NOTES AND CORRESPONDENCE

Observed Influences of Autumn–Early Winter Eurasian Snow Cover Anomalies on the Hemispheric PNA-like Variability in Winter

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ABSTRACT

The impact of the Eurasian snow cover extent on the Northern Hemisphere (NH) circulation is investigated by applying a lagged maximum covariance analysis (MCA) to monthly satellite-derived snow cover and NCEP reanalysis data. Wintertime atmospheric signals significantly correlated with persistently autumn–early winter snow cover anomalies are found in the leading two MCA modes. The first MCA mode indicates the effect of Eurasian snow cover anomalies on the Arctic Oscillation/North Atlantic Oscillation (AO/NAO). The second MCA mode corresponds with the forcing of Eurasian snow cover anomalies on the hemispheric Pacific–North America (PNA)-like atmospheric variations. This snow–atmosphere relationship may present a significant potential for wintertime predictability.

1. Introduction

The significant impact of Siberian autumn snow cover anomalies on the wintertime Arctic Oscillation/North Atlantic Oscillation (AO/NAO)-like atmospheric circulation in the North Atlantic sector has been identified in observations (e.g., Cohen and Entekhabi 1999; Saito et al. 2001) and through numerical modeling (e.g., Gong et al. 2003; Fletcher et al. 2009, hereafter F09). There are also indications that Eurasian snow cover influences atmospheric circulation over Asia and the North Pacific. A deepened Aleutian low is associated with the extension of wintertime/springtime Eurasian snow cover in modeling studies (Walsh and Ross 1988; Yasunari et al. 1991; Walland and Simmons 1996). Extensive East Asian snow cover in wintertime leads to negative height anomalies over the North Pacific in observations (Clark and Serreze 2000). Coupling between springtime Asian snow cover and the subsequent summer monsoon has been addressed in many papers (e.g., Barnett et al. 1989; Baran and Shukla 1999; Zhao et al. 2007). In a recent numerical modeling study (Orsolin and Kavaler 2009), variability in autumn–early winter snow cover extent over eastern Eurasia has been demonstrated to cause circulation anomalies over the North Pacific and the North Atlantic in winter through the development of the Aleutian–Icelandic low seesaw teleconnection, a PNA-like pattern with a strong eastward extension toward Europe (Honda and Nakamura 2001; Orsolin et al. 2008).

The main purpose of our study is to reexamine the coupling between Eurasian snow cover and NH atmospheric circulation and to provide observational evidence of the influence of autumn–early winter Eurasian snow cover anomalies on the wintertime hemispheric PNA-like atmospheric circulation. We investigate the possible influence of the Eurasian snow cover anomaly on NH atmospheric circulation in observations using a lagged maximum covariance analysis (MCA), also known as singular value decomposition (SVD) analysis (Bretihert et al. 1992). As noted by Czaja and Frankignoul (2002), lagged covariance is powerful in distinguishing between cause and effect. On monthly or longer time scales, the atmosphere primarily acts as a white noise forcing on the ocean, sea ice, and snow cover. If the snow cover only responds passively, then there should be no covariance when the snow cover anomaly leads by more than the atmospheric persistence time. If snow cover fluctuations have an impact on the atmospheric variable, then significant cross covariance should exist when the snow cover
anomaly leads. Such signatures are searched for here as a function of time lag, along the course of a year. If the Eurasian snow cover exerts a strong influence on climate variability beyond the AO/NAO, then it should also be present in the MCA results. The rest of this paper is arranged as follows: section 2 describes the data sources and analysis techniques, results are presented in section 3, and a summary is provided in section 4.

2. Datasets and methodology

The atmospheric datasets used here are geopotential height at 500 hPa (Z500) and wind at 300 hPa in the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996). Snow cover is obtained from monthly visible satellite data (Robinson et al. 1993) for 1979–2008. The climate (30-yr mean) of each month is removed separately for both atmosphere and snow cover datasets. To reduce the influence of trends and ultra-low-frequency changes, a second-order polynomial was removed by least squares fit at each grid point for all fields. At each grid point for both the snow cover and atmosphere anomalies, we filter out the tropical Pacific influence using a regression against the Niño-3.4 (5°N–5°S, 120°–170°W) sea surface temperature (SST) anomalies of the preceding months. The regression coefficient is selected as the maximum regression coefficient within the preceding 6 months.

