NAO impact towards the springtime snow disappearance in the western Eurasian continent

M. E. Hori
Institute of Geoscience, University of Tsukuba, Ibaraki, Japan

T. Yasunari
Hydrospheric Atmospheric Research Center, Nagoya University, Aichi, Japan

Received 6 July 2003; revised 2 August 2003; accepted 2 September 2003; published 3 October 2003.

[1] The atmospheric control over the disappearance of snow in western Eurasia and its relationship to the North Atlantic Oscillation (NAO) are investigated. While the NAO has a strong effect on the lower tropospheric temperature in the region, its ability to force control over the snow extent is confined to western Europe in January and February. Successive composite differences and heat budget analysis reveals a northeastward propagation of snow cover anomaly maintained by an anomalous zonal heat advection related to the NAO from January to March, and the intrusion of warm climatological southerly maintaining the snow anomaly until April. *INDEX TERMS:* 3309 Meteorology and Atmospheric Dynamics: Climatology (1620); 3319 Meteorology and Atmospheric Dynamics: General circulation; 3322 Meteorology and Atmospheric Dynamics: Land/atmosphere interactions. **Citation:** Hori, M. E., and T. Yasunari, NAO impact towards the springtime snow disappearance in the western Eurasian continent, *Geophys. Res. Lett.*, *30*(19), 1977, doi:10.1029/2003GL018103, 2003.

1. Introduction

[2] Interannual variations and the seasonal predictability of monsoon activity has been of great interest to researchers and local population of the monsoon region alike, due to its large social-economical impact.

[3] While many explanations regarding the source of monsoon interannual variability have been given, it is well recognized that the wintertime snow cover over the Eurasian continent has a strong negative correlation with the subsequent Indian monsoon rainfall (IMR) thus being a good predictor [*Hahn and Shukla*, 1976]. *Bamzai and Shukla* [1999] showed that such correlation is created primarily by the variations in snow extent over western Europe during winter (Dec.–Mar.), and in eastern Europe during spring (Apr.–May). This negative correlation has grown stronger since the 1970s and can be related to the wintertime lower tropospheric temperature variability in the region [*Chang et al.*, 2001; *Liu and Yanai*, 2002].

[4] On the other hand, *Clark et al.* [1999] have shown that the snow extent in this particular region is subject to strong influence by the North Atlantic Oscillation (NAO). The NAO is characterized by a strong barotropic dipole structure of pressure/height centered over the Atlantic basin. Particular importance of NAO is its dominant strength in the wintertime atmospheric variability and its relation to the recent positive temperature trend in the Eurasian continent [*Hurrell*, 1996].

[5] *Robock et al.* [2003] have shown that correlations between DJF-averaged NAO and two-month running averaged snow extent produces a seasonal change similar to that reported by *Bamzai and Shukla* [1999]. Their result shows a negative snow cover anomaly which progresses eastward from winter to spring. *Saito and Cohen* [2003] have also shown a lag-lead correlation between the AO/NAO variability and Eurasian snow cover which shows an atmosphere leading correlation originating in January persisting through to the following spring.

[6] However, it should be noted that atmospheric variability such as the NAO has a relatively short memory in monthly time scale, and since snow disappears within 2–3 weeks [*Ueda et al.*, 2003], the timing is crucial of when and where such atmospheric signals can be imprinted on the surface snow cover.

[7] The objective of this study is to investigate the source of the strong seasonal correlation between the NAO and the snow extent, and how an anomalous snow extent persists through the following months.

2. Data

[8] Two types of snow cover data are used throughout this research. The NOAA/NESDIS & CPC northern hemisphere snow cover data is used for the period of 1973–2002 following the methods of *Gutzler and Rosen* [1992].

[9] In addition, we use the 10-day mean snow depth data from the Former Soviet Union Hydrological Snow Surveys [*Krenke*, 1998]. 705 stations from 20° through 60°E and 35° through 70°N are used from 1966 to 1990. Data after 1990 is not used due to a reduction in the number of stations.

