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The Role of snow cover in the Northern Hemisphere winter to summer transition

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[1] This paper examines the role of North Hemisphere snow cover in the linkage between the winter North Atlantic Oscillation (NAO) and the summer Northern Annular Mode (NAM). This transition is partially supported by the persistence of the NAO-induced snow cover anomalies and the asymmetric thermal distribution induced by summer snow cover. We define an index of subpolar temperature difference which links winter NAO with the subsequent summer NAM. The index is also significant in the linkage between summer and winter climates and can be used as an useful predictor of the upcoming winter NAO. **Citation:** Barriopedro, D., R. García-Herrera, and E. Hernández (2006), The Role of snow cover in the Northern Hemisphere winter to summer transition, *Geophys. Res. Lett.*, *33*, L14708, doi:10.1029/2006GL025763.

1. Introduction

[2] The dominant pattern of Northern Hemisphere (NH) climate variability is characterized by a barotropic structure with simultaneous and opposite-signed oscillations of atmospheric mass between high and midlatitudes. Regionally it is commonly referred to as the North Atlantic Oscillation (NAO) [*Wallace and Gutzler*, 1981] and hemispherically as the Arctic Oscillation (AO)/Northern Annular Mode (conventional NAM hereafter) [*Thompson and Wallace*, 2000] or the seasonally varying Northern Annular Mode (NAM hereafter) [*Ogi et al.*, 2004]. Since winter shows larger atmospheric variability, the structure of the conventional NAM/AO mostly reflects the winter pattern. However, the NAM shows a large spatial variation, with its meridional structure varying with season.

[3] Even when the NAO is considered an internal mode of variability the recent increase of interannual variance and its upward trend are in excess of that expected from internal fluctuations of the atmosphere, which has been partially attributed to a coupling with other components of the climatic system [*Feldstein*, 2002]. The persistence of these boundary conditions of long-term memory can make seasonal atmospheric conditions predictable at a later time. Attempts to predict the winter NAO from general circulation models (GCMs) have yielded little skill to date, but recent empirical studies have reported significant lagged links between winter NAO and slow-varying variables in the preceding seasons such as JJASO North Atlantic sea surface temperatures [*Saunders and Qian*, 2002] or snow cover in March to October [*Bojariu and Gimeno*, 2003] and summer [*Saunders et al.*, 2003].

[4] On the other hand, winter NAO has been identified as a potential forcing of the subsequent spring [*Hori and Yasunari*, 2003] and summer [*Ogi et al.*, 2003] snow cover and atmospheric circulation. The summertime NAO signal is annular but its meridional scale is much smaller than the winter annular mode. The winter and the following summer NAM are also significantly correlated at 99%, the linking pattern being quite similar to the summer NAM. This winter-to-summer linkage has been interpreted as a preferred transition from one polarity of the winter NAO/NAM to the same polarity of the summer NAM [*Ogi et al.*, 2004]. However, the underlying mechanisms involving the continuation of the same phase are not completely understood.

[5] This paper examines the role of NH snow cover in supporting the winter to summer linkage. It is shown that the preferred winter NAO-to-summer NAM transition is favored by the subpolar temperature difference induced by snow cover, which simultaneously arises as an useful predictor of the next winter NAO.

2. Data

[6] 31 years (1972–2002) of monthly snow data were obtained from NOAA visible satellite charts in an 89x89 NH matrix, where each element represents the percentage of snow-covered surface of each cell. Monthly series of area of snow extent for Eurasia, North America (without Greenland) and NH were also used. Both data sets were provided by the Rutgers University Climate Lab (RUCL) [*Robinson et al.*, 1993].

[7] Additionally, monthly 2-m air temperature and 1000–200 hPa vertically integrated fields of geopotential height and zonal wind were derived from the ERA-40 reanalysis data set over a 2.5° latitude by 2.5° longitude NH grid [*Simmons and Gibson*, 2000].

