Cryosphere-atmosphere interaction related to variability and change of Northern Hemisphere annular mode

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1. Introduction

Factors such as storminess, atmospheric temperature, and the intensity and duration of precipitation events over the Atlantic–European sector are determined primarily by the phases of the Northern Hemisphere annular mode also known as the Arctic Oscillation/North Atlantic Oscillation (AO/NAO) [1]. The positive AO/NAO phase is characterised by stronger sub polar westerlies, with the opposite conditions occurring during a negative AO/NAO phase. These patterns are particularly pronounced during cold months, but are nevertheless evident throughout the year. The growth of weather systems across the North Atlantic storm track and the interaction between these systems and the larger scale mean flows are of central importance in determining the characteristics of the AO/NAO [2]. However, the AO/NAO is also influenced by the interaction of troposphere dynamics with external factors in the stratosphere, and is subject to the influences of gradual changes that occur at the Earth's surface. Ocean temperatures, in particular, represent the most significant forcing mechanism for AO/NAO-related atmospheric variability, given the time scales of oceanic circulation and the large heat capacity of the oceans [3]. However, the effects of land–atmosphere interaction have also received attention. Seasonally frozen ground and snow cover are the largest components of the cryosphere. Snow covers up to 46.5% of the Earth's total land surface in January and about 98% of the total seasonal snow cover is located in the Northern Hemisphere [4].

Two types of out-of-phase relationships between the winter AO/NAO and Eurasian snow-cover extent have been described in previous studies. These are: (1): a quasi-simultaneous relationship (January-March) [5] that can persist until summer ([6]; [7]) and (2) a lagged relationship linking summer, autumn and/or warm-season (April-October) snow cover with the following AO/NAO ([5]; [8]; [9]; [10]). The simultaneous correlation between positive (negative) AO/NAO and negative (positive) snow cover anomalies is mostly due to warm (cold) thermal advection over the continental areas in mid- and late winter. In this case, modelling studies have also revealed a positive feedback: positive (negative) snow anomalies favour negative (positive) AO/NAO phase which in turn enhances those anomalies [11]. On the other hand, snow anomaly persistence from winter to summer is associated with Northern Hemisphere blocking variability [12]. However, the fundamental mechanisms determining the lagged relationship between snow cover in the interval April-October and the next winter AO/NAO are not yet completely understood. The main aim of this paper is to investigate how the variability in snow cover is related to AO/NAO-type atmospheric circulation, in order to explain this lagged relationship.

2. Data and methods

The primary data set consisted of monthly data from the European Centre for Medium-Range Weather Forecasts (ECMWF), comprising a 40-year reanalysis (ERA-40) of zonal wind and geopotential heights over the Northern Hemisphere from 1958–2002, obtained from the ECMWF data server (http://data.ecmwf.int/data/). The frequency and extent of anomalies of Eurasian snow cover were derived from the weekly 89X89 Snow Cover Data, which is available from the NOAA Climate Prediction Centre (CPC, see http://www.cpc.ncep.noaa.gov/data/snow/). This snow archive covers the period 1973 to the present day. A given grid point is said to be snow covered when the snow extent exceeds 50% of the cell area. The present study used snow data from 1973 to 2004.

In our analysis we have defined the warm and cold periods as April–October and respectively December–February because this replicates the annual evolution of correlations between the extent of monthly snow cover and the AO/NAO presented in [8]. In addition, this temporal partitioning approach reflects the seasonal cycles of both atmosphere and cryosphere variability over mid and high latitudes of Eurasia [2]. Reconstructions of the extent of snow cover in April and October were taken from [13] and their average was used as an approximation of the warm-season value for the period 1930–1990 since the largest variations in snow cover come from thawing and freezing periods (climatically confined in April and October, respectively (eg. [14]; [15]; [16]).

Frozen ground variables for Northern Asia were taken from [17]. Permafrost is a layer of soil or rock, at some depth beneath the surface, in which the temperature has been continuously below 0 °C for two or more years. The active layer depth is defined as the layer of ground that is subject to annual thawing and freezing in areas underlain by permafrost, while the seasonal freeze depth is defined as the layer of ground in non-permafrost regions that is subject to freezing during the cold season. In addition, we used monthly soil temperatures for 1956–2000 at seven depths from 89 Siberian stations located between 50°–60°N and 80°–115°E [16]. The December to March AO/NAO index computed by [18] (http://www.cgd.ucar.edu/~jhurrell/AO/NAO.html) was used to quantify AO/NAO variability.