[10] Also, daily averaged geopotential height and temperature field and omega velocity from the NCEP/NCAR reanalysis [*Kalnay et al.*, 1996] are used for 1973–2002.

[11] The NAO index based on a long-term normalized pressure difference between Gibraltar and Iceland provided by the Climate Research Unit is also used [*Jones et al.*, 1997; *Osborn et al.*, 1999].

3. Results

[12] To illustrate the impact of the NAO towards snow extent, we begin by presenting the composite difference of

0094-8276/03/2003GL018103$05.00

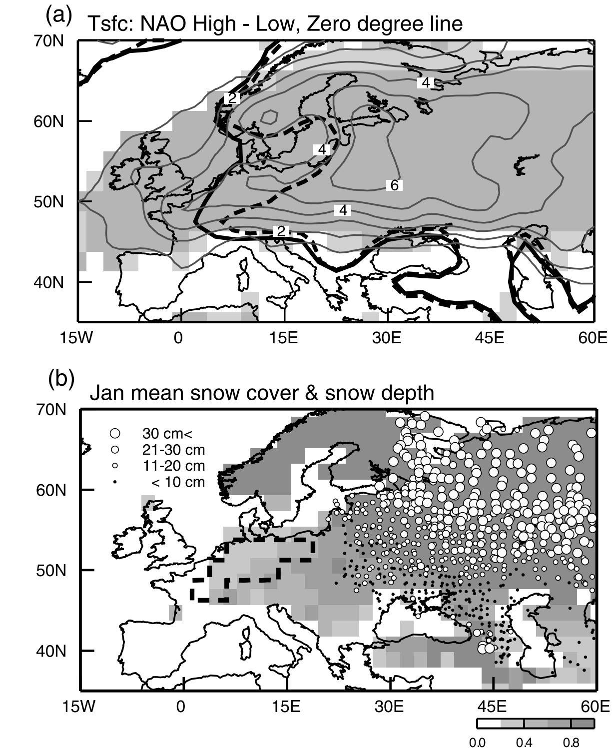

Figure 1. (a) Composite difference of the January surface temperature for 14 positive NAO years and 7 negative years. Units in °C. The region of light (dark) shades corresponds to 95% (99%) confidence level. Thick dashed (thick solid) line denotes the zero degree line for positive (negative) NAO composites respectively. (b) Climatological January snow cover frequency (shading) and snow depth (circles).

January surface temperature for 14 years with January NAO index exceeding 1.0 and 7 years below the −1.0 threshold in (Figure 1a). Light and dark shading corresponds to differences occurring less than 5% and 1% by chance, based on Monte-Carlo test of 1000 random composite differences. Also, the composite zero-degree line for high (low) NAO years is shown in thick dashed (thick solid) line.

[13] Significant positive temperature difference can be seen across western Eurasia, with the center of strong warming located around 30°E, 55°N. However, an area with sub-zero temperature in the low NAO years and above-zero temperature in the high NAO years is confined to an area around 5–15°E, 48–52°N.

[14] In this region, the persistence of snow and precipitation phase will be greatly affected by the NAO signature, whereas an above-zero temperature accompanying high NAO events favor a short persistent snow cover and precipitation consisting mainly of rain instead of snow and vice versa for the low NAO events [*Clark et al.*, 1999]. Presence of snow cover in this region may also have a local effect towards the surface temperature favoring such persistence of snow [*Walsh et al.*, 1985; *Shinoda et al.*, 2001]. Whereas in the warming center, the climatological feature reveals a temperature below −10°C and not even a strong warming accompanying a positive NAO phase could overcome the zero-degree threshold. We define this region excluding the ocean grids as the NAO Impact Region (NIR) where the atmosphere controls the persistence of snow.