[8] Since the winter to summer linkage is well sustained between winter and summer months [*Ogi et al.*, 2004] here we use the standard four-season definition (DJF, MAM, JJA and SON). We use the Hurrell NAO index, defined as the standardized mean sea-level pressure difference between Iceland and Ponta Delgada in the Azores [*Hurrell*, 1995] to characterize the winter dominant mode. Other NAO indices as those defined by *Jones et al.* [1997] or by the Climate Prediction Center (CPC) of the NOAA [www.cpc.ncep.noaa.gov/] did not show significant differences. As the summer NAM captures better the summer atmospheric variability, the NAM index is used as the dominant mode in summer [http://poplar.ees.hokudai.ac.jp/svnam/]. It is defined as the first mode in EOF analysis of monthly and zonally averaged geopotential height field from 1000 hPa to

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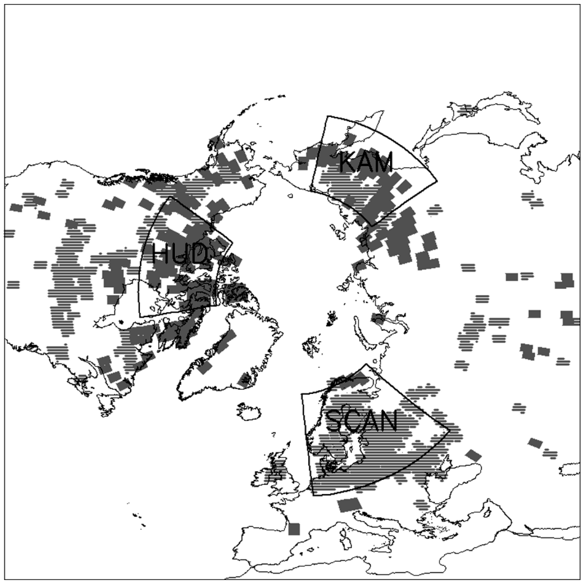

Figure 1. The winter-to-summer linkage. Correlation coefficient map between winter NAO and snow covered cells in spring (dashed cells) and June–July (shaded cells). Only significant negative correlation coefficients up to $p < 0.1$ significance level are shown. Significant correlated regions are marked with black solid lines and referred to as Scandinavia (SCAN), Kamchatka (KAM) and Hudson's Bay (HUD).

200 hPa and poleward of 40°N for each individual calendar month [*Ogi et al.*, 2004]. All series were detrended in order to minimize the influence of the temporal trends in the correlation and composites analyses.

3. Results

[9] The relationship between winter NAO and the upcoming and previous NH snow cover is dominated by two kind of linkages (not shown). The first one links winter NAO with the contemporaneous winter (Pearson's correlation coefficient $r = -0.43$) and subsequent spring ($r = -0.41$) and summer ($r = -0.43$) snow cover ($p < 0.05$). The other one implies a snow-leading relationship, with positive summer snow cover anomalies preceding a negative winter NAO-like pattern ($r = -0.57$) ($p < 0.01$). These teleconnections confirm the role of snow cover in winter NAO predictability and identify snow cover as a feasible candidate to memorize and support the winter NAO-to-summer NAM transition.

[10] The winter NAO impact on monthly snow cover is especially evident over Eurasia (North America) in the spring months (June–July) (Figure 1). As the climatological snow melting in the northern regions of North America is delayed relative to other regions of Eurasia, both linkages may simply represent the inter-continental propagation of the snow cover anomalies. In fact, the snow cover response in Eurasia starts to be significant two or three months earlier than in North America (not shown). Furthermore, JJ snow anomalies over the Kamchatka Peninsula (KAM) and the Hudson's Bay (HUD) are preceded by snow fluctuations in spring over the Scandinavian sector (SCAN) with correlation coefficients of 0.61 and 0.55 ($p < 0.01$), implying a remote teleconnection between snow covered areas of Eurasia and North America.

[11] Successive lag-correlation maps between winter NAO and snow cover reveal that the snow cover response in winter is essentially restricted to central Europe (not shown) while the NAO control in northern Eurasia, where winter temperatures are below freezing, exerts little influence on snow extent [*Clark et al.*, 1999]. In spring the snow cover signal progresses toward the SCAN sector ($r = -0.49$, $p < 0.01$) (see Figure 1). The Eurasian snow cover response has been attributed to a NAO influence in the location of the snow cover line and the timing of snow cover disappearance during the melting season, with positive winter NAO phases determining temperatures above-zero and a shorter persistence of snow over central Europe and western Eurasia [*Hori and Yasunari*, 2003].

