Paper IV

Shujie Ma, Nils Gunnar Kvamstø, Yongjia Song and Asgeir Sorteberg (2007)

The potential contribution of autumn Eurasian snow cover to wintertime NAO.

The potential contribution of autumn Eurasian snow cover to wintertime NAO

Shujie Ma[1] Nils Gunnar Kvamstø[1,2] Yongjia Song[1]
Asgeir Sorteberg[1]
[1] Bjerknes Centre for Climate Research, University of Bergen, Norway
[2] Geophysical Institute, University of Bergen, Norway

ABSTRACT

In this paper we show on the basis of SVD analysis that the extensive (deficient) snow cover over large areas of Northern Europe and Siberia in October heralds the negative (positive) DJF NAO. When the Eurasian snow cover extent is anomalously high in October, it is driven simultaneously by the upstream meridional circulation at mid- and high latitudes, and the subsequently atmospheric dynamical evolution is important for the wintertime negative NAO. Under conditions with less than normal Eurasian snow cover, the possible response of the positive NAO to Eurasian snow is mainly through the persistent thermal heating over Siberia in lower troposphere and the enhanced effect of Polar/Eurasia teleconnection pattern on the circumpolar vortex.

1. Introduction

The North Atlantic Oscillation (NAO), which shows the concurrent variation in the strength of the Icelandic Low and the Azores High, is an important teleconnection pattern in the extratropical atmosphere, ranging from the surface to lower stratosphere (Wallace and Gutzler, 1981; Hurrell, 1995 and 1996; Thompson and Wallace, 1998; Rodwell et al., 1999; Wallace, 2000). It appears in all seasons, but is most prominent during winter.

In recent years several physical mechanisms to influence the variability of wintertime NAO have been proposed. It is pointed out that a large portion of the NAO variability arises from internal atmospheric processes (Osborn et al., 1999; Feldstein, 2002). However, there are indications that surface forcings, such as sea surface temperature anomalies (SSTA) and continental snow cover variations, play a significant role for the variability of the NAO. For example, the positive trend in the NAO during the last thirty years has been linked to a progressive warming of the tropical Indian and Pacific SST (Bader and Latif, 2003; Hoerling et al., 2001 and 2004). The tripole SSTA pattern over North Atlantic is also coherent with the NAO variations on most time scales from intraseasonal to interdecadal (Walter and Graf 2002; Paeth et al., 2003). But the coherence was not stationary during the whole time period investigated. On the other hand studies have indicated that the extent of autumn Eurasian snow cover may modulate the variability of the NAO through its thermodynamic influence on atmospheric anomalies in lower troposphere, largely by snow-albedo feedback (Cohen and Entekhabi, 1999; Kumar and Yang, 2003), and through its regional effect on vertical propagation of wave fluxes associated with the interaction between stratosphere and troposphere (Saito et al., 2001; Cohen et al. 2002; Gong et al., 2003).

Seasonal lagged connections between boundary forcings and the NAO have been studied as well. The North Atlantic SSTs in May contributes to wintertime NAO through

Atlantic air-sea interaction (Rodwell and Folland, 2002). Due to the weak persistence in seasonal changes of snow cover comparing to the SSTs, the potential modulation of snow fluctuations to the variability of the NAO may be important on shorter time scales. The Eurasian snow cover in October can influence subsequent extratropical atmospheric variability that modulates the winter NAO (Saito and Cohen, 2003).

Due to the poor seasonal predictability of the NAO based on snow cover forcing (Cohen et al., 2002), the dynamical and thermodynamic effect of snow forcing on the variability of the NAO need to be studied further. On the basis of observation analysis on the evolution of atmospheric anomalies from October to December under different Eurasian snow conditions, the goal of this paper is to study the potential role of autumn Eurasian snow cover on wintertime NAO.