To search a snow cover anomaly influence on the atmosphere, the MCA is applied as a function of time lag and season to monthly snow cover: and Z500 anomalies in the domain from 20° to 90°N. Sets of three successive months [e.g., January–March (JFM)] were considered with snow cover leading or lagging Z500 anomalies. For each season and lag, the MCA is based on 29-yr (87 months) data. We evaluate the significance of statistics in the MCA, the squared covariance (SC), and the temporal correlation (r) between the expansion coefficients of Z500 and snow cover using the Monte Carlo approach described in Czaja and Frankignoul (2002). One hundred ensembles of MCA between the scrambled Z500 and original snow cover are performed for each observed MCA.

3. Results

The MCA results are summarized in Fig. 1 for the leading two MCA modes between lag –5 and lag 2. Throughout most of the year, significant SCs are found at lags 0 and 1 (snow cover follows), showing that the dominant air–snow interaction is atmospheric control on Eurasian snow extent, consistent with Clark et al. (1999). On the other hand, significant SCs associated with the first MCA mode are only found when snow cover leads Z500 by 3–4 months during fall and winter [primarily October–December (OND), November–January (NDJ), and December–February (DJF)] and 1–2 months during spring [primarily February–April (FMA), March–May (MAM), and April–June (AMJ)]. For the second MCA mode, there are significant SCs when snow cover leads Z500 by 3–4 months during winter (primarily DJF and JFM).
a. Coupled patterns between autumn–early winter Eurasian snow cover anomalies and wintertime atmospheric variability

To illustrate the association of the wintertime atmospheric signal with the autumn–early winter snow cover, we have examined the MCA patterns of the leading two MCA modes for DJF Z500 and snow cover at the time lags between −3 and +1 months when significant SCs are found at the 10% level (Fig. 1). Each pair of patterns is formed from the heterogeneous covariance map for Z500 and the homogeneous covariance map for snow cover, which are constructed by regressing the Z500 and snow cover fields onto the normalized MCA–snow cover time series at each lag. At lag −2, the coupled patterns associated with the first MCA mode primarily reflect the forcing of the negatively polarized AO/NAO-like atmosphere by the positive Siberian snow cover (Fig. 2), as in Cohen and Entekhabi (1999). However, Fig. 2 indicates that OND European snow cover anomalies might also contribute the AO/NAO-like variability. Similar AO-like MCA patterns are found at lags −4 and −2 for DJF Z500 and lags −1 to −3 months for OND and NDJ Z500, with significant SCs at the 10% level. Since the AO-like response has been well studied (e.g., Cohen et al. 2007; F09), next we focus on results associated with the second MCA mode.

The coupled patterns associated with the second MCA mode, shown in Figs. 3a–3c, suggest that hemispheric PNA-like variability in winter, with both PNA and eastern Atlantic (EA) teleconnection features (Wallace and Gutzler 1981), can be strengthened by Eurasian autumn–early winter snow cover anomalies. Similar hemispheric PNA-like patterns are found (not shown) at lags −1 to −4 months for Z500 in JFM, with significant SC at the 10% level. When snow cover leads, the second MCA mode explains 16%–22% of the total covariance in winter, and correlations between the MCA–Z500 and snow cover time series are about 0.70 (significant at the 95% confidence level with the t test) (Table 1).