[12] While earlier works suggested that the snow cover response does not last after April, recent studies have revealed that the NAO effect may remain until summer [*Ogi et al.*, 2003]. Figure 1 confirms that in June–July (JJ) the winter NAO signal is evident over snow covered areas of eastern Eurasia and North America, near the KAM sector ($r = -0.41$, $p < 0.05$) and the HUD sector ($r = -0.56$, $p < 0.01$), respectively. As the winter NAO does not have memory longer than one month, this lagged effect can be partially attributed to the persistence of snow cover, which has larger thermal inertia. The memory effect of snow cover, computed as the number of lagged months with significant autocorrelation ($p < 0.05$), shows that early spring Eurasian snow cover anomalies persist during the next three months

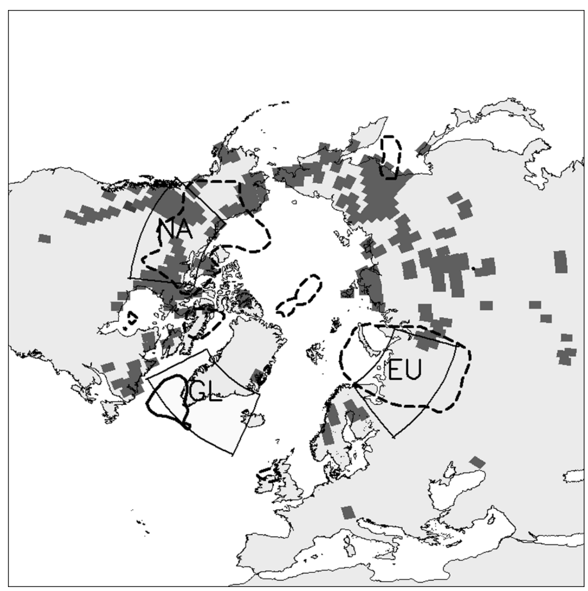

Figure 2. The summer-to-winter linkage. Correlation coefficient map between JJ snow covered cells (JJ North American snow cover) and the subsequent winter NAO (JJ 2-m air temperature). Shaded areas (solid/dashed lines) indicate negative (positive/negative) significant correlations with winter NAO (2-m air temperature) at $p < 0.1$ significance level. Significant temperature correlated areas are marked with black solid lines and referred in the text.

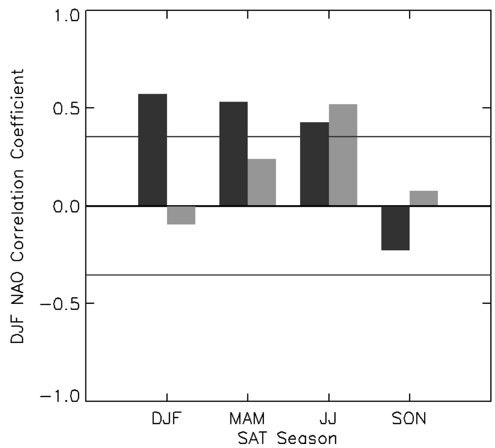

Figure 3. Seasonal evolution of the linkage between winter NAO and $\Delta(SAT)$. Dark (light) bars indicate correlations with the following (previous) $\Delta(SAT)$. Solid lines represent the p = 0.05 significance level.

(not shown), which enables to sustain the imprinted snow cover signatures until summer. These results are in agreement with *Saito et al.* [2004], who reported a loss of autumn Eurasian snow autocorrelation along the 1980's replaced by an emerging winter-spring persistence. Such snow persistence may partially result from the snow cover suppression of the total diabatic (sensible and latent) heating release in spring via the albedo effect. The cooling effect of snow cover produces persistent atmospheric responses in local and remote regions including a later warming of the Eurasian continent and altered land-ocean contrasts, the eastward expansion of a more zonal east Asian jet, the weakening (deepening) of the Okhotsk ridge (Aleutian low) or the strengthening of the North Pacific storm track [e.g., *Barnett et al.*, 1989; *Kodera and Chiba*, 1989]. Even when in summer snow cover vanishes over most of Eurasia these teleconnection patterns persist due to the additional cooling induced by the increase of soil moisture (snow-hydrological effect) after the snow melt [e.g., *Yasunari et al.*, 1991]. As the snow-induced changes in atmospheric circulation have an impact in remote locations they may initiate the snow cover response over other still covered areas in early summer, providing a physical mechanism in the propagation of the NAO-induced snow cover anomalies from Eurasia to North America.