We identified land and atmosphere related signals by performing canonical correlation analysis (CCA) and ciclostationary empirical orthogonal function analysis (CEOF) on the observed data. The CCA selects coupled spatial patterns of two variables such that their time evolution is optimally correlated (e.g. [19]). Before canonical correlation analysis, the original data are usually projected onto their Empirical Orthogonal Functions (EOFs), retaining only a limited number of them in order to minimize the noise. The CEOF analysis extracts patterns that are phase locked with the annual cycle and reveals their evolution throughout the year [20]. In our analysis the nested period was 12 months. For the monthly time series, the CEOF procedure yields periodic loading vectors, one for each month of the year, which represent the nested fluctuations.

The cryosphere - atmosphere relationship over Eurasia

We began by analysing the active layer depth and maximum seasonal freeze depth for Eurasia. It's worth noting that extreme values of soil freezing depth are of interest for practical applications (e.g. engineering design specifications). The active layer and freeze depths were spatially averaged for Russian regions using the available time series, as in [17]. We focused our analysis on the warm (April–October) and cold (December–February) periods. Figure 1 shows that seasonal freeze and active layer depths show insignificant positive correlations with the AO/NAO up to the 1970s, but they divert after that and evolve opposite correlations with the AO/NAO. Since around 1965 the 15-year running correlation coefficients between the AO/NAO and the concurrent maximum freeze depth is consistently significant at the 95%-level. Negative anomalies of maximum freeze depth are physically consistent with positive temperature and snowfall anomalies due to a positive phase of AO/NAO leading to an intensification of storm activity over northern Eurasia. As for the active layer depth, positive anomalies are physically consistent with a positive phase of previous AO/NAO which enables higher soil temperatures leading to enhanced thawing in spring and summer. Since around 1973 the active layer depth is correlated significantly with the previous winter's AO/NAO. The delayed onset of significant correlation between active layer depths and AO/NAO could be explained by the higher thermal inertia of soils underlined by permafrost. These changes coincide with a reported eastward shift in the AO/NAO centre of action, and with the strong increasing trend in the AO/NAO index [21] which may help to explain why the AO/NAO-cryosphere relationship is significant in more recent years, but less so before this. In recent decades the eastward shift has been associated with a spatially coherent climate regime during which the AO/NAO index, Siberian wintertime temperatures, and North Atlantic storm activity have all become correlated significantly with each other [22]. In figure 1 it is worth noting the presence of a decadal component superimposed on the interdecadal snow-AO/NAO variability structure. Saito et al. [23] have also revealed this decadal behaviour analysing the hemispheric-scale covariability of snow and atmosphere for autumn-to-winter and winter-to-spring periods in the last three decades.

We find that the AO/NAO's influence on the Eurasian cryosphere has also increased significantly in recent decades, reaching values of −0.79 and 0.71 for seasonal freeze depth and active layer thickness respectively (Figure 1). At the same time, the extent of the snow cover in Eurasia in the warm season has become correlated more significantly with the following winter's AO/NAO [9], which suggests that there is a causal link between changes in the cryosphere over Eurasia and AO/NAO variability and predictability. However, the active layer depth and seasonal freeze depth for Eurasia do not show any predictive AO/NAO-related signal. It would appear that although AO/NAO circulation affects almost all mid- and high Eurasian latitudes, any feedback to the atmosphere that affects the AO/NAO itself may only be observed at a regional scale. Attempts to detect if there is a predictive signal in seasonal freeze and/or active layer depth at regional scales have suggested that even though these variables have a response to the NAO, they are not useful predictors. Observed data limitations due to incomplete spatial cover of seasonally frozen ground and active layer underlain by permafrost might be an explanation. Hence, these two variables will not be used in the following sections of the paper.

