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Links between the late wintertime North Atlantic Oscillation and springtime vegetation growth over Eurasia

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Abstract In the present study, the linkages between the late wintertime (January–February–March; JFM) North Atlantic Oscillation (NAO) and springtime (April–May–June; AMJ) vegetation growth over Eurasia is investigated. Here, the proxy of vegetation growth is represented by normalized difference vegetation index (NDVI) gridded data, obtained from the advanced very high resolution radiometer. Over the period 1982–2006, the NAO (JFM) correlated well with the NDVI (AMJ) over Eurasia, wherein a positive NAO tended to increase the NDVI (AMJ) over Eurasia and vice versa. The results show that a positive phase of the late wintertime NAO leads to an increase in surface air temperature, soil temperature and rainfall in most parts of Eurasia in winter. These changes tend to produce weaker and thinner snow cover in spring compared to that that forms in a negative NAO phase. Corresponding to this, the albedo decreases and the surface air temperature increases over Eurasia in spring, which contributes an earlier snowmelt. Subsequently, the land surface over Eurasia becomes warmer and wetter earlier, as the snow melts. These conditions can then facilitate higher than average vegetation growth over Eurasia, in comparison to the conditions that occur in a negative NAO phase.

Keywords NDVI · NAO · Non-simultaneous · Snow · Eurasia

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1 Introduction

Vegetation plays a key role in climate change due to its influence on potential feedback mechanisms, such as albedo, evapotranspiration, roughness and surface processes. In turn, changes in climate strongly influence vegetation greenness, through variations in air or soil temperature, precipitation, radiation, atmospheric circulations and other meteorological variables. Recently, correlations between vegetation and changes in climate have attracted wide attention by scientists (Betts et al. 1997; Dai and Zeng 1997; Hoffmann and Jackson 2000; Jiang et al. 2011; Rodriguez-Iturbe et al. 1999; Yuan et al. 2011; Zeng et al. 2000). Gao et al. (2007) pointed out that the land use change influenced the circulation and surface energy budget by altering the surface roughness, leaf-area-index (LAI) and albedo.

Global mean surface air temperature has increased since the late nineteenth century; each of the past three decades has been warmer than all the previous decades, and the 2001–2010 decade was the warmest since instrumental records began. The global combined land and ocean temperature data shows an increase of about 0.89 °C over the period 1901–2012 (IPCC 2013). As a result of global warming during this period, the active growing season has become longer, with a much earlier spring and later autumn than before, leading to an increase of photosynthetic activity by vegetation (Bogaert et al. 2002; Zhou et al. 2001).

The changes described have been detected in satellite data (Tucker et al. 2001; Zhou et al. 2001). Considering the synoptic coverage and repeated temporal sampling that satellites offer, the potential to use remotely sensed data for monitoring vegetation dynamics at regional to global scales is huge (Myneni et al. 1997). To quantify the spatial and temporal variations in vegetation growth and activity, the normalized difference vegetation index (NDVI) (Tucker 1979) is calculated from near-infrared and red visible reflectance in satellite images; the NDVI is used as an indicator of vegetation greenness (Myneni et al. 1997) and has been widely developed to monitor vegetation phenology (Cleland et al. 2007; Jia et al. 2006, 2009).

Vegetation activity is strongly influenced by the local weather variables, so researchers have identified the local meteorological variables associated with the NDVI, such as air temperature, precipitation and evapotranspiration (Di et al. 1994; Kawabata et al. 2001; Nemani and Running 1989; Wang et al. 2003). However, large-scale climate systems also play essential roles in how the NDVI changes in response to global change (Gong and Shi 2003). For example, some studies have found possible connections between the regional NDVI and the Southern Oscillation (SO) over lower latitudes (Kogan 2000; Myneni et al. 1996). Over the mid- to high-latitudes other climate systems, such as the North Atlantic Oscillation (NAO), Arctic Oscillation (AO) or the Pacific/North American (PNA) pattern, might exert greater influence on the NDVI. Therefore, while traditionally research has focused on the effects of local weather variables on the regional NDVI, the linkage between it and large-scale climate systems has become a hot topic (Anyamba and Eastman 1996; Cho et al. 2014; Gong and Shi 2003).

The NAO is a dominant winter climate pattern over the Atlantic Ocean and Europe; it refers to a large-scale seesaw of atmospheric mass between the subtropical high (centered on the Azores) and the polar low (centered on Iceland) (Hurrell et al. 2003). The NAO is known to influence the climate variability over the wintertime Northern Hemisphere (Rodwell et al. 1999; Watanabe and Nitta 1999) and even spring-summertime atmospheric circulations (Hahn and Shukla 1976; Ogi et al. 2003; Qian and Saunders 2003; Sun and Wang 2012; Sun et al. 2008; Tian and Fan 2012). For example, Ogi et al. (2003) showed that when the wintertime NAO is in a positive phase, the summertime surface air temperatures over circumpolar regions in northern Eurasia and subarctic North America become warmer and the geopotential height is higher than when NAO is in a negative phase. They also suggested that the wintertime NAO is memorized in the snow, sea ice and ocean surface in the circumpolar regions and that these anomalies subsequently influence the summertime atmospheric circulation in the extratropics. Similar conclusions to these can be seen in other studies (Bamzai and Shukla 1999; Qian and Saunders 2003). The seasonally lagged signal of the winter NAO in the north Pacific is also investigated by Zhao and Moore (2006). They showed the spring sea-level pressures and surface temperatures in the region are positively correlated with the characteristics of the NAO during the preceding winter. They identified two distinct mechanisms responsible for this lagged signal: sea-surface temperature anomalies in the North Pacific and Eurasian snow anomalies. Zhou (2013) studied the relationship between the winter (December–February) NAO and the precipitation over southern China in the following spring (March–May). Results showed that the wave train propagated the NAO signal eastward to East Asia and affected local upper-tropospheric atmospheric circulation. Zhou and Cui (2014) also investigated the relationship between the NAO and the tropical cyclone frequency over the western North Pacific in summer.

Studies on the relationship between the winter NAO and the regional NDVI have become of interest to scientists. Non-simultaneous correlations between the NAO and variations in vegetation growth over the northern hemisphere and Asia have been shown (Gong and Shi 2003; Wang and You 2004; Zhou et al. 2013). Gong and Shi (2003) used the multivariate regression analysis technique to estimate quantitatively the connection between nine climate indices and NDVI values. Wang and You (2004) focused on the impact of NAO on the productivity of vegetation in Asia at various time lags without explaining what the underlying mechanism for the non-simultaneous correlations. Zhou et al. (2013) investigated the relationship between springtime NAO and the year-to-year increment of summer maize and rice yield in Northeast China. They revealed the key factor for this non-simultaneous relationship might be sea surface temperature (SST) anomalies in the North Atlantic induced by springtime NAO. Additionally, Vicente-Serrano and Heredia-Laclaustra (2004) demonstrated significant spatial differences in the NDVI developed between 1982 and 2000 on the Iberian Peninsula; these were characterized by a positive trend of the NDVI in the north of the Iberian Peninsula, and both stable and negative trends of the NDVI in the southern areas. This spatial pattern of the NDVI was identified to be significantly related to the NAO, because the latter could impact the former by greatly influencing the precipitation distribution even the atmospheric pattern on the Iberian Peninsula (Vicente-Serrano and Heredia-Laclaustra 2004). Li et al. (2012) showed that the regions where the NAO and NDVI correlate well are mainly concentrated in the mid- and high-latitude areas of the northern hemisphere (around the 60°N belt) and the African zone (around the 15°N belt), as well as the vast regions of the southern hemisphere around the 10°S–30°S belt.

There are a few studies that have focused on the statistical relationships between the NAO and the NDVI (Gouveia et al. 2008; Maignan et al. 2008); however, this research has mainly concentrated on Europe, North America and North Africa and there has been little research focused on Eurasia (Martínez-Jauregui et al. 2009; Vicente-Serrano and Heredia-Laclaustra 2004; Wang 2003; Wang et al. 2003). Therefore, in the present study, we primarily focus on the linkage between variability in the springtime NDVI over Eurasia (30°E–150°E, 45°N–70°N) and the winter NAO, to investigate how the late wintertime NAO affects the following springtime NDVI activity in the region. Studies into the physical mechanisms that lead to the remote linkage between the NAO and the NDVI remain limited, so the present study also aims to provide probable physical mechanisms for the non-simultaneous relationship.

This paper is divided into five sections. The data and methods employed in this work are first described in Sect. 2. The analysis of spatio-temporal variability of the vegetation over Eurasia and the lagged relationship between the late wintertime NAO and springtime NDVI are presented in Sect. 3. The physical mechanisms responsible for the non-simultaneous NAO–NDVI relationship are discussed in Sect. 4 and a summary and discussion are provided in Sect. 5.