2. Coupled pattern between autumn Eurasian snow cover and wintertime NAO

To explore possible large-scale links between autumn Eurasian snow cover and wintertime NAO, we first conduct a Singular Value Decomposition (SVD) analysis on the standardized fields of winter-mean (DJF) sea level pressure (SLP) over the Atlantic sector (20-80°N, 90°W-40°E) and autumn-mean (SON) snow cover extent over Eurasia (30-75°N, 0°-179°E). The SVD technique identifies corresponding modes in two fields through maximizing the covariance (correlation) between the two (standardized) fields (Bretherton et. al. 1992; Wallace et al. 1992). This implies that the coupled pairs of spatial patterns statistically tend to occur synchronously and could be physically related to each other. The linearly coupled degree of two fields is evaluated by Squared Covariance Fraction (SCF) (resembling explained variance) (Bretherton et. al. 1992). We have used weekly snow cover data from 1972 to 2000 from the National Snow and Ice Data Center (NSIDC) (Armstrong and Brodzik, 2002). All the monthly reanalysis data in this study are obtained from European Centre for Medium-Range Weather Forecasts (ECMWF references Uppala et al. 2005; ECMWF 2002).

The leading two SVD modes account for 70% of the total variance between the two fields (Table 1). From the third mode the individual SCF begins to be less than 0.1. This means that the linear relationship between autumn Eurasian snow cover anomaly and wintertime circulation variation over the Atlantic sector is sufficiently represented by the first and second SVD modes (SVD1 and SVD2).

	SVD1	SVD2	SVD3	SVD4
squared covariance fraction (SCF)	0.5558	0.1451	0.0697	0.0539
accumulative SCF	0.5558	0.7009	0.7706	0.8246

Table 1. The squared covariance fraction (SCF) and accumulative SCF for the first four pairs of singular vectors from SVD analysis on wintertime SLP over Atlantic sector (20-80°N, 90°W-40°E) and autumn snow cover over Eurasia (30-75°N, 0°-179°E) .

The SVD1 explains 55.6% of the total variance and the correlation between the first pair of extension coefficients is significant (r=0.71) at the 99.9% critical level (Fig. 1c). It is interesting to see that an evident NAO pattern exhibits in Fig. 1a, and that the time series of SVD1 for wintertime SLP is significantly correlated (r=0.99) with the observed DJF NAO index (http://www.cgd.ucar.edu/cas/jhurrell/indices.html) (Fig. 1d). So the spatial pattern and temporal variation of wintertime NAO is well represented by SVD1. Corresponding to the convincing NAO representation by SVD1, the coupled pattern of snow cover is characterized by a large negative area of snow cover anomaly extending eastward and southward from northern Europe to Siberia (Fig. 1b).