Table 1. Cross Correlation of Monthly NAO Index and NIR-Averaged Snow Cover for December Through February

	NAO Index		
NIR Snow	Dec	Jan	Feb
Dec	−0.37	−0.16	0.10
Jan	−0.17	**−0.69**	−0.37
Feb	0.11	−0.32	**−0.79**

Bold types denotes a statically 95% significant correlation based on a two-tailed Student's t-test. All timeseries are normalized prior to taking the correlation.

[15] The frequency and depth of snow cover for the climatological January from 1972–2002 are shown in (Figure 1b). NIR, denoted by an area surrounded by a thick dashed line, lies mostly in an area where large interannual snow cover variability can be seen.

[16] An equivalent figure for December (not shown) shows that the NIR is primarily in the same western European region but with a reduced area due to smaller influence of the NAO on the surface temperature than that in January. The equivalent figure for February (not shown) shows that the NIR shares most areas with that of January with similar underlying snow variability. This resemblance is due to similarity in climatological features between January and February.

[17] Table 1 shows the lag-lead correlation for the NAO index and NIR-averaged snow cover. Simultaneous correlation in January and February is significant which is consistent with the findings of *Saito and Cohen* [2003]. Considering the weak correlation between January NAO and February NAO (0.12), atmospheric control over NIR snow cover can occur in either January or February.

[18] In this study, the following lag composite analysis is based on January to study a long-lasting snow cover anomaly.

[19] A timeseries of the January NAO index and NIR-averaged snow cover frequency are shown in Figure 2. It can be seen that the strength of correlation shown in Table 1 comes mostly from the late 1970s and 1980s where negative NAO was accompanied by a large snow cover over the NIR, and a deficit in snow cover during the early 1990s when the NAO was positive.

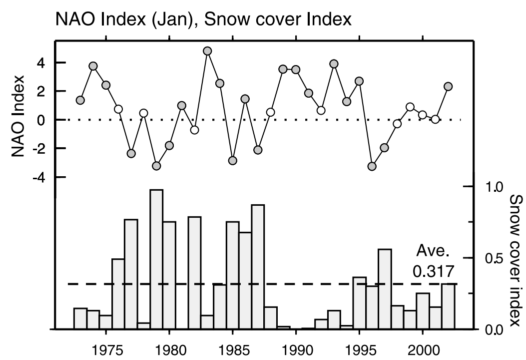

Figure 2. Timeseries of the January NAO index and snow cover frequency averaged over the NIR. The positive and negative extreme NAO year are shown with shaded circles. Thick dashed line denotes the climatological snow cover over the NIR.