In order to identify key regions for AO/NAO predictability on the basis of the snow cover in Eurasia, we computed Northern Hemisphere CCA modes of snow frequencies during the previous warm season and winter sea level pressure (SLP) for 1973–2002 respectively (Figure 2, filled contours and lines). The snow frequencies are defined at each 2°×2° grid point as the number of weeks with snow divided by the total number of weeks in a time interval. The correlation coefficient between the time series modulating CCA spatial patterns is 0.81 and the fraction of total variance explained by CCA mode is 0.18 and 0.15 for SLP and snow frequency, respectively. Strong anomalies are found in snow frequencies over southern Siberia, centered around 55°N and to a lesser extent over Northern Europe. Other studies have also pointed these regions as teleconnection areas in the cryosphere-AO/NAO linkage [eg. 24], but their role appears to be limited to certain months within the warm period. A lower (higher) snow frequency in the previous warm season is followed by a positive (negative) AO/NAO phase. The correlation coefficient between the following winter's AO/NAO and snow frequency averaged over southern Siberia (from 50°N to 60°N and from 80°E to 115°E) in the warm season is -0.54 (statistically significant at the 99%-level) for the period 1973–2002. This raises the issue of what might be the reasons for snow-related processes being more strongly linked with AO/NAO variability in the regions identified previously.

By performing a CEOF analysis of soil temperature at seven different depths, averaged over Southern Siberia (from 50°N to 60°N and from 80°E to 115°E), we were able to describe the nature of the local soil-atmosphere interaction on a multiannual scale. The nested period of CEOF analysis was 12 months and the analysed period was January 1967 to December 2000. The period 1967-2000 was used because it is the one that reveals a significant linkage with the AO/NAO (Figure 1). Results shown in Figure 3, which illustrates the first CEOF of primary and detrended data, suggest that the late winter/early spring soil temperature anomalies can re-emerge in October. At the beginning of the warm season, the thermal signal associated with the thawing propagates into the soil. Its signature reflects an early or delayed snow melting, depending on the conditions of the previous winter. An early (delayed) snow melting produces positive (negative) thermal anomalies and reduced (enhanced) soil moisture as the season progresses. In midsummer, thermal anomalies occur deeper in the soil. A number of studies (e.g. [25]; [26];[27]) showed that soil moisture-rainfall feedback is an important mechanism affecting hydrological persistence during late spring and summer. This positive feedback continues to produce thermal and moisture anomalies in the soil during the warm season. In October, during soil cooling, thermal anomalies are transferred to the atmosphere, causing geopotential, and zonal wind anomalies. Land-atmosphere coupling is modulated by the frequency of snow cover and this interaction is significantly diminished when snow covers a significant part of the area under analysis. At the end of the warm season, the soil-related anomalies are likely to cause jet stream anomalies in the upper troposphere and to excite non-local responses that will lead to AO/NAO-type variability the following winter. The October soil conditions depend on spring anomalies, as our CEOF analysis of soil temperature has revealed (Figure 3), explaining the lagged signal between the winter AO/NAO and the preceding April to October snow frequency over Eurasia.

In order to further investigate the nature of cryosphere-atmosphere relationship we performed a CEOF analysis on zonal wind anomalies at the 300 hPa level for the period 1967-2000 to identify multiannual patterns of jet stream fluctuations. Four of 12 patterns of the first CEOF mode are presented in Figure 4. Out-of-phase zonal wind anomalies centered along 70°N, 50°N and 30°N alternate over Asia and Europe, showing a similar pattern in April and October. It is interesting to note that the tripole of zonal wind anomalies at 300 hPa is shifted northward in Europe when compared with Asia, which is consistent with the location of snow frequency anomalies over Northern Europe and Southern Siberia. Thus, at the same latitudes the zonal wind pattern is opposite in Europe and Asia in these months, but in phase in winter, when the snowline is located further southward both in Europe and Asia. Opposite wind anomalies along 70°N and 50°N over Europe and Asia are also consistent with acceleration and deceleration of polar jet stream in high troposphere. The October wind pattern at 300 hPa supports the existence of a mechanism in which thermal soil fluctuations leading to snow frequency anomalies over the key regions influence atmospheric processes.