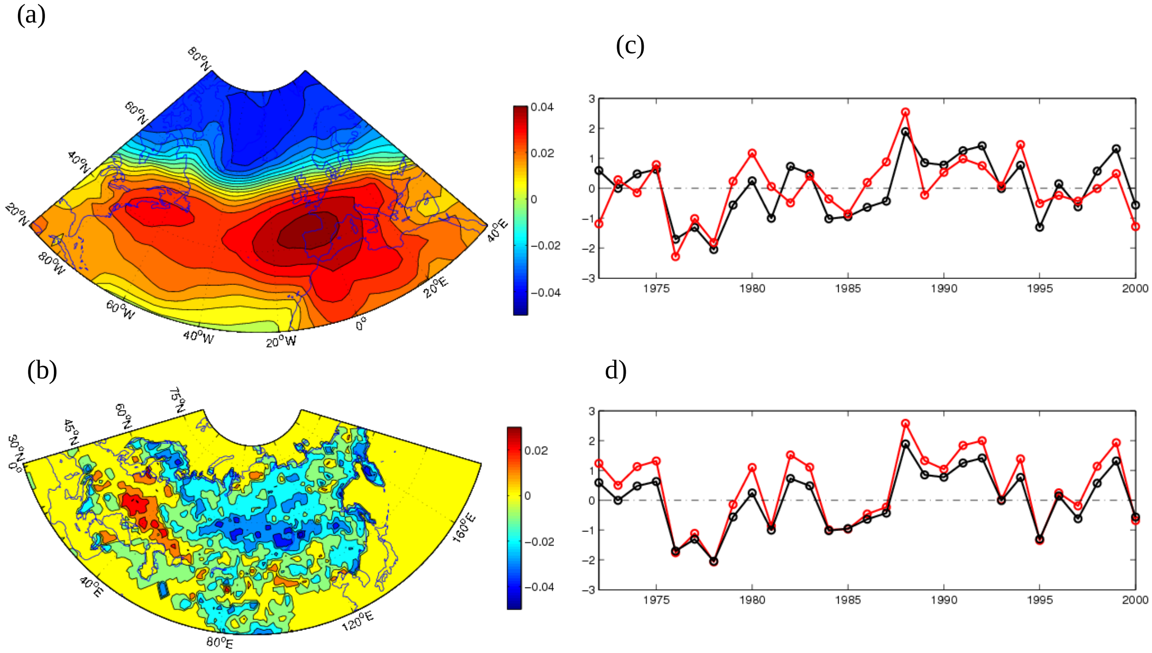

Figure 1. The first pair of singular vectors from SVD analysis on normalized fields of monthly sea level pressure (SLP) in winter (a) (contour interval is 0.005) and Eurasian snow cover in autumn (b). (c) The standardized expansion coefficients for SVD1, black line is for SLP, red line for snow cover, the indicated year corresponds to the Dec. of the winter season, e.g. 1975 means the winter in 1975/1976. (d) The DJF NAO index got from (http://www.cgd.ucar.edu/cas/jhurrell/indices.html) (red line) and expansion coefficients for SVD1 (black line).

The extension coefficient of snow SVD1 (Fig. 1c) shows close connection (r=0.79) with area-averaged snow cover over Eurasia. In addition, it is very consistent with the result of Robinson (1996) who pointed out that snow cover got peak values in late 1970’s and record minimum values in late 1980’s and early 1990’s. This indicates that the snow SVD1 may represent the large-scale anomaly of autumn Eurasian snow cover and the major contribution to the NAO variability.

To gain more confidence on the linear link in the leading modes between both fields, an Empirical Orthogonal Function (EOF) analysis is first performed on autumn Eurasian snow cover. When the DJF SLP field over the Atlantic sector is regressed against the principal component of the leading snow EOF mode (EOF1), this gives rise to the NAO pattern (not shown). The snow EOF1 and its regressed SLP spatial pattern are very similar to the one obtained by SVD analysis, and the correlation between the time series of snow EOF1 and SVD1 is significant (r=0.9) at a 99.9% confidence level. This indicates that the coupled pattern of SVD1 is robust.

It should also be mentioned that the correlation coefficient between snow EOF1 and observed NAO index is -0.48, significant at a 95% confidence level. It is less than that for SVD1 (0.71). The latter is higher than the correlation between area-mean Eurasian snow cover in early seasons and the DJF NAO index (-0.31 and -0.55) in Cohen and Entekhabi

(1999) and Bojariu and Gimeno (2002) and similar to the correlation between the DJF AO index and October SLP/snow index (0.7) in Cohen et al. (2001). This means that the SVD technique picks out the most dominant and closest linear leading modes between the two fields. So the coupled correlation of SVD1 indicates that deficient (extensive) snow cover over northern Europe and Siberia northward of 45°N in SON heralds the positive (negative) DJF NAO, a result in agreement with other studies (Cohen and Entekhabi 1999; Saito et al. 2001).

The SVD2 reflects the coupled relationship between Eurasian snow cover anomaly and wavetrain-like Eastern Atlantic teleconnection (EA) (Wallace and Gutzler 1981) (not shown), which seems less connected with this paper's theme. So we will not discuss this coupled pattern.

3. Evolution of atmospheric anomalies under typical coupled relationship between autumn Eurasian snow anomaly and wintertime NAO

The purpose of this paper is to give useful information on the evidently coupled connection between an early-seasonal snow cover anomaly and the DJF NAO. It is clearly seen in Fig. 1d that the values of SVD1 time series are lower than that of the NAO index, but at the same phase as the observed NAO index. The years when the NAO index and/or time series of SVD1 are nearly zero are excluded in the following discussion since no apparent NAO events and/or no linear contribution of snow anomaly to the NAO are presented in these years. In addition, the year 1999 is excluded too because of the opposite sign of snow EOF1 and SVD1. Therefore, we choose the special years, in which the coupled relationship between DJF NAO and SON Eurasian snow cover anomaly is well established, to study possible physical links between the two fields.

The linear correlation between DJF NAO and SON Eurasian snow indicates two dominantly categories over time: positive SON Eurasian snow cover anomalies and negative DJF NAO, hereafter termed as P-N, and correspondingly negative SON Eurasian snow cover anomalies and positive DJF NAO, denoted as N-P. The years 1976, 1977, 1978, 1984, 1985, 1995 and 2000 belong to the P-N category, and the years 1975, 1983, 1988, 1990, 1991, 1992 and 1994 to the N-P category. Because of significant leading relation between October Eurasian snow cover anomalies and DJF NAO (see Introduction), evolution of atmospheric anomalies from October to December will be studied for both categories using composite analysis with use of Student's t-test for evaluation of statistical significance.