When the snow cover anomalies lead, the MCA maps in Fig. 3 are similar, and the MCA–snow cover time series are significantly correlated with each other (correlations greater than 0.65), indicating that the progression of the snow cover anomalies is coherent, both spatially and temporally. The significant SC of DJF atmospheric snow cover at the lags −1 to −3 appears to be relevant to the persistence of positive snow cover anomalies in the Tibetan Plateau (TP) and negative snow cover anomalies in the midlatitude, especially in Mongolia and northwest China, in Figs. 3a–3c. The below-normal (above normal) air temperatures in the TP (midlatitude Eurasia) due to the increased (reduced) albedo act to enhance (weaken) the local meridional temperature gradient in the vicinity of the East Asian subtropical jet (East Asian polar front jet) (Ren et al. 2010) and then increase (reduce) the strength of the zonal westerly wind. Indeed, projections of the 500-hPa zonal wind in DJF on the MCA–snow cover time series associated with Figs. 3a–3c show a strengthening and eastward expansion of the East Asian subtropical jet and a weakening of the East Asian polar front jet (not shown).

Since the TP is situated in the belt of the subtropics, cold air can either be carried over great distances by the prevailing westerlies and/or descend along its northern slopes. Consistent with these changes, the 500-hPa height in Figs. 3a–3c depicts the amplification of the tropospheric wave train over North America and north-central Eurasia. The relationship between the DJF Z500 and Eurasian snow cover several months earlier are so strong that strong correlation is also found implicitly in the snow cover field alone. Such persistence of snow cover anomalies explains why the lagged MCA can detect the across-season air–snow interactions taking place between atmosphere in DJF and snow cover from fall to early winter.

In contrast to results in Figs. 3a–3c describing the snow forcing on the atmosphere, the dominant relationships depicted in Figs. 3d and 3e are the PNA-like atmospheric forcings on snow cover at lags 0 and 1, respectively. These indicate that positive snow cover anomalies in the TP and strong negative snow cover anomalies from southern Europe to central Asia are induced by anomalous surface winds and snow–air fluxes associated with the PNA and EA.

b. Verification of MCA results with regression analyses

We attempted to determine which part of the snow cover anomaly had the strongest influence on the
PNA-like pattern by considering the averaged snow cover anomaly index in boxes over 30°–40°N, 85°–100°E (the TP center) and 40°–50°N, 90°–110°E (the Mongolia center). We mainly discuss the lag –2 in DJF (snow cover in OND), where the SC is the largest (Table 1), but similar results are found for other negative lags. Projecting the DJF Z500 anomaly on these two snow cover time series in OND suggests that snow cover anomalies in both TP and Mongolia centers exert a strong influence on the atmosphere (Figs. 4a and 4b). Meanwhile, projecting the DJF Z500 anomaly onto the difference between the TP and Mongolia snow cover indices in OND gave a Z500 anomaly pattern (Fig. 4c) that resembled the hemispheric PNA-like pattern derived from the MCA in Fig. 3b.

The maximum atmospheric response is about 25 m (20 m) for the TP (Mongolia) center. These are substantially smaller than the corresponding atmospheric signals in Fig. 3b, which is not surprising since the linear regression estimates in Fig. 4 are based on box-averaged
TABLE 1. Statistics associated with the second MCA mode when Z500 is assigned at DJF and JFM. Snow cover leads Z500 at negative lags. Statistics include (i) the SC ($\times 100$), (ii) the $r$ between the MCA–Z500 and snow cover time series, and (iii) the squared-covariance fraction (SCF). Note that the SC is formed by dividing by the number of grid points. The percentages in parentheses give their estimated significance level estimated by the Monte Carlo approach.

<table>
<thead>
<tr>
<th>Season</th>
<th>Z500 in DJF</th>
<th>Z500 in JFM</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$\text{LAG}$</td>
<td></td>
</tr>
<tr>
<td></td>
<td>$-4$</td>
<td>$-3$</td>
</tr>
<tr>
<td>SC</td>
<td>45.0</td>
<td>70.8</td>
</tr>
<tr>
<td></td>
<td>(49%)</td>
<td>(23%)</td>
</tr>
<tr>
<td>$r$</td>
<td>0.64</td>
<td>0.72</td>
</tr>
<tr>
<td></td>
<td>(34%)</td>
<td>(10%)</td>
</tr>
<tr>
<td>SC (%)</td>
<td>22.2</td>
<td>17.7</td>
</tr>
</tbody>
</table>

values rather than the maximum, and the MCA maximizes the covariance. Note that the above calculations are independent from the MCA, except for the choice of season and box locations, suggesting that the MCA results in Figs. 3a–3c are robust.