CEOF analysis of observed zonal wind anomalies at 300 hPa shows that anomalies in October over our key regions are related to AO/NAO type patterns the following December (Figure 4). The correlation coefficient between the principal components associated with CEOF spatial patterns of zonal wind and soil temperature is 0.50 (statistically significant at the 99%-level). Time series of soil temperature and zonal wind show increasing trends in the period of interest (1967-2000) (Figure 5). The increasing trend is stronger for soil temperature and increases with depth. We also applied CEOF analyses to detrended data. These analyses show similar results in terms of patterns and correlation coefficients. The local forcing appears to be related to the presence of an area characterised by high seasonal snow variability [9] and, in addition for Southern Siberia, in a location with topographic forcing, which probably favours the propagation of stationary waves into stratosphere [28]. Figure 6 presents four of the first twelve CEOF patterns of geopotential height at 200 hPa applied to the same time period (1967-2000). Winter month patterns reveal AO/NAO-type structures. It is interesting to note the persistence of the same type of anomalies over Asia and Eastern Arctic throughout the year, even though the centres of variability are slightly displaced in location and their intensities vary. The correlation coefficient between the principal components associated with CEOF spatial patterns of geopotential height and zonal wind is 0.69 (statistically significant at the 99%-level) but 0.50 (statistically significant at the 99%-level) for geopotential height at 200 hPa and soil temperature. When the same type of analysis is applied to geopotential heights at 500 hPa, the correlation coefficient with the principal component of soil temperature is only 0.30 (statistically significant at the 99%-level) which suggests that the land-related signal becomes stronger with height in the atmosphere as one would expect from a thermal related mechanism.

Conclusions

There have been extensive efforts to explain recent upward trends in the AO/NAO index, by invoking external factors and natural variability. Several mechanisms have so far been proposed, such as: (1) North Atlantic sea surface temperature SST changes in the last-half century [29], (2) an upward trend in SSTs over the tropical Indian and Pacific Oceans [30]; [31], and (3) processes that affect the strength of the stratospheric polar vortex via downward wave propagation from the stratosphere [32]. Here, we propose a new mechanism for amplifying AO/NAO variability and change, namely land-atmosphere coupling in Eurasia.

Our study identifies the persistence of a geopotential height pattern resembling Arctic Oscillation (AO) throughout the annual cycle, even though the centres of variability are slightly displaced in location and their intensities vary seasonally (see Figure 6). This pattern is related to interannual AO/NAO persistence and our CEOF analysis of soil temperatures suggests that the persistence could be the effect of land-atmosphere interaction over the huge continental mass of Eurasia, which becomes stronger at the beginning and at the end of warm season. The CEOF analysis of zonal wind has suggested that in the cold season the atmospheric circulation control is taken by the Atlantic-European sector due to the main role of enhanced thermal gradients driven by the ocean-land contrast. Also, in the cold season, Asian climate variability is lead by the concurrent AO/NAO. On the other hand, in the warm season, Eurasian land mass is more actively involved in maintaining AO type geopotential height anomalies over high latitudes and higher in atmosphere.

The CEOF analysis of soil temperature suggests that diminished (enhanced) Eurasian snow cover in early spring that is associated with a positive (negative) phase of winter AO/NAO results in positive (negative) thermal anomalies that propagate in the soil during the warm season. A number of studies have demonstrated that soil moisture-rainfall feedback leads to the persistence of thermal and hydrologic anomalies during late spring and summer (e.g. [25]; [26]; [27]). In October, during soil cooling, soil heat and moisture anomalies are propagated back into the atmosphere, causing thermal, geopotential, and zonal wind anomalies. CEOF analyses of zonal wind at 300 hPa and geopotential height at 500 hPa add credence to the hypothesis that the April to October signal is transferred to the following winter, thus explaining the statistical relationship between the frequency of snow cover from April to October and the AO/NAO phase in the following December to March. These results suggest that a positive feedback loop links the warm and cold season anomalies of the Eurasian cryospheric components. This feedback could contribute to the amplification of the red shift in the AO/NAO spectra observed in recent decades.