3.1 P-N category

In Fig. 2 the green star marks the large area of positive snow anomaly over Siberia between 50°N and 65°N in October at a 95% confidence level. The extensive snow is associated with a simultaneous anomaly of large-scale atmospheric circulation. The warm ridge over the west coast of Eurasian continent at mid- and high latitudes is significantly stronger than normal (positive anomaly), so are also the cold troughs over west of Lake Baykal and northwestern Africa (negative anomalies). The strong meridional circulation structures at mid- and high latitudes over these regions guide the movement of the cold and warm air in lower troposphere and bring an extensive snow cover over Siberia (solid precipitation) at comparative high latitude.

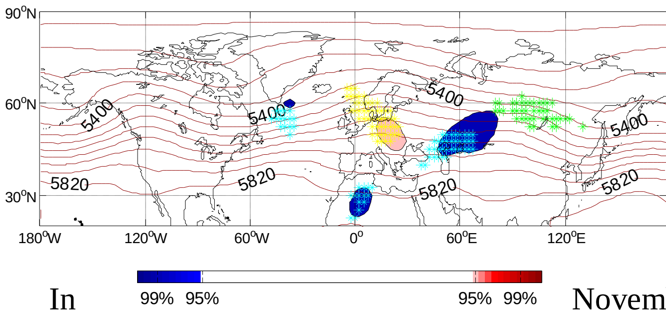

Figure 2. Climatological 500hPa height field in October. The contour interval is 60m. Shading areas indicate significant height anomaly at confident levels higher than 95% (red-positive anomaly, blue-negative anomaly). Green, blue and yellow star marks indicate significant anomalies for positive snow cover, negative 850hPa temperature and positive 850hPa temperature respectively at a 95% confident level.

In November, the high pressure ridge over North Atlantic is cold and weak (negative anomaly) (Fig. 3a), while the region at high latitude is mainly covered by a positive anomaly in 500hPa height. It implies a weak polar vortex, but statistically insignificant. This anomalous situation persists and strengthens in December (Fig. 3b). The significantly negative temperature anomaly over North Atlantic reduces the strength of the high pressure ridge further, while a significantly warm anomaly over Greenland and Arctic weakens the polar vortex. A typical negative AO/NAO pattern takes place.

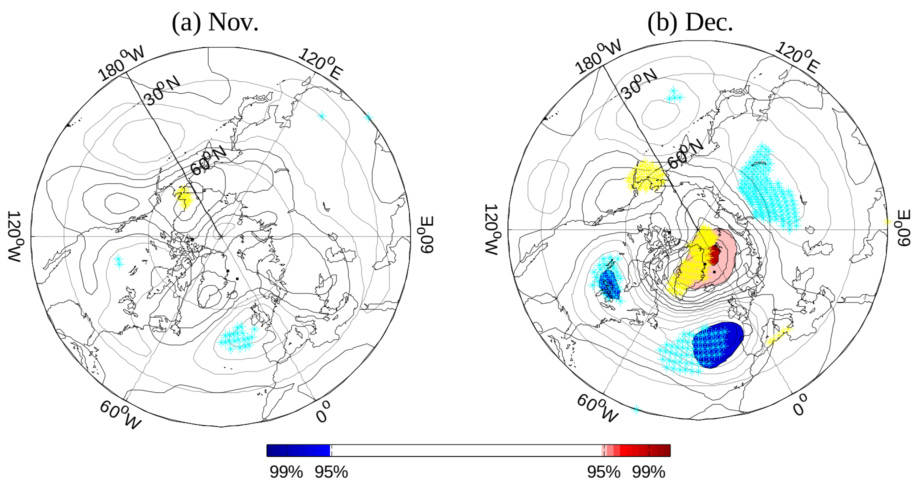

Figure 3. Monthly anomalous composite maps of 500hPa height field in November (a) and December (b) under P-N category, the contour interval is 15m. Shading areas indicate significant height anomaly at confident level higher than 95% (red-positive anomaly, blue-negative anomaly), blue and yellow star marks indicate negative and positive anomalies of 850hPa temperature respectively, significant at a 90% confident level.