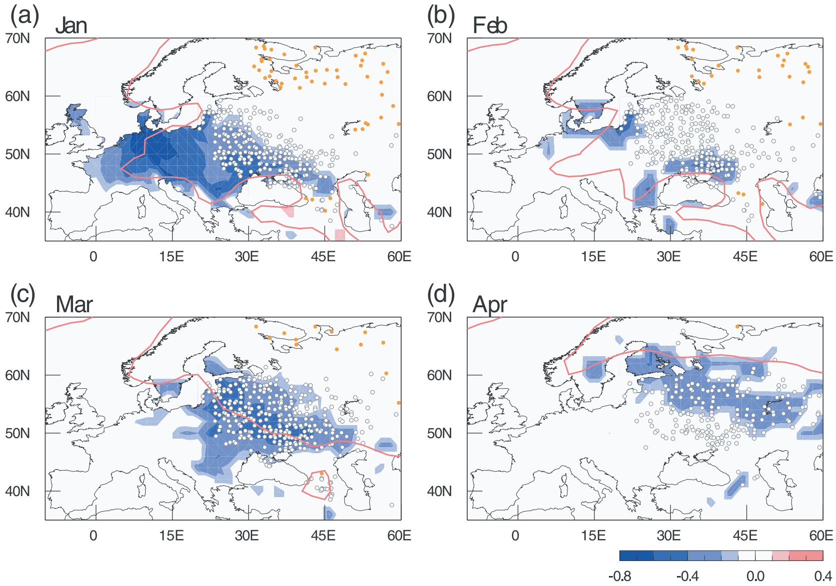

Figure 3. (a) Composite difference of snow cover frequency (shading) and snow depth (circles) for 14 positive NAO years and 7 negative NAO years. Plots were made only for differences exceeding the 95% confidence level. An orange (white) circle is plotted for significant positive (negative) difference in snow depth. Thick red line shows the climatological zero-degree line. (b) same as in (a) except for a lag composite difference in following February. (c) same as in (b) except for following March. (d) same as in (b) except for following April.

[20] To investigate the persistence of NAO signature on the snow cover, a lag composite difference based on January with high or low NAO is shown in Figure 3. Here, the shading corresponds to a rate of frequency change in satellite detected snow cover and circles the change in snow depth. In January, a strong negative difference in snow cover and snow depth centered around the NIR extends well eastward of the climatological zero-degree line, which marks the disappearance of snow in the positive NAO. In the subsequent February, significant snow differences propagating northeastwards can be seen in the snow depth, but not as clear in the satellite-derived snow cover which represents a change in larger area of snow.

[21] The progression of this signal can be seen in the following March and April and was consistent with the findings of *Bamzai and Shukla* [1999] and *Robock et al.* [2003]. It should be noted that there is no difference in the snow depth during January in the region where negative signals will be seen in April. This shows that the northeastward propagation of snow extent anomaly is sensitive to the January snow extent over the NIR region and not the excess or deficit in snow mass in the region.

[22] A quantitative heat budget analysis was conducted to explore the maintenance of such seasonal progression following the methods used by *Tanaka and Milkovich* [1990].

[23] The thermal advection, adiabatic expansion or contraction, and residual term (mainly radiation) were calculated using a daily atmospheric dataset, and were is mass integrated for 1000hPa through 100hPa and monthly averaged.

[24] The climatological surface wind and lag composite difference for the advective heating term based on high or low January NAO years is shown in Figure 4. The adiabatic term was found to be small throughout January to April

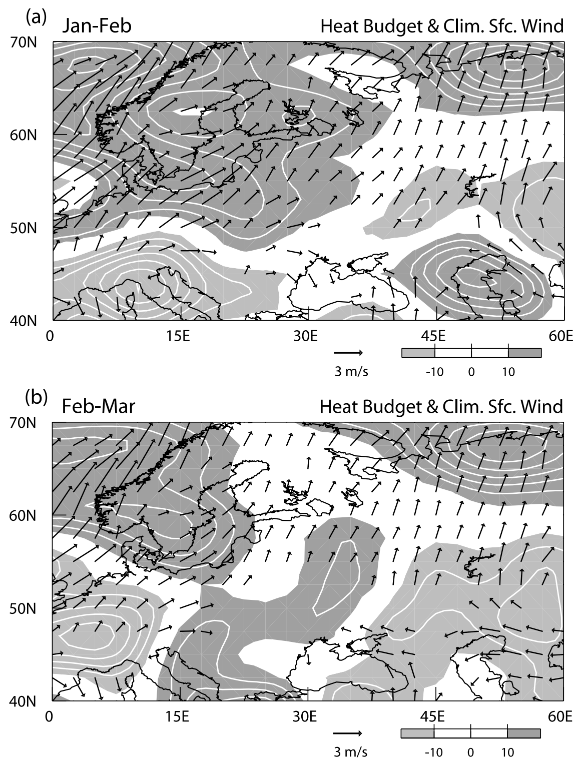

Figure 4. (a) A composite difference of January through February heat budget analysis. The dark (light) shading with white contours corresponds to positive (negative) advection. Contour interval is 20 W/m^2. Vectors denote the climatological surface wind. Grids with wind speed exceeding 1 m/s are plotted. (b) Same as in (a) but for February through March.

composites, and is not shown on the figure. During January through February, advective heating due to NAO induced westerly anomaly is present in western Europe, corresponding to the snow cover difference in (Figures 3a and 3b).