2 Data and methods

2.1 Vegetation data

Temporal variations in vegetation photosynthetic activity were investigated using the monthly NDVI dataset (Tucker 1979). The dataset was compiled at an 8 km spatial resolution from the AVHRR on board the National Oceanic and Atmospheric Administration (NOAA) series of meteorological satellites, by the Global Inventory Monitoring and Modeling Studies (GIMMS) research group (Tucker et al. 2005). The data were originally processed as 15 day composites for the period from 1982 to 2006 ($n = 25$) to further minimize the effects of clouds on the vegetation signal (Dai et al. 2010). The GIMMS–NDVI dataset was calibrated and corrected for view geometry, volcanic aerosols and other effects that are not related to vegetation change.

The NDVI is calculated from these individual measurements as follows:

$$NDVI = \frac{(NIR - VIS)}{(NIR + VIS)},$$

where VIS and NIR stand for the spectral reflectance measurements acquired in the visible (red) and near-infrared regions, respectively (see http://earthobservatory.nasa.gov/Features/MeasuringVegetation/measuring_vegetation_2.php). The NDVI itself varies between −1.0 and +1.0, wherein an area containing a dense vegetation canopy is denoted by positive values and clouds and snow fields result in negative values. Therefore, for the present analysis we select pixels with an NDVI value of more than 0.1, to represent density of vegetation, and those pixels with a mean NDVI <0.1 are excluded (Li et al. 2012). In this way, we improve the credibility of the NDVI data and also make the analysis of the correlation between the vegetation and atmospheric circulation more clear.

It is noted that the NDVI in mid- to high-latitude continents over Eurasia has a sharp increase from April to June, which is consistent with previous results that show the NDVI exhibits greater trends in spring than in other growing seasons (Gong and Shi 2003; Suzuki et al. 2000; Zhou et al. 2001). Therefore, we defined springtime as April–June, and late wintertime refers to January–March. Our target region of NDVI is limited to the mid- to high-latitude continents over Eurasia (30°E–150°E, 45°N–70°N).

2.2 Climate data

The National Centers for Environmental Prediction and the National Center for Atmospheric Research (NCEP/NCAR) reanalysis dataset (Kalnay et al. 1996), and the European Centre for Medium Range Weather Forecasts Interim reanalysis dataset (Dee et al. 2011), for the period from 1982 to 2006 (the same time span of the NDVI data) are employed in this study; the data utilized include monthly mean sea level pressure, u–v wind vector, 2 m air temperature, soil temperature, sensible heat flux and snow albedo. The Global Precipitation Climatology Project monthly precipitation dataset from the Climate Prediction Center (CPC) at a 2.5° × 2.5° resolution is also used. The snow cover data are extracted from the NOAA weekly snow cover extent (SCE) dataset, maintained at Rutgers University (http://climate.rutgers.edu/snowcover/). The satellite-based data provide weekly SCE data for the land masses of Eurasia, North America and the northern hemisphere as a whole (Déry and Brown 2007). The monthly NAO index was obtained from the public NOAA-CPC database (see http://www.cpc.ncep.noaa.gov/products/precip/CWlink/pna/nao.shtml). The NAO consists of a north–south dipole of anomalies, with one center located over Greenland and the other center, of opposite sign, spanning the central latitudes of the North Atlantic, between 35°N and 40°N. The NAO index is defined as the SLP difference between Stykkisholmur, Iceland and Ponta Delgada, Azores by Hurrell (1995) from 1982 to 2006. In the present study, the late wintertime NAO index is defined as the seasonal mean value for January, February, and March (JFM).

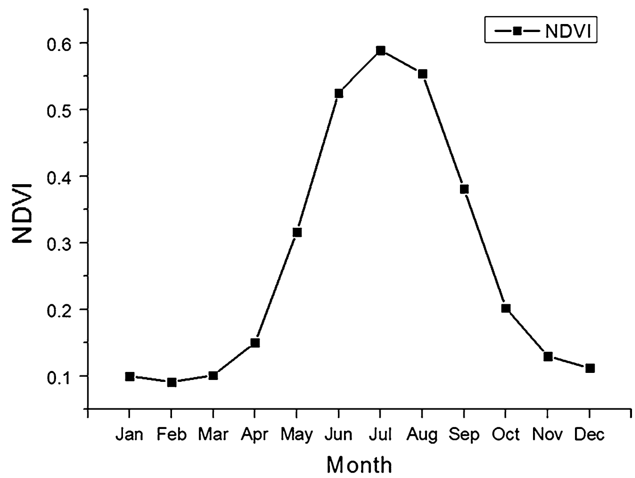

Fig. 1 Monthly area-average NDVI values over Eurasia (45°N–70°N, 30°E–150°E) between 1982 and 2006

Table 1 The autocorrelations of NDVI in 3 months (April–June) between 1982 and 2006

	April	May	June
April	1.00	0.55**	0.35*
May	0.55**	1.00	0.84**
June	0.35*	0.84**	1.00

* Significant correlation at 90 % level, and ** 99 % level, estimated by a local Student's *t* test

2.3 Methods

To analyze the spatio-temporal variability of the vegetation over Eurasia, an empirical orthogonal function (EOF) analysis of the NDVI is conducted. EOF is a good tool to extract a compact or simplified, but optimal, representation of spatio-temporal data. The statistical significance is determined by Student's *t* tests under the assumption that the sample data are independent. All analyses are carried out using de-trended data.

In order to illustrate the effect of the late wintertime NAO on springtime vegetation activity, atmospheric circulation and climate factors that are related to the NDVI and NAO, are investigated using correlation and composite difference analysis methods. Based on the criterion that the normalized EUNDVI (AMJ) is larger than 0.5 or less than −0.5 SD, the high-value EUNDVI years (1988, 1990, 1991, 1997, 2000, 2002) and low-value NDVI years (1983, 1987, 1998, 2003, 2004, 2006) are tagged. Based on the criterion that the NAO index is larger than 0.5 or less than −0.5 SD of NAO, the positive-phase NAO years (1989, 1990, 1992, 1993, 1995, 1997, 2000, 2002) and negative-phase NAO years (1985, 1987, 1996, 2001, 2004, 2005, 2006) are tagged.

3 Linkage between springtime NDVI over Eurasia and late wintertime NAO

3.1 Spatio-temporal evolution of the NDVI anomaly patterns

Vegetation growth activities can be divided into a growing season (April–October) and non-growing season (November–March), with the minimum NDVI value occurring in February and maximum in July (Fig. 1). The largest month-to-month increment of the NDVI occurs in April–May–June (AMJ); over this period there is a dramatic increase in the NDVI of 0.43 (between March and June), which accounts for 87.8 % of the total increase seen in the NDVI (0.49; between February and July). This indicates that the vegetation activity during AMJ is most vigorous, responding to the seasonal change from winter to spring. We also calculated the autocorrelations of NDVI in 3 months (April–June) (Table 1). The correlation between April and June NDVI is at the 0.1 significance level, others are all at the 0.01 significance level, as estimated from a standard Student's *t* test. Hence, AMJ is defined as the springtime period of the NDVI in this study, which is consistent with previous studies (Cho et al. 2014).

The first EOF pattern obtained for the period 1982–2006 (Fig. 2), accounts for more than 25 % of the total variance in the springtime NDVI and explains the major distribution of vegetation over Eurasia. In the first EOF pattern, positive values are present over most parts of Eurasia, indicating that the change in the springtime NDVI is almost the same across Eurasia. There are significant variations in the springtime NDVI within the area north of 50°N and between 80°E and 135°E; this implies that the vegetation varies more significantly in this high latitude region.

The time-coefficients of the first EOF pattern, which represent the temporal characteristics in vegetation phenology and depict the various vegetation growing periods, show both an interannual and interdecadal variability. Negative NDVI values occur during 1983–1987 and mostly positive values occur between 1988 and 2002. The normalized NDVI values are less than −1.0 SD in 1983, 1987, 2004, 2006 and greater than +1.0 SD in 1990 and 1997, respectively.