[13] Since the snow cover response does not last after summer, the memorized winter NAO signal should be feedbacked into the summer circulation. In order to explore the physical basis of the summer transition, the correlation map between contemporaneous JJ North America snow cover (where the NAO signal is especially remarkable) and the 2-m air temperature field has been computed (Figure 2). It is seen that JJ snow extent is inversely correlated with the overlying temperature near the location of anomalous JJ snow cover (HUD and SCAN sectors) and positively with southern Greenland. Consequently, we define an index of subpolar temperature difference ($\Delta(SAT)$) from the 2-m air temperature averaged over those significantly linked sectors of Figure 2. Since the index reflects temperature variations associated with changes in JJ North American snow, the highest correlation coefficient for contemporaneous variations in JJ $\Delta(SAT)$ and those snow covered sectors of Figure 1 was for HUD (r = −0.64; p < 0.01), confirming the greater North American potential to induce the anomalous subpolar thermal distribution. However, as the winter NAO signal in western Eurasian snow is negligible in summer (see Figure 1), the EU sector contribution to JJ $\Delta(SAT)$ fluctuations was of secondary importance (not shown) and it has not been included in equation (1).

$$\Delta(SAT) = NA - GL \tag{1}$$

JJ $\Delta(SAT)$ is significantly correlated to the previous winter NAO (r = 0.43, p < 0.05), revealing a dynamical link between winter NAO and the summer climate (Figure 3). The summer anomaly patterns associated to JJ $\Delta(SAT)$ (Figure 4) show that summers of high JJ $\Delta(SAT)$ are

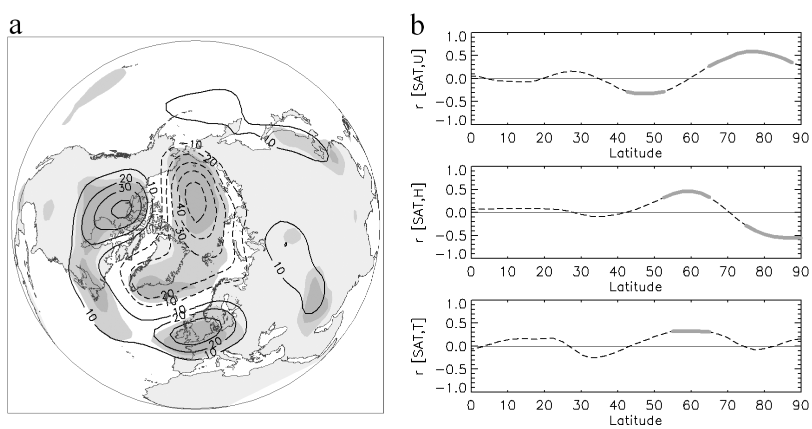

Figure 4. (a) Summer composite difference of 1000–200 hPa vertically integrated geopotential height (gpm) for high minus low JJ $\Delta(SAT)$ years (computed as those years above/below 0.5-sigma level). Shaded contoured areas indicate significant differences at p < 0.1 and p < 0.01 significance level; (b) Latitudinal distribution of the correlation coefficient between JJ $\Delta(SAT)$ and summer NH zonally averaged 2-m air temperature (lower panel) and 1000–200 hPa vertically integrated geopotential height (middle panel) and zonal wind (upper panel). Thick lines indicate significant correlations at p < 0.1 significance level.

characterized by a barotropic annular pattern of negative heights (cooling) over the Arctic Ocean and positive heights (warming) over the subarctic regions of North America, Europe and Eurasia (Figure 4a). Such pattern resembles a positive phase of the summer NAM, which is strongly associated with the Arctic front and the polar jet. Positive phases of the summer NAM display negative heights and temperatures over the Arctic Ocean and positive over the subarctic regions, especially northwestern Eurasia, the northern Sea of Okhotsk and Canada [*Ogi et al.*, 2004], coinciding with the temperature sectors controlled by summer snow cover. Thus, the surface temperature pattern associated with the summer NAM is almost identical to that of $\Delta(SAT)$. The correlation coefficient between JJ $\Delta(SAT)$ and the summer NAM reaches 0.57 ($p < 0.01$). These results confirm the potential of JJ snow-induced $\Delta(SAT)$ to support the transition between the winter and summer dominant modes. The linkage between JJ $\Delta(SAT)$ and summer NAM may arise from the favored warm continent/cold ocean pattern induced by the positive phases of JJ $\Delta(SAT)$. The enhanced thermal contrast induces a strengthened polar vortex (Figure 4b) and favors the transient eddy activity along the Arctic frontal zone. The anomalous westerlies may act as a waveguide for Rossby waves, developing a wave train pattern with ridges located over SCAN, KAM and HUD, which is characteristic of positive summer NAM phases.

4. Discussion and Conclusions

[14] We find that the winter NAO to summer NAM transition is favored by the asymmetrical thermal pattern associated to the NAO-induced snow anomalies in summer. Winters of high/low NAO phases induce low/high Eurasian snow cover, which, in turn, supports the continuation of the same winter NAO phase by propagating the snow cover anomalies to remote regions. Since JJ surface temperatures over NA are strongly controlled by HUD snow cover, ($r = -0.68$; $p < 0.01$) the impact of winter NAO over there results in thermal fluctuations and a positive JJ $\Delta(SAT)$. The enhanced thermal contrast along the Arctic coast enhances the zonal winds and the eddy activity, which maintain the summer NAM pattern [*Ogi et al.*, 2004].