From the atmospheric anomaly evolution under the P-N category it is clearly seen that the extensive Eurasian snow cover in October is driven by its upstream meridional atmospheric circulation at mid- and high latitudes. Since the significantly cold and weak ridge of subtropical North Atlantic persists from November to December, it is obvious that the nonlinear adjustment of the atmospheric circulation from the meridional circulation over North Atlantic and downstream regions in October to subsequent nearly circumpolar circulation anomaly may relate to negative NAO. The significant thermal effect (approximate -3.5°C anomaly) over Siberia in lower troposphere, which is associated with extensive autumn Eurasian snow cover, takes place in December (negative temperature anomaly in Fig. 3b). Therefore it seems that the atmospheric dynamical evolution in autumn, in years with extensive Eurasian snow early in the season, plays an important role for the DJF negative NAO.

3.2 N-P category

Fig. 4a shows the composite result of the height anomaly at 500hPa in October in years with less Eurasian snow cover. An atmospheric anomalous dipole over polar region and northeastern China and Mongolia, called Polar/Eurasia (PEA) teleconnection pattern from NOAA NCEP CPC website, takes place. This contributes to a deeper polar vortex than average over eastern part of Arctic and to a stronger high pressure ridge of Lake Baykal. This strong ridge plays a key role in preventing the boreal cold air from moving southward into Eastern Europe, resulting in less Eurasian snow. The significant warming effect (2°C anomaly) in lower troposphere over Siberia, associated with less snow, strengthens the local high pressure ridge.

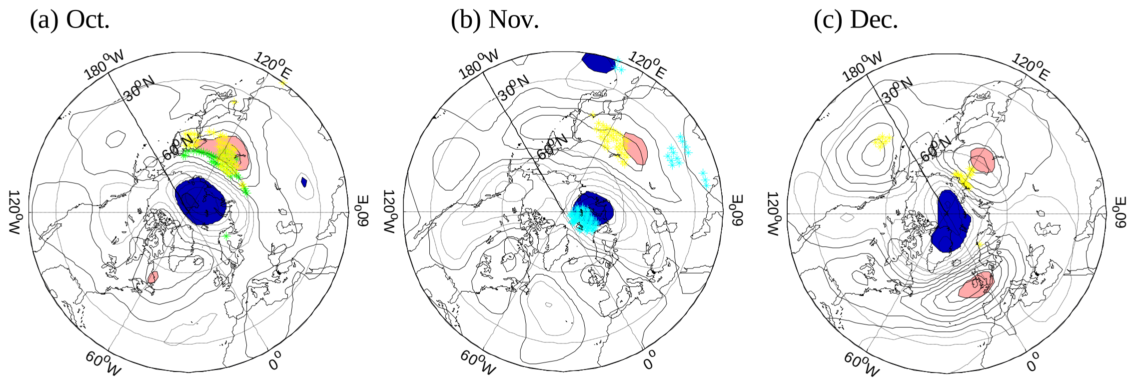

Figure 4. Monthly anomalous composite maps of 500hPa height field in October (a), November (b) and December (c) under N-P category. The contour interval is 15m. Shading areas indicate significant height anomaly at confident levels higher than 95% (red-positive anomaly, blue-negative anomaly), green, blue and yellow star marks indicate negative snow cover anomaly, negative and positive 850hPa temperature anomaly respectively, significant at a 95% confident level.

The PEA pattern persists in November (Fig. 4b), and the significant warming over Siberia (2°C anomaly) maintains the strong ridge of Lake Baykal. In the mean time a cooling over Arctic (near -4°C), significant at a 95% confidence level, indicates a strengthening of the polar vortex. Then in December the polar vortex develops even more and expands northward, covering the whole Arctic. Simultaneously a significant positive height anomaly occurs over North Atlantic (Fig. 4c), and the typical positive AO/NAO pattern establishes.

In our study, the significant distribution of temperature anomaly at 2m height under less October Eurasian snow cover is very similar to that at 850hPa height (not shown). The warm anomalous center over Siberia and Mongolia in lower troposphere persists from October to November (Fig. 4). This implies that the consequence of surface warming due to less snow through snow-albedo feedback could be a diabatic heating source to the lower atmosphere (Yeh et al. 1983; Kumar and Yang 2003). Studies have found that PEA teleconnection is closely related to Eurasian surface air temperature from autumn to spring (KOIDE and KODERA 1999). Hence it is interesting to see the consistency that the positive phase of PEA pattern appears throughout the period investigated.