Assuming that the variability in NDVI is constant, the area-average NDVI (AMJ) index over Eurasia was calculated (30°E–150°E, 45°N–70°N). The correlation coefficients between the time series of the first EOF pattern of the NDVI (AMJ) and the area-average NDVI (AMJ) over Eurasia are 0.99 (raw data) and 0.98 (with the linear trend removed); both correlations are significant (α 0.01). Therefore, the time-coefficient of the first EOF pattern of NDVI (AMJ) over Eurasia (EUNDVI) is used to present

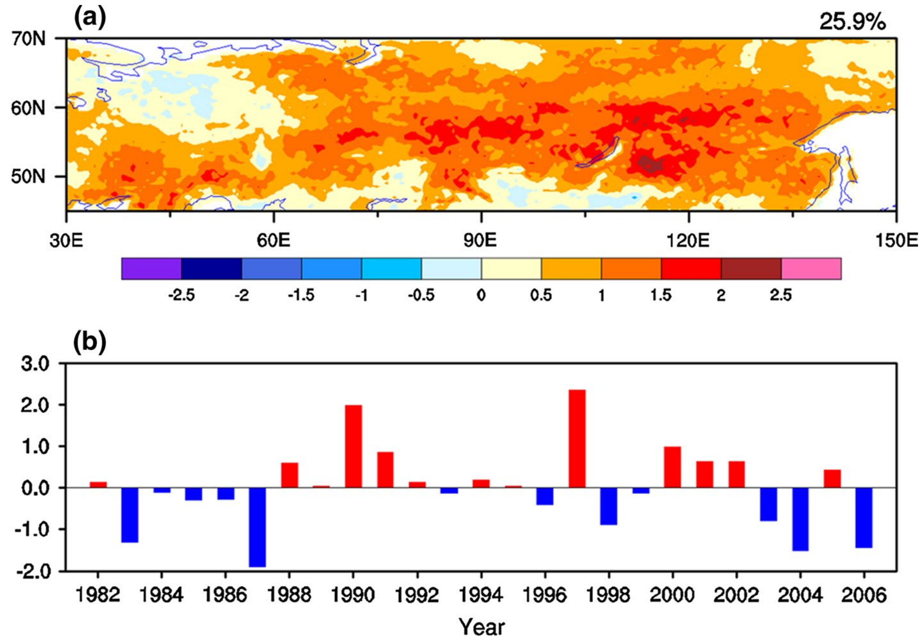

Fig. 2 The first EOF pattern obtained from the springtime NDVI (AMJ) over Eurasia (45°N–70°N, 30°E–150°E). **a** The spatial distribution and **b** its time-coefficient

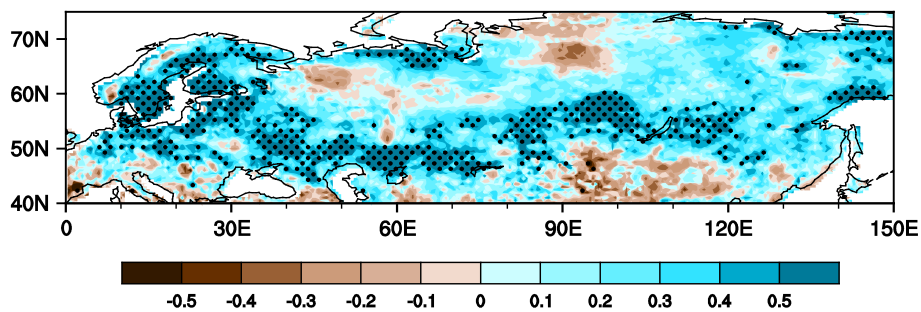

Fig. 3 The correlation coefficient between the spring NDVI (AMJ) and the previous winter NAO (JFM), over Eurasia. The *dotted regions* are the statistical confidence levels at 95 %, estimated by a local Student's *t* test

the primary vegetation phenology during springtime over Eurasia.

3.2 Correlation between the NAO (JFM) and the NDVI (AMJ)

The correlation coefficients between the late wintertime NAO (JFM) and the NDVI (AMJ) are positive across most parts of Eurasia (Fig. 3). That is, when the late wintertime NAO is in a positive phase, the vegetation activity over most parts of Eurasia in the coming spring tends to be stronger. The statistically significant correlation coefficients dominate the 45°N–55°N belt, extending from western Europe to Siberia and reaching the Lake Baikal region.

The time series of the detrended NAO (JFM) and the EUNDVI (AMJ), during 1982–2006 (Fig. 4), show that the variation of the NDVI (AMJ) over Eurasia is generally consistent with the variation of the late wintertime NAO (JFM), except in some years, including 1983, 1989, 1993, 2001 and 2005. The correlation coefficients between the two series are 0.50 (containing the linear trend) and 0.56

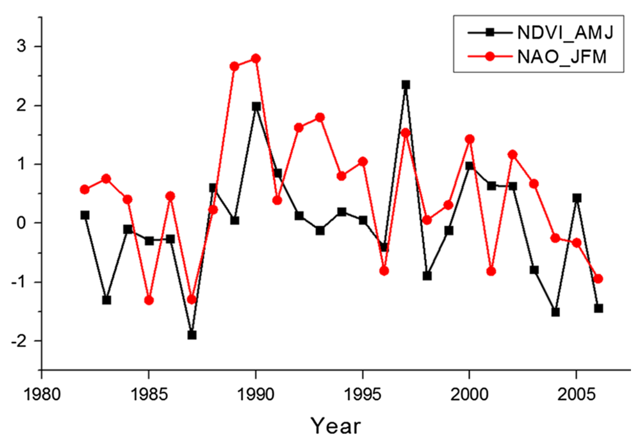

Fig. 4 Changes in the EUNDVI (AMJ) (*black squares*) and the NAO (JFM) index (*red dots*) over time

(de-trended); both are significant (α 0.01), as estimated from a standard Student's *t* test. This demonstrates that the late wintertime NAO is an important factor that influences the spring vegetation activity in Eurasia, as it can be used to

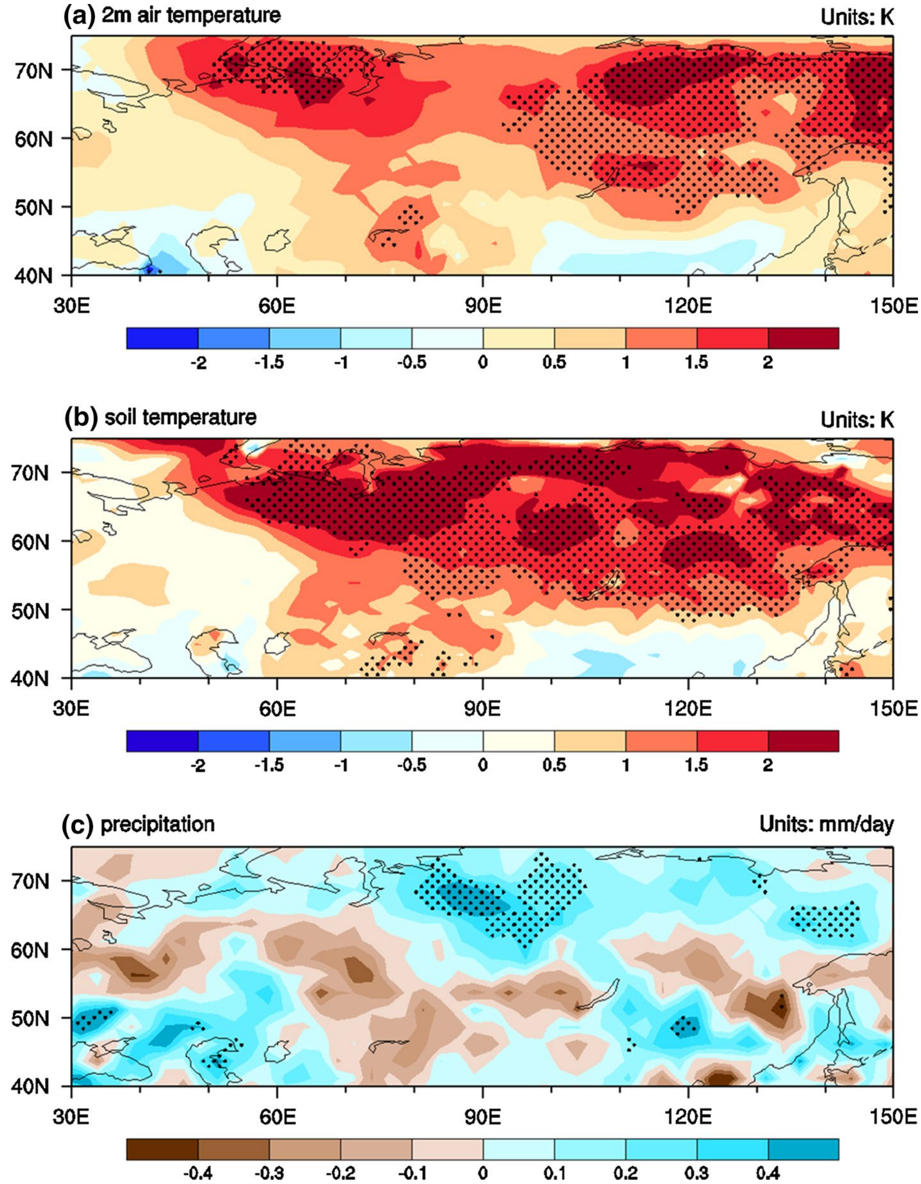

Fig. 5 Spring (AMJ) differences in temperature and precipitation between the high- and low-value NDVI (AMJ) years. **a** Air temperature at 2 m (K), **b** soil temperature (K), **c** precipitation (mm/day). *Dotted areas* show significant changes at the 95 % level, estimated by a local Student's *t* test

explain 31.4 % of the variance that occurs in the greenness index in the subsequent spring, over the mid- and high-latitude region of Eurasia (Figs. 3, 4).