[25] While anomalous westerlies are weak during March, the onset of warm climatological southerly over central Europe around 35°E, 55°N [*Ueda et al.*, 2003], gives a strong (weak) southwesterly feature above the region of snow cover anomaly in (Figure 3c) on high (low) NAO cases, which corresponds to the heat advection anomaly feature in (Figure 4b). Heat advection anomaly is weak during April (not shown), and the snow cover anomaly signals caused in the previous months are carried mainly by the intrusion of warm southerly.

4. Discussion and Conclusions

[26] Lag composite difference has been used to demonstrate the atmospheric regulation of snow extent over the western Eurasia.

[27] The strong coupling of the NAO and snow extent during January and February is confined to a relatively small area centered in Western Europe. This region is closely related to the area used in the study of *Chang et al.* [2001], where a recent strong correlation between the IMR and the temperature of western Eurasia was noted.

[28] Strong persistence of wintertime NAO signature into the following season was responsible for the northeastward propagation of snow cover anomaly from January to April, which was found in both a satellite-derived snow cover chart and in situ measurements of snow depth. Heat budget analysis reveals that anomalous zonal heat advection centered n western Europe is mostly responsible for maintaining the snow cover anomaly from January to February. While such anomalous heat advection is weak during March, the onset of climatological warm southerly is sufficient in carrying the snow cover anomaly from March to April.

[29] While the result strongly supports the findings of *Robock et al.* [2003], it was also found that the progress of this signal is not a result of continental scale excess/deficit in snow mass as hypothesized in several model experiments [*Barnett et al.*, 1989; *Yasunari et al.*, 1991; *Vernekar et al.*, 1995].

[30] It should be noted that such a seasonal march in snow cover can occur with regards to the January NAO signature alone, and does not require the following Indian monsoon rainfall to be in a particular state. The signal of disappearance of snow in western Eurasia rapidly diminishes by April [*Ueda et al.*, 2003], and soil moisture signature is probably not an efficient carrier of climatic signals for the following season [*Shinoda*, 2001].

[31] However, snow extent may have a vital role in regulating the circulation pattern which gives background for the monsoon [*Yasunari and Seki*, 1992]; thus further analysis of the effect of an anomalous snow extent on the monsoon is required.

[32] **Acknowledgments.** The authors would like to thank Dr. Fujio Kimura and Dr. Hiroaki Ueda for their helpful discussions. The reviewers provided valuable comments and suggestions. Figures in this article are made by the Generic Mapping Tools [*Wessel and Smith*, 1991].

References

Bamzai, A. S., and J. Shukla, Relation between Eurasian snow cover, snow depth, and the Indian summer monsoon: An observational study, *J. Clim.*, *12*, 3117–3132, 1999.

Barnett, T. P., L. Dümenil, U. Schlese, F. Roeckner, and M. Latif, The effect of Eurasian snow cover on regional and global climate variations, *J. Atmos. Sci.*, *46*, 661–685, 1989.

Chang, C.-P., P. Harr, and J. Ju, Possible roles of Atlantic circulations on the weakening Indian monsoon rainfall-ENSO relationship, *J. Clim.*, *14*, 2376–2380, 2001.

Clark, M. P., M. C. Serreze, and D. A. Robinson, Atmospheric controls on Eurasian snow extent, *Int. J. Climatol.*, *19*, 27–40, 1999.

Gutzler, D. S., and R. D. Rosen, Interannual variability of wintertime snow cover across the northern hemisphere, *J. Clim.*, *5*, 1441–1447, 1992.

