (2003). This figure is available in colour online at wileyonlinelibrary.com/journal/joc

4. Discussion and conclusions

Winter temperature extremes have been shown to be warming faster than annual mean warm extremes (Aguilar *et al.*, 2009) and are therefore particularly sensitive to future change. We have examined the spatial and temporal distributions of trends for four winter temperature extreme indices, using 303 stations in China over the period 1961–2003. For the majority of stations, significant decreases in cold days/nights (TX10/TN10) are observed with mean rates of −1.33 and −2.98 d/decade, respectively, while significant increases in warm d/nights (TX90/TN90) are also observed with mean rates of 0.92 and 2.35 d/decade, respectively. Such changes are consistent with previous studies in other parts of the world (Peterson *et al.*, 2002, 2008; Aguilar *et al.*, 2005, 2009; Alexander *et al.*, 2006), and show that changes in winter temperature extremes reflect the consistent winter warming in China (You *et al.*, 2008a,b, 2011a). The asymmetric changes in minimum and maximum temperature result in the declining DTR with rate of −0.25 °C/decade. In most cases, stations in the north of China have the largest trend magnitudes, again consistent with rapid warming in the region (Wang and Gong, 2000; Ding *et al*., 2007).

Changes in winter temperature extremes are consistent with the report of You *et al*. (2011a) at the annual scale, but the trend magnitudes are higher than those, suggesting the winter is more sensitive to the extremes. The causes about the temperature extremes change have been studied, but require further study. Besides gas greenhouse gas emissions, You *et al*. (2011a) considered that temperature extremes change is probably associated with rapid urbanization, increased industrial aerosols and non-climate factors such as population, economic activity and local energy usage. The influences become particularly significant in China because of its rapid urbanization and economic activity (Qian and Lin, 2004).

The AO influences surface air temperature not only over the bulk of the Eurasian continent (Hurrell, 1995; Thompson and Wallace, 1998) but also in northern China (Gong and Wang, 2003). The winter AO index is significantly negatively correlated with TX10/TN10 and DTR, and positively correlated with TX10/TN10, indicating that the AO influences winter cold/warm extreme temperature. During the strongly positive AO index years, enhanced cyclonic circulation over the Urals (focused near 50°N and 60 °E) brings more warm airflow into northern China, decreasing the strength of the East Asian winter monsoon and limiting its southward extension (Figure 10). This is consistent with previous research that shows that atmospheric circulation changes have contributed to the changes in climate extremes in China (You *et al*., 2011a). Other work has also suggested that an increase in strong positive AO phases could lead to a decreasing East Asian winter monsoon (Gong *et al*., 2001; Wu and Wang, 2002). Composites of atmospheric circulation shown in this study also support the relationship between AO variability and the strength of winter cold outbreaks in the northern China.

It is notable that the winter AO index has shifted several phases during 1873–2010 (Figure 11), derived from Li and Wang (2003). Winter AO index increases since the 1960s and has the downward trend since 1990 (Hurrell and Deser, 2010), confirmed by the weakening East Asian winter monsoon (Niu *et al*., 2010). The asymmetrical change in geopotential height, zonal and meridional wind may reflect a weakening of the East Asian winter monsoon. At the same time, more warming at high latitudes also reduce the thermal contrast, contributing to the weakening the East Asian winter monsoon (Figure 10). This will reduce the invasion of dry and cold air from the northern regions, creating a favourable background for temperature extremes. The limitation of this study is that the study period stops at the end of 2003, and the recent winters have strongly negative AO values in 2009 and 2010, especially in 2010 (Figure 11). In 2010, China has experienced the coldest winter since 1987, with the annual mean temperature of −4.7 °C (Ren *et al*., 2011). This supports the hypothesis that the AO can modulate the winter temperature extremes by the atmosphere conditions. Meanwhile, the AO can be modulated by the 11 year solar cycles (Kodera, 2002). In this study, winter temperature extremes have stronger values during solar maximum years in most cases. But the mechanical linkage between solar activities, the AO and temperature extremes need to be investigated in future studies. Our results indicate also that further investigation of the linkage between the AO and climate extremes in China is worthwhile.

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