4 The responsible physical mechanisms for the non-simultaneous relationship

4.1 The climate factors related to NDVI (AMJ)

Several relevant springtime (AMJ) climate factors are identified through the investigation of climatic differences between the high-value and low-value EUNDVI (AMJ) years including 2 m air temperature, soil temperature and precipitation (Fig. 5). A positive EUNDVI (AMJ) is closely related to a significant increase in both 2 m air temperature (Fig. 5a) and soil-temperature (Fig. 5b). A positive EUNDVI (AMJ) also corresponds to higher than normal rainfall in two areas north of 60°N, and in an area between 30°E–60°E and 40°N–50°N, along with lower than average rainfall south of 60°N (Fig. 5c).

The data show that the NDVI is, statistically, more closely linked with surface temperature than precipitation, in agreement with previous findings (Ichii et al. 2002; Kawabata et al. 2001; Los et al. 2001; Schultz and Halpert 1993). Schultz and Halpert (1993) found that the onset of precipitation generally appears to act as the stimulus for vegetation in regions where the magnitude of the annual temperature cycle is small, and vice versa. Los et al. (2001) observed a very strong connection between anomalies in the NDVI and land surface air temperature, existing in 3.4 years (NAO) signals in Europe, but showed weaker associations with precipitation, indicating that moisture generally does not limit vegetation growth in the region. Some evidence also indicated that the vegetation anomaly in the northern mid- and high-latitudes is predominantly a response to warmer surface temperatures (Myneni et al. 1997; Tucker et al. 2001; Zhang et al. 2004). It is clear that temperature is a dominant climate factor that influences the NDVI variability over Eurasia, with increased temperatures corresponding to a higher NDVI.

The next question then, is what is the relationship between the late wintertime NAO (JFM) and the climate factors that influence the spring NDVI in Eurasia? The composite differences in springtime (AMJ) climate factors that influence the NDVI and the positive- and negative-phase NAO (JFM) years indicate that a positive NAO tends to increase the springtime Eurasian air temperature and soil temperature, and also increases rainfall at both high latitudes and in southern Europe (Fig. 8). Clearly, the climate patterns in Fig. 8 resemble those seen in Fig. 5, showing the positive correlation coefficient between the NAO (JFM) and the NDVI (AMJ), in terms of the climate factors.

4.2 The physical mechanisms responsible for the link between the NAO (JFM) and the NDVI (AMJ)

The NAO bipolar mode of circulation consists of the exchange of air between the subtropical Azores high and the polar Icelandic low. The result in the previous section shows that the late wintertime NAO has a significant impact on the spring NDVI. Since the atmosphere itself does not have long memory (it is <1 month) and the winter NAO does not have a significant auto-correlation after March (Ogi et al. 2003), something else with a longer memory should link the winter to spring (even summer), pattern that occurs.

Firstly, an analysis of the variations in the previous atmospheric circulation field, associated with the wintertime NAO (JFM). Since the anomalies, relative to the positive- and negative-phase NAO years, tend to have an opposite distribution, the composite difference is discussed hereafter. The composite differences of 850, 500 and 200 hPa horizontal winds (JFM) between the positive- and negative-phase NAO (JFM) years show various anomalies (Fig. 6a–c). During the positive phase of the winter NAO, there are barotropic circulation anomalies over the North Atlantic and into Eurasia, with cyclonic circulation anomalies over the northern part and anticyclonic circulation anomalies over the southern part. The form of an anomalous high over Lake Baikal (85°E–150°E, 35°N–60°N) is closely related to Rossby wave activity. Arrows superimposed in Fig. 6d are the wave activity flux at the 200 hPa level, formulated by Takaya and Nakamura (Takaya and Nakamura 1997, 2001). The arrows clearly show that the Rossby wave propagates south-eastward from Novaya Zemlya to the regions around Lake Baikal (85°E–150°E, 35°N–60°N), which leads to the anomalous high over Lake Baikal. This anomalous high increases solar radiation and contributes to warming over Lake Baikal.

Corresponding to the barotropic circulation anomalies (Fig. 6a–c), anomalous westerlies prevail over Eurasia, around 40°N–60°N, bringing warm moist air from north of the Atlantic to most of Eurasia, which leads to warming over Eurasia (Fig. 7a, b) (Qian and Saunders 2003; Rodwell et al. 1999). Precipitation over the northern part of Eurasia (50°N–70°N) increases (Fig. 7c), due to the warm moist air being transported by the prevailing westerlies and increased vertical motion induced by the cyclonic circulation anomalies (Fig. 6a–c). In contrast, the precipitation decreases over the southern part of Eurasia (40°N–50°N; Fig. 7c) due to the anticyclonic circulation anomalies that dominate there (Fig. 6a–c). These changes, especially the warming of the 2 m air temperature (Fig. 7a) and soil temperature (Fig. 7b), could produce the negative anomalies of snow cover (Fig. 9a), which represent weaker and thinner snow cover. As a result of the decreasing snow over Eurasia, the snow albedo accordingly decreases (Fig. 9b) and the land surface would store more shortwave radiation and temperature.

After a positive winter NAO, in the following spring (AMJ), as the thin snow melts fast, the land releases more heat (accumulated from the previous winter). Therefore, the soil temperature (AMJ) increases (Fig. 8a), along with an increase in the sensible heat flux (AMJ) (Fig. 8b) over most parts of Eurasia. Accordingly, the air temperature (AMJ) (Fig. 8c) increases and precipitation (AMJ) also increases over most parts of Eurasia (Fig. 8d); the latter occurs dramatically so in the areas encompassed by 50°E–60°E and 40°N–60°N, and 80°E–130°E and 65°N–75°N. However, there is also a negative anomaly rainfall belt at 60°N, extending from west to east across Eurasia. The most significant negative anomaly center of precipitation lies to the west of Lake Baikal, in the Sayan Mountains, which suggests a strong influence of the complex topography there. Ultimately, these conditions facilitate the earlier arrival of springtime; warm and moist surfaces over Eurasia in spring after a positive NAO phase (JFM) favor higher than average vegetation growth in comparison to conditions that follow a negative NAO phase.

According to the analysis above, we hypothesized that snow cover might be responsible for the linkage between the winter NAO and the spring NDVI. Between high- and low-value NDVI (AMJ) years, snow cover has significant negative anomalies over the mid- and high-latitudes of Eurasia (around 50°N–70°N) in winter (JFM) and spring (AMJ), respectively (Fig. 9c, d; where the area is blank, denotes the snow cover is 100 % and is the same as in