The positive phase of PEA pattern reflects an enhanced circumpolar vortex, especially in December the polar vortex expands northward and even covers Greenland. Due to the equivalent-barotropic structure of atmospheric circulation over North Atlantic sector (Thompson and Wallace, 1998), concurrent variation in the strength of the Icelandic Low and the Azores High takes place, resulting in the occurrence of the positive NAO.

4. Conclusions

This study focuses on the linear coupled relationship between autumn Eurasian snow cover extent and wintertime NAO based on the leading modes of SVD analysis on the standardized monthly mean ECMWF SLP and satellite-obtained mean snow cover fields. The extensive (deficient) snow cover over large areas from northern Europe to Siberia in October heralds the negative (positive) DJF NAO. We provide evidence for possible physical linkage between early snow and wintertime NAO. It is found that the extensive Eurasian snow cover in October is driven by its upstream meridional atmospheric circulation at mid- and high latitudes. The evident thermal cooling effect associated with early-seasonal extensive Eurasian snow takes place in December. On the other hand, the surface warming over Siberia in years with less snow through snow-albedo feedback acts as a diabatic heating source to the lower troposphere from October to November.

The modulation effect in years with extensive Eurasian snow on the wintertime negative NAO has been suggested by previous studies (Watanabe and Nitta, 1998; Gong et al., 2003). The albedo effect is not the predominant thermodynamic mechanism behind a positive Siberia snow anomaly (Gong et al. 2004). From our study it is clearly seen that the negative temperature anomaly over Siberia is not significantly strong before December under extensive Eurasian snow cover, and the ridge of subtropical North Atlantic changes from stronger in October to colder and weaker than normal from November to December. This implies that the dynamical atmospheric evolution in mid- and late autumn and early winter plays an important role for the DJF negative NAO.

Gong et al. (2004) have found that realistic Siberia snow cover and snow depth anomalies in early season are required to produce a local temperature response which is strong enough to distinctly modulate the winter AO/NAO. In years with less snow in autumn over Eurasia, we found that the surface warming over Siberia is strong enough to heat the atmosphere in lower troposphere (2°C anomaly on 850hPa). Moreover, associated with heating effect, the positive phase of Polar/Eurasia teleconnection persists. Therefore the possible contribution of autumn deficient Eurasian snow to typical positive AO/NAO is mainly through the persistent thermal heating caused by snow variation and the enhanced effect of PEA pattern on the circumpolar vortex.

In this paper it is found that the main dynamical atmospheric response from an extensive autumn Siberia snow cover to a wintertime low NAO index is quite different in character from the response in years with less snow leading to a relative high NAO index. The causality for such different response from snow forcing derived from statistical analysis need further theoretical considerations. An interesting aspect of our study is that the significantly higher correlation found for SVD1 indicates some skill in seasonal prediction for wintertime NAO based on large anomalies in autumn Eurasian snow cover.

Acknowledgments. This investigation was supported by Research Council of Norway through MACESIZ project. We thank Prof. Sigbjørn Grønås for helpful comments and suggestions.

Reference:

Armstrong, R.L. and M.J. Brodzik, Northern Hemisphere EASE-Grid Weekly Snow Cover and Sea Ice Extent Version 2, *Boulder, CO, USA: National Snow and Ice Data Center*, CDROM, 2002

Bader J. and M. Latif, The impact of decadal-scale Indian Ocean sea surface temperature anomalies on Sahelian rainfall and the North Atlantic Oscillation, *Geophys. Res. Lett.*, *30*(22), 2169, 2003.

Bretherton, C. S., C. Smith and J. M. Wallace, An intercomparison of methods for finding coupled patterns in climate data, *J. Clim.*, 5**,** 541-560, 1992.

Cohen J. and D. Entekhabi, Eurasian snow cover variability and North Hemisphere climate variability, *Geophys. Res. Lett.*, 26(3), 345-348, 1999.

Cohen J., D. Salstein and K. Saito, A dynamical framework to understand and predict the major Norther Hemisphere mode, *Geophys. Res. Lett.*, 29(10), 1412, 2002.