The results yielded by this study are relevant for both seasonal forecasting (e.g. [33]) and climate change projections. Decreased snow cover in spring, coupled with active layer thickness and decreased seasonally frozen ground depth [10] under global warming conditions imply an increased capacity to store heat and moisture in the warm season, which are delivered to atmosphere exciting non-local positive AO/NAO-related perturbations through jet stream fluctuations in early winter. The AO/NAO may be considered an internal mode of extratropical tropospheric dynamics, and as such its predictability is likely to be the result of more than one mechanism, involving SSTs, snow cover, sea ice, and the dynamics of the lower stratosphere. Thus, climate models with resolved freeze/thaw cycles in soil, snowpack, and full low stratospheric dynamics, in combination with realistic topographic forcing, could be used to improve predictions of the AO/NAO under present and future climate conditions.

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Figure captions

Figure 1. Interdecadal evolution of AO/NAO-related atmosphere-cryosphere interaction: 15-year running correlation coefficients between Hurrell's December-March (DJFM) AO/NAO index and maximum freeze depth (dotted line), active layer thickness (thick solid line), April and October snow cover index (dashed line) and April–October snow cover index (thin solid line). The two horizontal lines represent statistically significant values at the 95% level.

Figure 2. Spatial pattern of canonical correlation analysis (CCA) between April to October snow frequencies (shaded) and the following December to February SLP (contours) (1973-2002). Primary data sets are standardised. The region marked with a rectangle is further analyzed from the standpoint of soil temperatures.

Figure 3. First CEOF of soil temperatures (in °C) averaged at 7 levels for the period January 1967- December 2000 (a) and the same analysis applied to detrended data (b). Depths are in cm.

Figure 4. Patterns of the first CEOF of zonal wind anomalies at 300 hPa (in m/s) for the period January 1967- December 2000.

Figure 5. Principal components (standardised anomalies) associated with the first CEOF of soil temperature (dashed line) and zonal wind at 300 hPa (solid line). Increasing trend lines for the two time series are also shown.

Figure 6. Patterns of the first CEOF of geopotential height anomalies at 200 hPa (in gpm) for the period January 1967- December 2000.

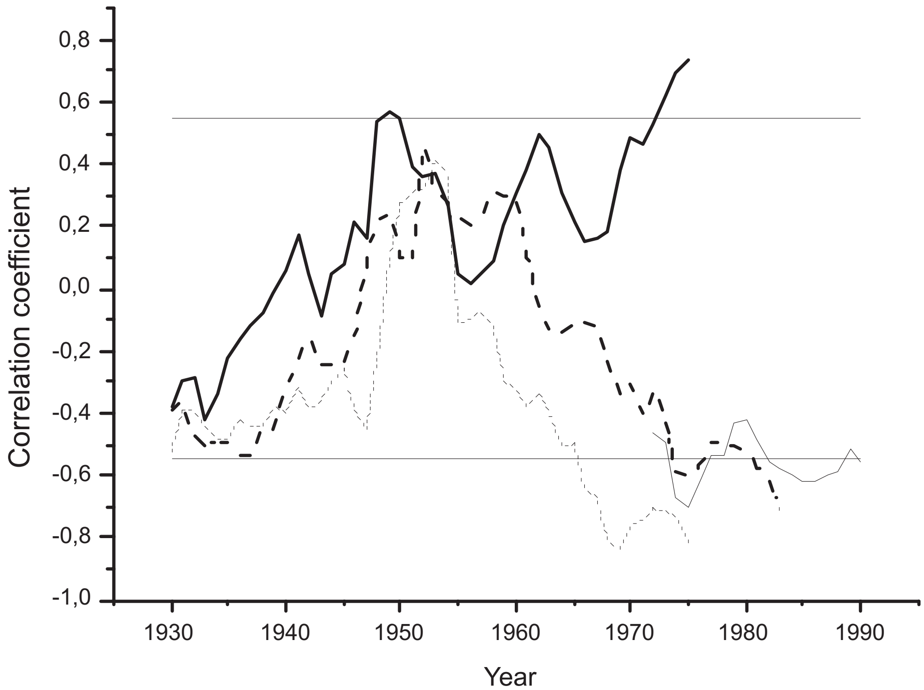
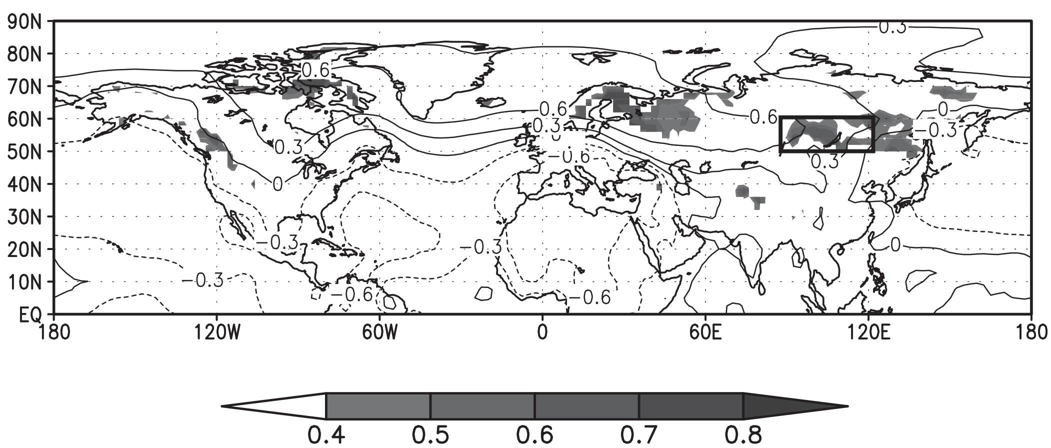