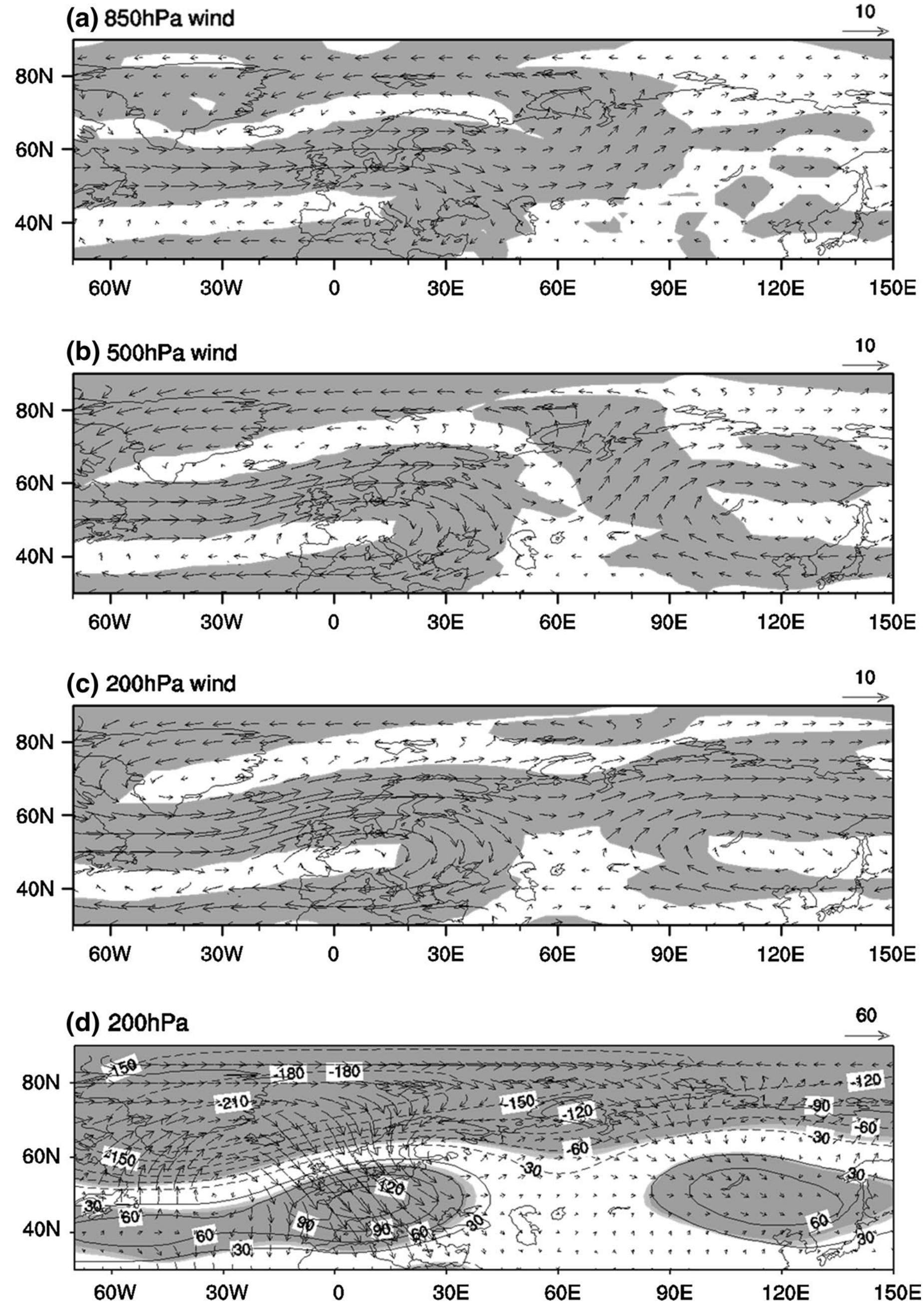

Fig. 6 Wind (JFM) differences between positive- and negative-phase NAO (JFM) years. **a** At 850 hPa, **b** at 500 hPa, and **c** at 200 hPa. **d** Winter (JFM) differences in geopotential height (*contours*) and wave activity flux (*arrows*) between positive- and negative-phase NAO (JFM) years at 200 hPa. *Shading areas* show significant changes at the 95 % level, estimated by a local Student's *t* test

Fig. 9a, b). The physical features of snow, including high albedo, low thermal conductivity, strongly influence the boundary-layer climate (Bednorz 2004). Snow is related to the change in air temperature, precipitation and soil moisture, therefore it is also, indirectly, an important factor for vegetation growth. Furthermore, Ogi et al. (2003) noted that even once melted, snowmelt continues to affect the heating through soil moisture. Thus, early warm conditions in spring over Eurasia can be caused by warm temperatures and low snowfall during the previous winter (Ogi et al. 2003). Snow cover also affects local atmospheric heating, by snow-albedo feedback during the melting season. Specifically, the thinner and weaker snow cover developed during a positive-phase late wintertime NAO year can decrease the albedo and increase temperature. Consequently, warm and moist surfaces over Eurasia can contribute to the vegetation growth.

On the contrary, in negative-phase NAO years, there are anomalies of larger Eurasian snow cover in late winter, which produce a delayed increase in spring surface air temperature and thus contribute to a lower NDVI over Eurasia (Bamzai and Shukla 1999; Hahn and Shukla 1976). The local impact of snow cover on atmospheric temperatures has been discussed in previous studies, which suggest that

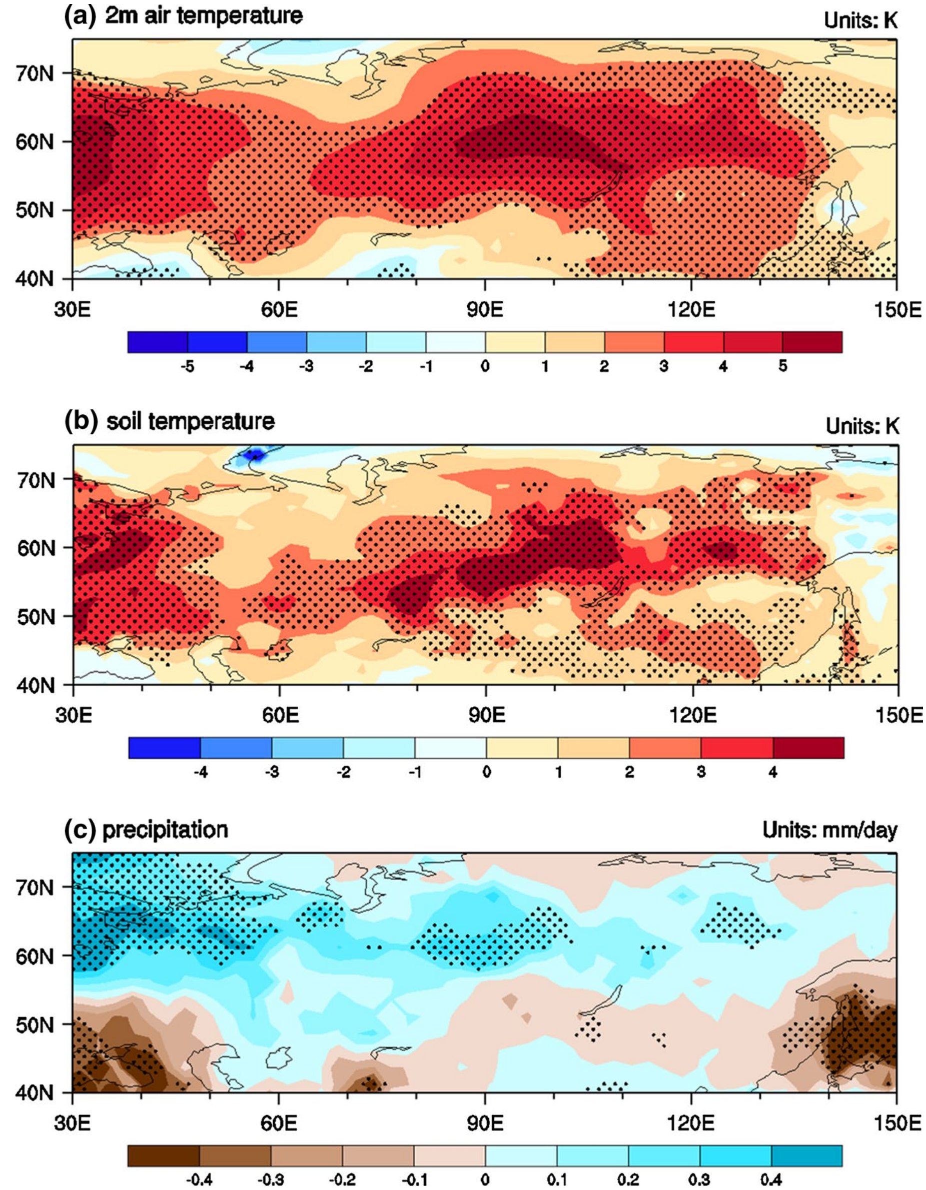

Fig. 7 Winter (JFM) differences in temperature and precipitation between positive- and negative-phase NAO (JFM) years. **a** Air temperature at 2 m (K), **b** soil temperature (K), **c** precipitation (mm/day). *Dotted areas* are the same as Fig. 5

positive feedback between snow cover and solar radiation occurs, due to the increased albedo, the absorption of latent heat by snow melting and sublimation, and the low thermal conductivity of the snowpack (Kim et al. 2013). These local cooling effects, in turn, strongly influence subsequent spring climates over Eurasia. Therefore, the delayed temperature increase in spring also delays the growth and development of the plants.

5 Conclusions

In the present study, we have focused on the significant lagged influence of the late wintertime NAO (JFM) on the subsequent springtime NDVI (AMJ) over Eurasia and its possible physical mechanisms.

We examined the correlation between the springtime EUNDVI (AMJ) and the winter NAO (JFM) index, and identified that their close relationship might enable the NAO index to serve as a predictor for subsequent spring NDVI values over the mid- and high-latitudes of Eurasia. Some 31.4 % of the spring vegetation variance is explained by the previous winter's NAO variations, while the correlation coefficients between the springtime EUNDVI (AMJ) and the late wintertime NAO (JFM) index are 0.50 (containing the linear trend), and 0.56 (de-trended), both of which are significant (α 0.01), as estimated from a standard Student's *t* test. Positive values of the winter