c. Response of storm tracks to autumn Eurasian snow cover anomalies

Snow cover feedbacks start from radiative and thermodynamical processes and then are amplified by the internal dynamics of the climate system (Clark and Sperrze 2000; Cohen et al. 2007). F09 show that the transition of the response from a thermally direct response to a Siberian snow cover in the onset phase to an AO-like response involves strong eddy-mean flow interactions. Here the snow-cover-related stormtrack changes are explored by computing the bandpass-filtered 300-hPa transient eddy kinetic energy $[\left(u^2 + v^2\right) / 2]$, associated with 2–8-day filtered fluctuations and its simultaneous and lagged regression patterns onto the normalized MCA–snow cover time series associated with the homogeneous snow cover patterns in Fig. 3b.

Figure 5a is indicative of reduced storm activity in Eurasia and south of Greenland, and increased storm activity in the Atlantic Ocean and western Europe in OND. In NDJ (Fig. 5b), storm activity is further significantly reduced in the extratropics because of the weakened westerlies. From OND to JFM (Figs. 5b–5d), the patterns are noted for the presence of dipole-like structures over the central and eastern parts of the Pacific and Atlantic basins, consistent with the enhanced (weakened) East Asian subtropical jet and North Atlantic polar front jet (East Asian polar front jet and North Atlantic subtropical jet). These patterns, depicting the southward (northward) migration of the storm tracks from their time-mean positions in the North Pacific (Atlantic), bear considerable similarity to the second (first) eigenvector variability of the storm tracks in the North Pacific (Atlantic) in Lau (1988), which can induce the PNA (EA) teleconnection.

![Fig. 4. Regression maps of DJF Z500 anomaly field against the indices of OND snow cover changes defined as (a) the snow cover averaged in the TP center (30°–40°N, 80°–100°E), (b) the Mongolia center (40°–50°N, 90°–110°E), and (c) the difference in snow cover between the TP center and the Mongolia center. Contour interval is 0 m, with negative contours dashed and the zero line omitted. The shading denotes values significant at the 95% confidence level.](image-url)
4. Summary

In addition to the well-documented influence of Siberian snow cover on the AO/NAO, we have detected the significant impact that autumn–early winter Eurasian snow cover has on wintertime PNA-like variability through the MCA. Regression analysis based on the snow cover anomaly centers of action confirms these findings. This snow–atmosphere relationship suggests a greater potential forecast skill for wintertime atmospheric variability. The link between the wintertime atmosphere and early-season snow cover appears to originate predominately from the remarkable persistence of snow cover anomalies in the TP and midlatitude Eurasia, in which a combined source of topographic and diabatic heating might be important (Hoskins and Karoly 1981). Our results support modeling experiments by Orsolini and Kvarnström (2009) that the autumn–early winter eastern Eurasian snow cover...
distinctly influences the year-to-year PNA variability in early winter and impacts variability over the North Atlantic in late winter.

The storm tracks respond strongly to early Eurasian snow cover anomalies associated with the second MCA mode, and its evolution of spatial patterns from fall to winter is closely related to the PNA and E-A teleconnection patterns in winter. This confirms the previous understanding that strong transient eddy forcing is important for explaining the extratropical response to midlatitude SST, sea ice, and snow anomalies (Doser et al. 2007, F09).

As a word of caution, we cannot exclude the possibility that the relationship between wintertime atmospheric and early-season snow cover anomalies may be caused by a third exogenous variable, although the influence of the long-term trend and ENSO has been removed. Modeling studies are needed to investigate the growth and maintenance of the hemispheric PNA-like atmospheric response to Eurasian snow cover anomalies found here.

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REFERENCES