Figure 1



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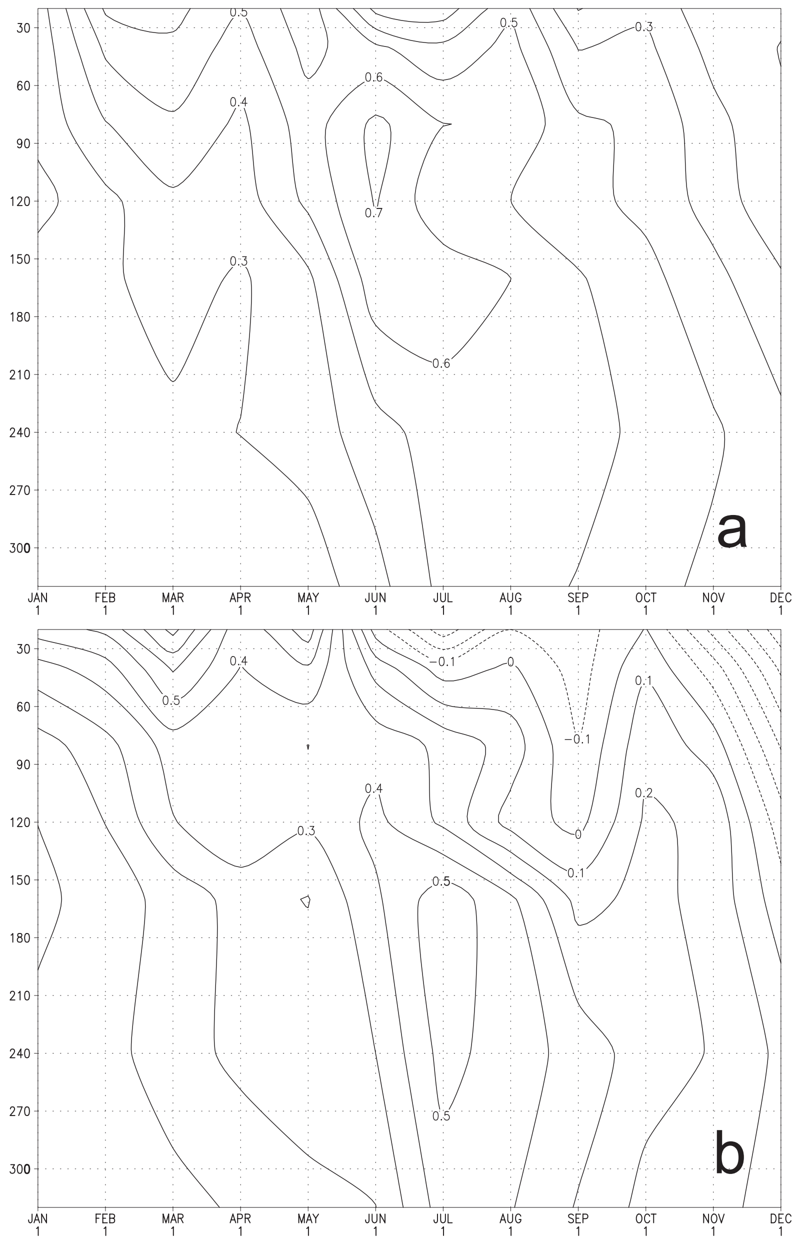