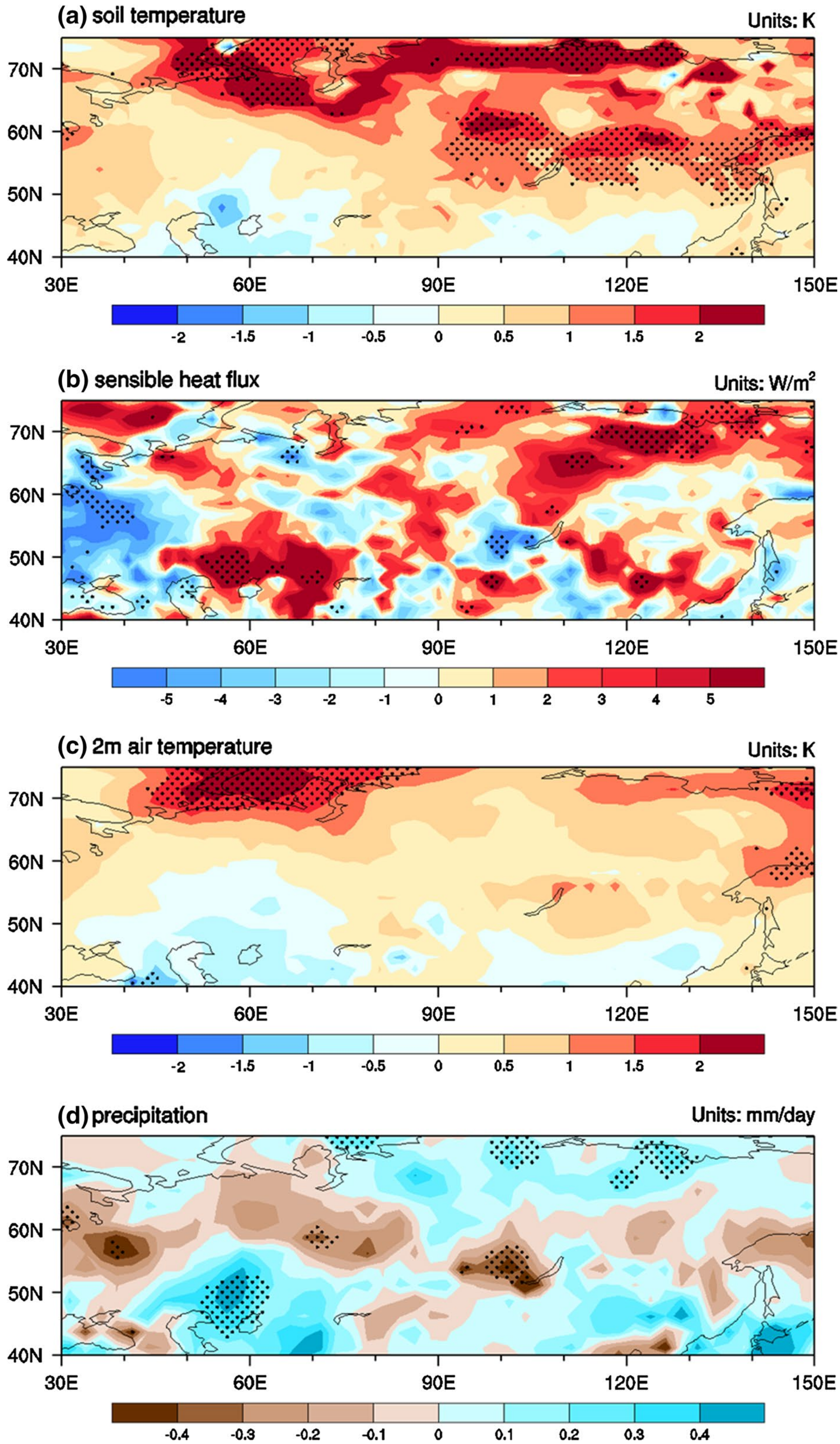

Fig. 8 Spring (AMJ) differences in meteorological variables between positive- and negative-phase NAO (JFM) years. **a** Soil temperature (K), **b** sensible heat flux (W/m^2), **c** 2 m air temperature (K), **d** precipitation (mm/day). *Dotted areas* are the same as Fig. 5

NAO contribute to high vegetation activity in the following spring, and vice versa.

The possible physical mechanisms for the non-simultaneous relationship between the winter NAO and the spring NDVI are inter-related (Fig. 10). A positive phase of the wintertime NAO is associated with an increase in air and soil temperature and more precipitation anomalies throughout most of Eurasia, due to enhanced westerly

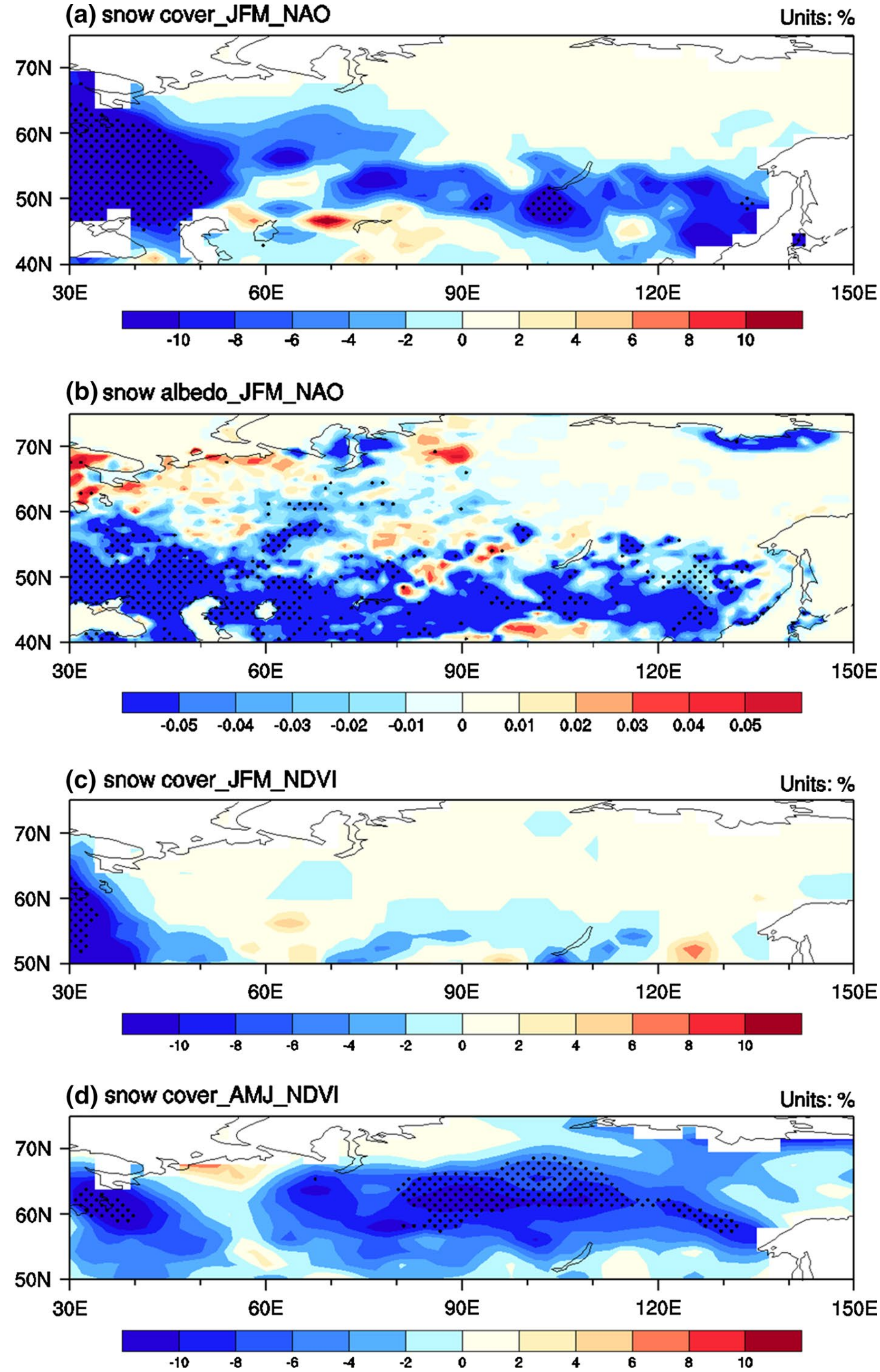

Fig. 9 Differences in winter (JFM) snow cover and snow albedo between positive- and negative-phase NAO (JFM) years. **a** Snow cover (%), **b** snow albedo. Differences in different season of snow cover between the high- and low-value NDVI (AMJ) years. **c** In JFM, **d** in AMJ. *Dotted areas* show significant change at 95 % level, estimated by a local Student's *t* test. Persistent coverage of snow (i.e. coverage is 100 %) is denoted by *blank areas*

winds, which correspond to barotropic cyclonic circulation anomalies over the northern part of Eurasia and anticyclonic circulation anomalies over the southern part of Eurasia. A positive phase of the winter NAO (JFM) also generates the Rossby wave, which propagates south-eastward from Novaya Zemlya to the regions around Lake Baikal (85°E–150°E, 35°N–60°N), and leads to an anomalous high over Lake Baikal. The anomalous high over Lake Baikal in turn leads to an increase in solar radiation also contributes to warming. All the described changes produce negative anomalies of snow cover, which then decreases the albedo and increases temperature further. Warming and moistening of the air occurs over Eurasia and leads to faster snow melt during subsequent spring. Together, these conditions favor vegetation growth at higher than average rates to produce green vegetation cover conditions.