Hahn, D. J., and J. Shukla, An apparent relationship between Eurasian snow cover and Indian monsoon rainfall, *J. Atmos. Sci.*, *33*, 2461–2462, 1976.

Hurrell, J. W., Influence of variations in extratropical wintertime teleconnections on Northern Hemisphere temperature, *Geophys. Res. Lett.*, *23*, 665–668, 1996.

Jones, P. D., T. Jónsson, and D. Wheeler, Extension to the North Atlantic Oscillation using early instrumental pressure observations from Gibraltar and South-West Iceland, *Int. J. Climatol.*, *17*, 1433–1450, 1997.

Kalnay, E., et al., The NCEP/NCAR 40-Year reanalysis project, *Bull. Am. Meteorol. Soc.*, *77*, 437–471, 1996.

Krenke, A., Former Soviet Union hydrological snow surveys, edited by NSIDC, http://nsidc.org/data/docs/noaa/g01170fsu_snow/index.html, World Data Cent. for Glaciol. [Snow and Ice], Boulder, Colo, Digital media, 1998, updated 2003.

Liu, X., and M. Yanai, Influence of Eurasian spring snow cover on Asian summer rainfall, *Int. J. Climatol.*, *22*, 1075–1089, 2002.

Osborn, T. J., K. R. Briffa, S. F. B. Tett, P. D. Jones, and R. M. Trigo, Evaluation of the North Atlantic Oscillation as simulated by a coupled climate model, *Clim. Dyn.*, *15*, 685–702, 1999.

Robock, A., et al., Land surface conditions over Eurasia and Indian summer monsoon rainfall, *J. Geophys. Res.*, *108*, 4131–4147, 2003.

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*, 1594–1597, 2003.

Shinoda, M., Climate memory of snow mass as soil moisture over central Eurasia, *J. Geophys. Res.*, *104*, 33,393–33,403, 2001.

Shinoda, M., M. Shinoda, and H. Utsugi, Spring snow-disappearance timing and its possible influence on temperature fields over central Eurasia, *J. Meteorol. Soc. Japan*, *79*, 37–59, 2001.

Tanaka, H. L., and M. F. Milkovich, A heat budget analysis of the polar troposphere in and around alaska during the abnormal winter of 1988/89, *Mon. Wea. Rev.*, *118*, 1628–1639, 1990.

Ueda, H., M. Shinoda, and H. Kamahori, Spring northward retreat of Eurasian snow cover relevant to seasonal and interannual variations of atmospheric circulation, *Int. J. Climatol.*, in press, 2003.

Vernekar, A. D., J. Zhou, and J. Shukla, The effect of Eurasian snow cover on the Indian monsoon, *J. Clim.*, *8*, 248–266, 1995.

Walsh, J. E., W. H. Jasperson, and B. Ross, Influences of snow cover and soil moisture on monthly air temperature, *Mon. Wea. Rev.*, *113*, 756–768, 1985.

Wessel, P., and W. H. F. Smith, Free software helps map and display data, *Eos Trans. AGU*, *441*, 445–446, 1991.

Yasunari, T., A. Kitoh, and T. Tokioka, Local and remote responses to excessive snow mass over Eurasia appearing in the northern spring and summer climate—A study with the MRI-GCM-, *J. Meteorol. Soc. Japan*, *69*, 473–487, 1991.

Yasunari, T., and Y. Seki, Role of the Asian monsoon on the interannual variability of the global climate system, *J. Meteorol. Soc. Japan*, *70*, 177–189, 1992.

M. E. Hori, Institute of Geoscience, University of Tsukuba, 1-1-1 Tennoudai, Tsukuba City, Ibaraki, 305-0001, Japan. (mhori@kankyo.envr.tsukuba.ac.jp)

T. Yasunari, Hydrospheric Atmospheric Research Center, Nagoya University, Chikusa-ku, Nagoya City, Aichi, 464-8601, Japan. (yasunari@ihas.nagoya-u.ac.jp)