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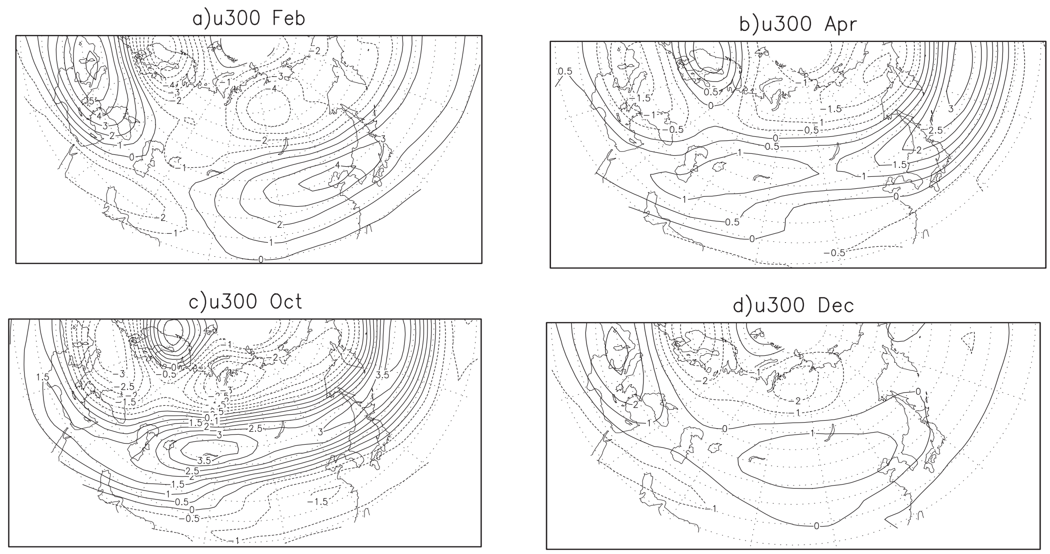
Figure 3.



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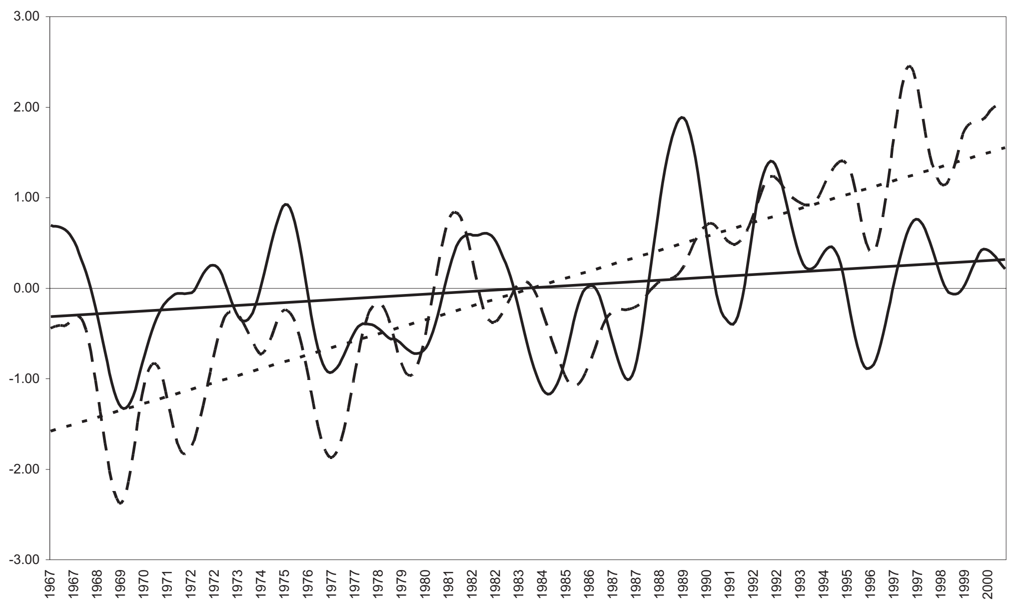

Figure 5.