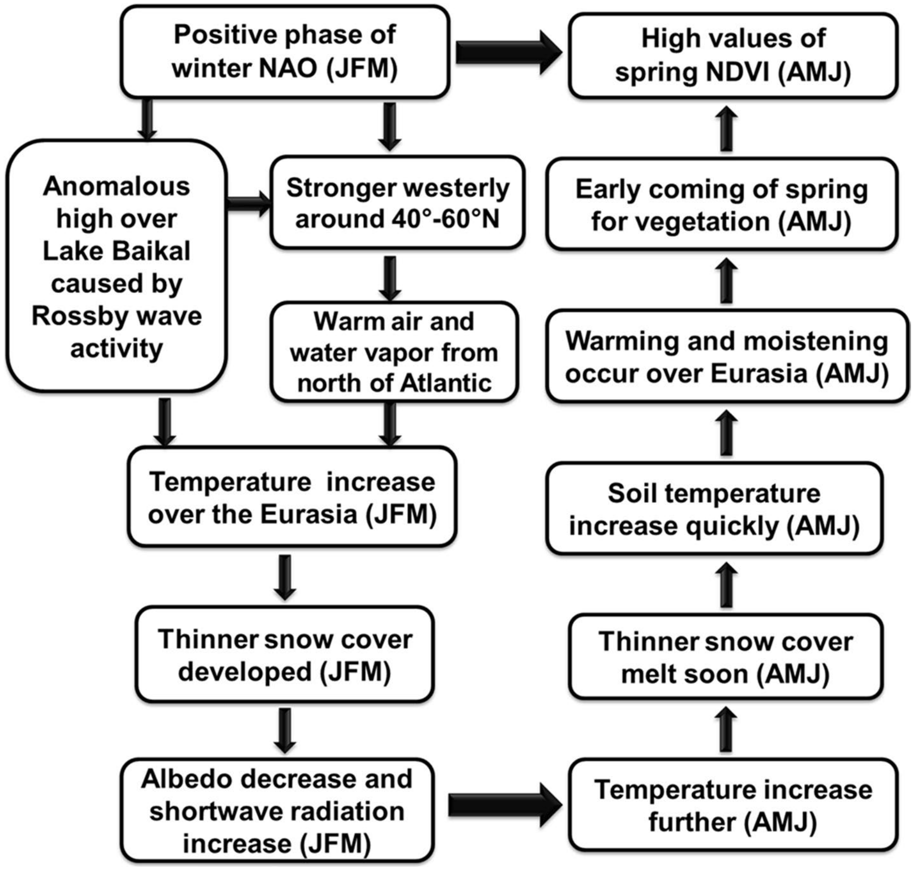

Fig. 10 A schematic diagram of possible physical process that lead to the non-simultaneous linkage between the NAO (JFM) and the NDVI (AMJ) over Eurasia

A negative phase of the late wintertime NAO leads to the opposite effect.

This study demonstrates that one of the likely climatic factors that effectively memorizes the winter NAO and links it to the subsequent spring climate, might be snow cover, which can strongly influence the boundary-layer climate, owing to its physical features. However, it is likely that there is more than one important climatic factor that does this; SST and sea-ice are two other probable candidates that warrant further research.

It has been long recognized that fluctuations in SST and the strength of the NAO in the North Atlantic are related (Bjerknes 1964). Stronger eastward flow generally increases evaporation, thereby cooling SST; on the other hand, SST anomalies, including those in middle latitudes, have been thought to be partly responsible for changes in atmospheric circulation over the North Atlantic and Europe in winter or spring (Rodwell et al. 1999). Czaja and Frankignoul (1999) found a NAO-like signal in early winter is associated with SST anomalies east of Newfoundland and in the eastern subtropical North Atlantic during the preceding summer. Rodwell and Folland (2002) also analyzed that significant links exist between wintertime NAO and SST anomalies in the preceding spring, summer, and autumn. Recently, Zhou (2013) revealed that springtime NAO can induce sea surface temperature anomalies (SSTA) in the North Atlantic, which display a tripole pattern. The spring Atlantic SSTA pattern that could persists to summer, can trigger a high-level tropospheric Rossby wave response in the Eurasia continent, resulting in atmospheric circulation anomalies over the Siberia-Mongolia region.

Meanwhile, sea ice may be another important link between the late wintertime NAO and spring NDVI. Some studies have showed the relationship between NAO and sea ice, which further affected the climate conditions (Kwok 2000; Seierstad and Bader 2009). For example, Hu et al. (2002) found that winter sea ice is thinner in high-NAO index years than in low-NAO index years in the Eurasian coastal region. They indicated that the thinner wintertime ice combined with strengthened southerlies in spring promotes an earlier break-up of the ice pack in the Eurasian coastal region, resulting in significant sea ice export. The higher ice efflux, in turn, further reduces the ice compactness, thus more solar radiation is absorbed by the oceans which enhances the melting process. Chapman and Walsh (1993) indicated that the arctic sea-ice variations over the past several decades are compatible with the corresponding air temperatures, which show a distinct warming that is strongest over northern land areas (northwestern North America, northern Asia) during the winter and spring.

Therefore, the SST and sea ice anomalies may be both other bridges between late wintertime NAO and spring NDVI over Eurasia during 1982–2006, and the physical mechanisms need further research.

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References

Anyamba A, Eastman JR (1996) Interannual variability of NDVI over Africa and its relation to El Niño/Southern Oscillation. Remote Sens 17(13):2533–2548

Bamzai AS, Shukla J (1999) Relation between Eurasian snow cover, snow depth, and the Indian summer monsoon: an observational study. J Clim 12(10):3117–3132

Bednorz E (2004) Snow cover in eastern Europe in relation to temperature, precipitation and circulation. Int J Climatol 24(5):591–601

Betts RA, Cox PM, Lee SE, Woodward FI (1997) Contrasting physiological and structural vegetation feedbacks in climate change simulations. Nature 387(6635):796–799

Bjerknes J (1964) Atlantic air–sea interaction. Adv Geophys 10(1):82

Bogaert J, Zhou L, Tucker CJ, Myneni RB, Ceulemans R (2002) Evidence for a persistent and extensive greening trend in Eurasia inferred from satellite vegetation index data. J Geophys Res Atmos (1984–2012) 107(D11):ACL 4-1–ACL 4-14

Chapman WL, Walsh JE (1993) Recent variations of sea ice and air temperature in high latitudes. Bull Am Meteorol Soc 74(1):33–47

Cho M-H, Lim G-H, Song H-J (2014) The effect of the wintertime Arctic Oscillation on springtime vegetation over the northern high latitude region. Asia Pac J Atmos Sci 50(1):567–573

Cleland EE, Chuine I, Menzel A, Mooney HA, Schwartz MD (2007) Shifting plant phenology in response to global change. Trends Ecol Evol 22:357–365

Czaja A, Frankignoul C (1999) Influence of the North Atlantic SST on the atmospheric circulation. Geophys Res Lett 26(19):2969–2972

Dai Y, Zeng Q (1997) A land surface model (IAP94) for climate studies part I: formulation and validation in off-line experiments. Adv Atmos Sci 14:433–460

Dai S, Zhang B, Wang H (2010) Spatio-temporal change of vegetation index NDVI in northwest China and its influencing factors. J Geoinf Sci 12:315–321

Dee DP et al (2011) The ERA-interim reanalysis: configuration and performance of the data assimilation system. Q J R Meteorol Soc 137:553–597

Déry SJ, Brown RD (2007) Recent Northern Hemisphere snow cover extent trends and implications for the snow-albedo feedback. Geophys Res Lett 34(22). doi:10.1029/2007GL031474

Di L, Rundquist DC, Han L (1994) Modelling relationships between NDVI and precipitation during vegetative growth cycles. Int J Remote Sens 15:2121–2136

Gao X, Zhang D, Chen Z, Pal J, Giorgi F (2007) Land use effects on climate in China as simulated by a regional climate model. Sci China Ser D Earth Sci 50:620–628

Gong D-Y, Shi P-J (2003) Northern hemispheric NDVI variations associated with large-scale climate indices in spring. Int J Remote Sens 24:2559–2566

Gouveia C, Trigo RM, DaCamara CC, Libonati R, Pereira J (2008) The North Atlantic oscillation and European vegetation dynamics. Int J Climatol 28:1835–1847

Hahn DG, Shukla J (1976) An apparent relationship between Eurasian snow cover and Indian monsoon rainfall. J Atmos Sci 33:2461–2462

Hoffmann WA, Jackson RB (2000) Vegetation-climate feedbacks in the conversion of tropical savanna to grassland. J Clim 13:1593–1602

Hu A, Rooth C, Bleck R et al (2002) NAO influence on sea ice extent in the Eurasian coastal region. Geophys Res Lett 29(22):10-1–10-4

Hurrell JW (1995) Decadal trends in the North Atlantic Oscillation: regional temperatures and precipitation. Science 269(5224):676–679

Hurrell JW, Kushnir Y, Ottersen G, Visbeck M (2003) An overview of the North Atlantic oscillation. Geophys Monogr 134:1–35