[15] In addition, this index is also helpful to explain the summer snow-to-winter NAO linkage. Recently, a similar index (T_{SP}) has been identified as the best lagged predictor of the next winter NAO among those found in the literature [*Fletcher and Saunders*, 2006]. Here, JJ $\Delta(SAT)$ also correlates significantly with the upcoming winter NAO ($r = 0.52$, $p < 0.01$). Since $\Delta(SAT)$ is strongly controlled by summer HUD snow cover, our results also reveal that winter NAO predictability resides, in greater extent, over JJ snow covered areas of North America (HUD) ($r = -0.49$; $p < 0.01$) more than in Eurasia (see Figure 2). However, given that JJ HUD snow extent is previously modulated by spring SCAN snow, a portion of the winter NAO predictability may be originated in spring. The significant link between spring SCAN snow cover and the upcoming winter NAO ($r = -0.45$; $p < 0.01$) is consistent with the intercontinental propagation of snow cover anomalies and suggests that winter NAO may be earlier anticipated by spring snow cover, although with a marginal skill.

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References

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

Bojariu, R., and L. Gimeno (2003), The role of snow cover fluctuations in multiannual NAO persistence, *Geophys. Res. Lett.*, *30*(4), 1156, doi:10.1029/2002GL015651.

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

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

Fletcher, C. G., and M. Saunders (2006), Winter North Atlantic Oscillation hindcast skill, 1900–2001, *J. Clim.*, in press.

Hori, M. E., and T. Yasunari (2003), NAO impact towards the springtime snow disappearance in the western Eurasian continent, *Geophys. Res. Lett.*, *30*(19), 1977, doi:10.1029/2003GL018103.

Hurrell, J. W. (1995), Decadal trends in the North Atlantic Oscillation: Regional temperature and precipitation, *Science*, *269*, 676–679.

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

Kodera, K., and M. Chiba (1989), Western Siberian spring snow cover and east Asian June 500 mb height, *Pap. Meteorol. Geophys.*, *40*, 51–54.

Ogi, M., Y. Tachibana, and K. Yamazaki (2003), Impact of the wintertime North Atlantic Oscillation (NAO) on the summertime atmospheric circulation, *Geophys. Res. Lett.*, *30*(13), 1704, doi:10.1029/2003GL017280.

Ogi, M., K. Yamazaki, and Y. Tachibana (2004), The summertime annular mode in the Northern Hemisphere and its linkage to the winter mode, *J. Geophys. Res.*, *109*, D20114, doi:10.1029/2004JD004514.

Robinson, D. A., K. F. Dewey, and R. R. Heim (1993), Global snow cover monitoring, An update, *Bull. Am. Meteorol. Soc.*, *74*, 1689–1696.

Saito, K., T. Yasunari, and J. Cohen (2004), Changes in the sub-decadal covariability between Northern Hemisphere snow cover and the general circulation of the atmosphere, *Int. J. Climatol.*, *24*, 33–44.

Saunders, M. A., and B. Qian (2002), Seasonal predictability of the winter NAO from north Atlantic sea surface temperatures, *Geophys. Res. Lett.*, *29*(22), 2049, doi:10.1029/2002GL014952.

Saunders, M. A., B. Qian, and B. Lloyd-Hughes (2003), Summer snow extent heralding of the winter North Atlantic Oscillation, *Geophys. Res. Lett.*, *30*(7), 1378, doi:10.1029/2002GL016832.

Simmons, A. J., and J. K. Gibson (2000), The ERA-40 project plan, *ERA-40 Proj. Rep. Ser.*, *1*, 63 pp., Eur. Cent. for Med.-Range Weather Forecasts, Reading, U. K.

Thompson, D. W. J., and J. M. Wallace (2000), Annular modes in the extratropical Circulation, Part I: Month-to-month variability, *J. Clim.*, *13*, 1000–1016.

Wallace, J. M., and D. S. Gutzler (1981), Teleconnections in the geopotential height field during the Northern Hemisphere winter, *Mon. Weather Rev.*, *109*, 784–812.

Yasunari, T., A. Kitoh, and T. Tokioka (1991), 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. Jpn.*, *69*, 473–487.

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