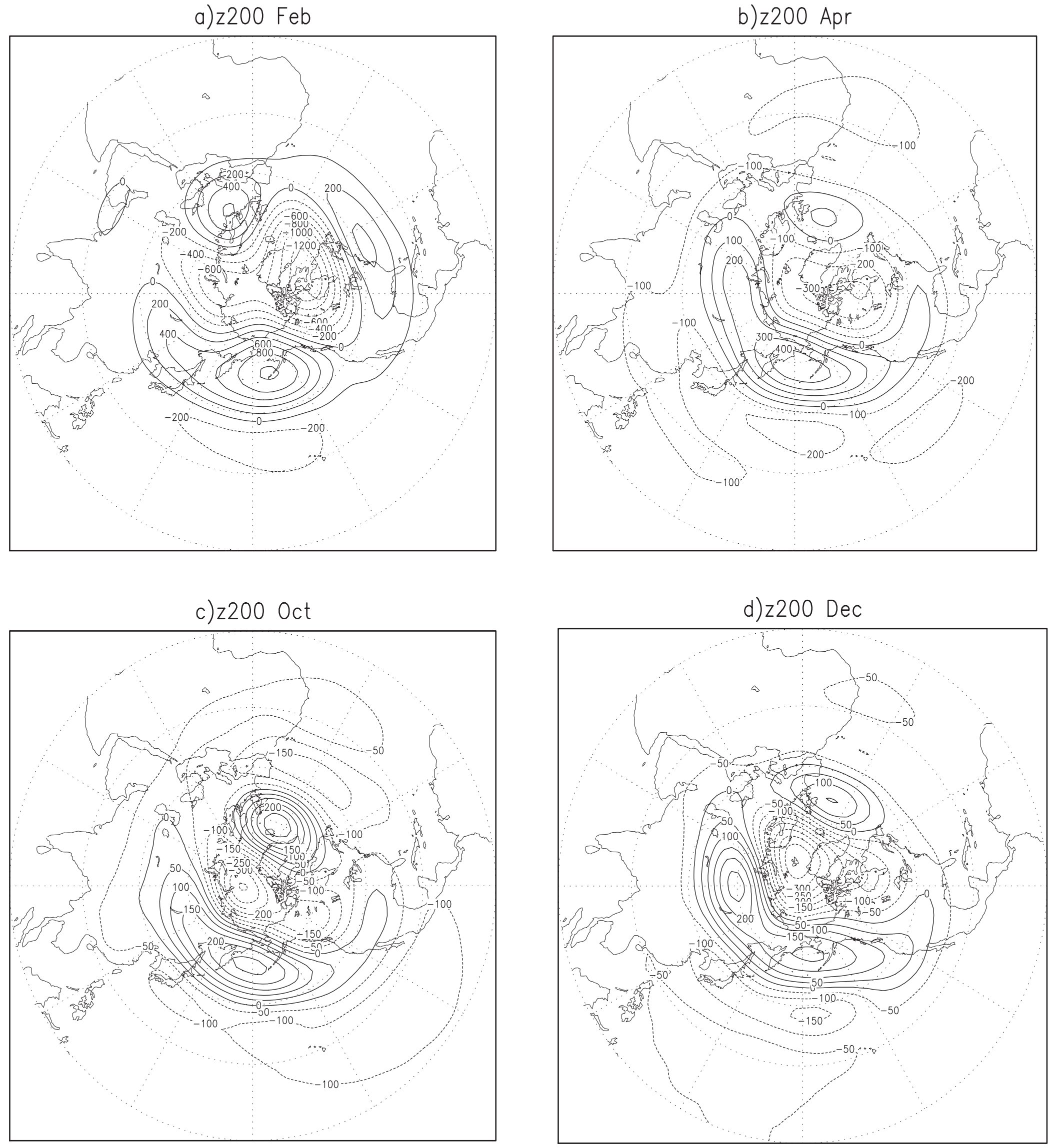

Figure 6.