European Centre for Medium-Range Weather Forecasts, The ERA-40 Archive. Reading, ECMWF, 40, 2002.

Feldstein S. B., The recent trend and variance increase of the annular mode, *J. Clim*., 15, 88-94, 2002.

Gong G., D. Entekhabi, and J. Cohen, Modeled Northern Hemisphere winter climate response to realistic Siberian snow amomalies, *J. Clim*., 16, 3917-3931, 2003.

Gong G., D. Entekhabi, and J. Cohen, 2004: Sensitiveity of atmospheric response to modeled snow anomaly characteristics. *J. Geophys. Res.*, 109, D06107

Hoerling, M. P., Hurrell, J. W., and T. Y.Xu, Tropical origins for recent North Atlantic climate change, *Science*, 292, 90-92, 2001 .

Hoerling, M.P., J.W. Hurrell, T. Xu, G.T.Bates, and A.S. Phillips, Twentieth Century North Atlantic Climate Change. Part II: Understanding the Effect of Indian Ocean Warming. *Clim. Dyn*, 23, 391－405, 2004.

Hurrell, J. W., Decadal trends in the North Atlantic Oscillation: regional temperatures and precipitation, *Science*, 269, 676-679, 1995.

Koide H. and K. Kodera, A SVD analysis between the winter NH 500-hPa height and surface temperature fields, *J. Meteorol. Soc. Jpn.*, 77, 47-61, 1999.

Kumar A. and F. Yang, Comparative influence of snow and SST variability on extratropical climate in Northern winter, *J. Clim.*, 16, 2248-2261, 2003.

Paeth H., M. Latif and A. Hense, Global SST influence on twentieth century NAO variability, *Clim. Dyn.*, 21, 63-75, 2003.

Ringler, T. D. and K. H. Cook, Understanding the seasonality of orographically forced stationary waves: Interaction between mechanical and thermal forcing, *J. Atmos. Sci.*, 56, 1154-1174, 1999.

Rodwell, M. J. and B. J. Hoskins, Monsoons and the dynamics of deserts, Quart. J. Roy. Meteor. Soc., 122, 1385–1404, 1996.

Rodwell M. J., D. P. Rowell and C. K, Folland, Oceanic forcing of the wintertime North Atlantic Oscillation and European climate. *Narure*, 398, 320-323, 1999 .

Rodwell M. J. and C. K, Folland, Atlantic air-sea interaction and seasonal predictability. Q. J. R. Met. Soc., 128, 1413-1443, 2002.

Saito K., J. Cohen and D. Entekhabi, Evolution of atmospheric response to early-season Eurasian snow cover anomalies, *Mon.Wea.Rev*., 129, 2746-2760, 2001.

Saito K. and J. Cohen, The potential role of snow cover in forcing interannual variability of the major Northern Hemisphere mode, *Geophys. Res. Lett*, 30(6), 1302, 2003.

Tompson D. W. J. and J. M. Wallace, The Arctic Oscillation signature in the wintertime geopotential height and temperature fields, *Geophys. Res. Lett.*, 25, 1297-1300, 1998.

Uppala, S. and Coauthors, The ERA-40 re-analysis, *Q. J. R. Meteorol. Soc.*, 131, 2961–3012, 2005.

Wallace J. M. and D. S. Gutzler, Teleconnection in the geopotential height field during the Northern Hemisphere winter, *Mon. Wea. Rev*., 109, 784-812, 1981.

Wallace, J. M., C. Smith and C. S. Bretherton, 1992: Singular value decomposition of wintertime sea surface temperature and 500-mb height anomalies. *J. Clim.*, 5, 561-576.

Walter K. and H.F. Graf, On the changing nature of the regional connection between the North Atlantic Oscillation and sea surface temperature, *J. Geophy. Res.*, 107, D17, 4338, 2002.

Watanabe M. and T. Nitta, Relative impacts of snow and sea surface temperature anomalies on an extreme phase in the winter atmospheric circulation, *J. Clim.*, 11, 2837-2857, 1998.

Yeh T.-C. , R.T. Wetherald and S. Manabe, 1983: A Model Study of the Short-Term Climatic and Hydrologic Effects of Sudden Snow-Cover Removal. *Mon.Wea.Rev*., 111, 1013–1024.