Ichii K, Kawabata A, Yamaguchi Y (2002) Global correlation analysis for NDVI and climatic variables and NDVI trends: 1982–1990. Int J Remote Sens 23(18):3873–3878

IPCC (2013) Climate change 2013: the physical science basis. Contribution of working group I to the fifth assessment report of the intergovernmental panel on climate change [Stocker TF, Qin DH, Plattner GK, Tignor M, Allen SK, Boschung J, Nauels A, Xia Y, Bex V, Midgley PM (eds)]. Cambridge University Press, Cambridge and New York, NY

Jia GJ, Epstein HE, Walker DA (2006) Spatial heterogeneity of tundra vegetation response to recent temperature changes. Glob Chang Biol 12:42–55

Jia GJ, Epstein HE, Walker DA (2009) Vegetation greening in the Canadian Arctic related to decadal warming. J Environ Monit 11:2231–2238

Jiang D, Zhang Y, Lang X (2011) Vegetation feedback under future global warming. Theor Appl Climatol 106:211–227

Kalnay E et al (1996) The NCEP/NCAR 40-year reanalysis project. Bull Am Meteorol Soc 77(3):437–471

Kawabata A, Ichii K, Yamaguchi Y (2001) Global monitoring of interannual changes in vegetation activities using NDVI and its relationships to temperature and precipitation. Int J Remote Sens 22:1377–1382

Kim Y, Kim K-Y, Kim B-M (2013) Physical mechanisms of European winter snow cover variability and its relationship to the NAO. Clim Dyn 40:1657–1669

Kogan FN (2000) Satellite-observed sensitivity of world land ecosystems to El Nino/La Nina. Remote Sens Environ 74(3):445–462

Kwok R (2000) Recent changes in Arctic Ocean sea ice motion associated with the North Atlantic Oscillation. Geophys Res Lett 27(6):775–778

Li A, Liang S, Wang A, Huang C (2012) Investigating the impacts of the North Atlantic Oscillation on global vegetation changes by a remotely sensed vegetation index. Int J Remote Sens 33:7222–7239

Los SO, Collatz GJ, Bounoua L, Sellers PJ, Tucker CJ (2001) Global interannual variations in sea surface temperature and land surface vegetation, air temperature, and precipitation. J Clim 14:1535–1549

Maignan F, Bréon FM, Bacour C, Demarty J, Poirson A (2008) Interannual vegetation phenology estimates from global AVHRR measurements: comparison with in situ data and applications. Remote Sens Environ 112:496–505

Martínez-Jauregui M, San Miguel-Ayanz A, Mysterud A, Rodríguez-Vigal C, Clutton-Brock TIM, Langvatn R, Coulson TIM (2009) Are local weather, NDVI and NAO consistent determinants of red deer weight across three contrasting European countries? Glob Chang Biol 15:1727–1738

Myneni RB, Los SO, Tucker CJ (1996) Satellite-based identification of linked vegetation index and sea surface temperature Anomaly areas from 1982–1990 for Africa, Australia and South America. Geophys Res Lett 23(7):729–732

Myneni RB, Keeling CD, Tucker CJ, Asrar G, Nemani RR (1997) Increased plant growth in the northern high latitudes from 1981 to 1991. Nature 386:698–702

Nemani RR, Running SW (1989) Estimation of regional surface resistance to evapotranspiration from NDVI and thermal-IR AVHRR data. J Appl Meteorol 28:276–284

Ogi M, Tachibana Y, Yamazaki K (2003) Impact of the wintertime North Atlantic Oscillation (NAO) on the summertime atmospheric circulation. Geophys Res Lett 30(13). doi:10.1029/2003GL017280

Pan LL (2005) Observed positive feedback between the NAO and the North Atlantic SSTA tripole. Geophys Res Lett 32(6). doi:10.1029/2005GL022427

Qian B, Saunders MA (2003) Summer UK temperature and its links to preceding Eurasian snow cover, North Atlantic SSTs, and the NAO. J Clim 16:4108–4120

Rodriguez-Iturbe I, D'Odorico P, Porporato A, Ridolfi L (1999) On the spatial and temporal links between vegetation, climate, and soil moisture. Water Resour Res 35:3709–3722

Rodwell MJ, Folland CK (2002) Atlantic air–sea interaction and seasonal predictability. Q J R Meteorol Soc 128(583):1413–1443

Rodwell MJ, Rowell DP, Folland CK (1999) Oceanic forcing of the wintertime North Atlantic Oscillation and European climate. Nature 398:320–323

Schultz P, Halpert M (1993) Global correlation of temperature, NDVI and precipitation. Adv Space Res 13:277–280

Seierstad IA, Bader J (2009) Impact of a projected future Arctic sea ice reduction on extratropical storminess and the NAO. Clim Dyn 33(7–8):937–943

Sun J, Wang H (2012) Changes of the connection between the summer North Atlantic Oscillation and the East Asian summer rainfall. J Geophys Res Atmos (1984–2012) 117, D08110. doi:10.1029/2012JD017482

Sun J, Wang H, Yuan W (2008) Decadal variations of the relationship between the summer North Atlantic Oscillation and Middle East Asian air temperature. J Geophys Res Atmos (1984–2012) 113(D15). doi:10.1029/2007JD009626

Suzuki R, Tanaka S, Yasunari T (2000) Relationships between meridional profiles of satellite-derived vegetation index (NDVI) and climate over Siberia. Int J Climatol 20:955–967

Takaya K, Nakamura H (1997) A formulation of a wave-activity flux for stationary Rossby waves on a zonally varying basic flow. Geophys Res Lett 24:2985–2988

Takaya K, Nakamura H (2001) A formulation of a phase-independent wave-activity flux for stationary and migratory quasigeostrophic eddies on a zonally varying basic flow. J Atmos Sci 58:608–627

Tian B, Fan K (2012) Relationship between the late spring NAO and summer extreme precipitation frequency in the middle and lower reaches of the Yangtze River. Atmos Ocean Sci Lett 5:455–460

Tucker CJ (1979) Red and photographic infrared linear combinations for monitoring vegetation. Remote Sens Environ 8(2):127–150

Tucker CJ, Slayback DA, Pinzon JE, Los SO, Myneni RB, Taylor MG (2001) Higher northern latitude normalized difference vegetation index and growing season trends from 1982 to 1999. Int J Biometeorol 45:184–190

Tucker CJ et al (2005) An extended AVHRR 8-km NDVI dataset compatible with MODIS and SPOT vegetation NDVI data. Int J Remote Sens 26:4485–4498

Vicente-Serrano SM, Heredia-Laclaustra A (2004) NAO influence on NDVI trends in the Iberian Peninsula (1982–2000). Int J Remote Sens 25:2871–2879

Wang G (2003) Reassessing the impact of North Atlantic Oscillation on the sub-Saharan vegetation productivity. Glob Chang Biol 9:493–499

Wang G, You L (2004) Delayed impact of the North Atlantic Oscillation on biosphere productivity in Asia. Geophys Res Lett 31(12). doi:10.1029/2004GL019766

Wang J, Rich PM, Price KP (2003) Temporal responses of NDVI to precipitation and temperature in the central Great Plains, USA. Int J Remote Sens 24:2345–2364

Watanabe M, Nitta T (1999) Decadal changes in the atmospheric circulation and associated surface climate variations in the Northern Hemisphere winter. J Clim 12:494–510

Yuan H, Dai Y, Xiao Z, Ji D, Shangguan W (2011) Reprocessing the MODIS leaf area index products for land surface and climate modeling. Remote Sens Environ 115:1171–1187

Zeng X, Dickinson RE, Walker A, Shaikh M, DeFries RS, Qi J (2000) Derivation and evaluation of global 1-km fractional vegetation cover data for land modeling. J Appl Meteorol 39:826–839

Zhang X, Friedl MA, Schaaf CB, Strahler AH (2004) Climate controls on vegetation phenological patterns in northern mid-and high latitudes inferred from MODIS data. Glob Chang Biol 10:1133–1145

Zhao H, Moore GWK (2006) A seasonally lagged signal of the North Atlantic Oscillation (NAO) in the North Pacific. Int J Climatol 26(7):957–970

Zhou BT (2013) Weakening of winter North Atlantic Oscillation signal in spring precipitation over southern China. Atmos Ocean Sci Lett 6(5):248–252

Zhou BT, Cui X (2014) Interdecadal change of the linkage between the North Atlantic Oscillation and the tropical cyclone frequency over the western North Pacific. Sci China Earth Sci 57:2148–2155

Zhou L, Tucker CJ, Kaufmann RK, Slayback D, Shabanov NV, Myneni RB (2001) Variations in northern vegetation activity inferred from satellite data of vegetation index during 1981–1999. J Geophys Res Atmos (1984–2012) 106:20069–20083

Zhou M, Wang H, Yang S, Fan K (2013) Influence of springtime North Atlantic Oscillation on crops yields in Northeast China. Clim Dyn 41:3317–3324