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The role of the Siberian high in Northern Hemisphere climate variability

Judah Cohen

Atmospheric and Environmental Research, Inc., Cambridge, Massachusetts

Kazuyuki Saito[1] and Dara Entekhabi

Massachusetts Institute of Technology Cambridge, Massachusetts

Abstract. The dominant mode of sea level pressure (SLP) variability during the winter months in the Northern Hemisphere (NH) is characterized by a dipole with one anomaly center covering the Arctic with the opposite sign anomaly stretched across the mid-latitudes. Associated with the SLP anomaly, is a surface temperature anomaly induced by the anomalous circulation. We will show that this anomaly pattern originates in the early fall, on a much more regional scale, in Siberia. As the season progresses this anomaly pattern propagates and amplifies to dominate much of the extratropical NH, making the Siberian high a dominant force in NH climate variability in winter. Also since the SLP and surface temperature anomalies originate in a region of maximum fall snow cover variability, we argue that snow cover partially forces the phase of winter variability and can potentially be used for the skillful prediction of winter climate.

Introduction

Over the past decade great strides have been made in understanding how heating anomalies in one region of the globe can force the overlying atmosphere to deviate from its mean state in a consistent and predictable manner. This deviation from normal can then propagate from the local heating anomaly to remote regions around the globe. For example, El-Niño/Southern Oscillation (ENSO) describes quasi-periodic anomaly patterns in sea surface temperatures (SSTs) which are confined to the tropical Pacific Ocean basin. Yet, ENSO through atmospheric teleconnection patterns, influences regional climates around the globe [*Barnston*, 1994].

In the NH mid- to high-latitude cold season there are three semi-permanent and quasi-stationary surface features that dominate the synoptic and even seasonal weather patterns. Two of these semi-permanent surface features reside in the major ocean basins of the NH - the Icelandic and Aleutian lows. The third, the Siberian high, resides over the world's greatest land mass - Asia. The intensity and coverage of this coldest and densest air mass in the NH, are influenced by diabatic heating anomalies associated with the underlying snow cover [*Foster et al.*, 1983; *Sahsamanoglou et al.*, 1991; *Clark and Serreze*, 2000].

Eurasian snow cover has large interannual variability and heating anomalies forced by snow cover in the region of the Siberian high can propagate remotely through favorable atmospheric teleconnection patterns. As with SSTs, anomalies occur on relatively large scales and the residence time of anomalies range from weeks [*Clark and Serreze*, 2000] to months [*Iwasaki*, 1991; *Walland and Simmonds*, 1997; *Cohen and Entekhabi*, 1999]. In this letter, we demonstrate associations between the Siberian high and remote regional climates on seasonal time scales and suggest the importance of snow cover to this teleconnection.

Results

The dominant mode of variability for December, January and February (DJF) NH SLP has been described as two phases of an Arctic high/low SLP center barotropically coupled with the stratosphere [*Thompson and Wallace*, 1998] or the expansion/contraction of the Siberian high, forced in part by seasonal surface heating anomalies [*Cohen and Entekhabi*, 1999]. What has been lacking in previous studies, is more definitive demonstration of the origins of the DJF SLP anomaly, which stretches from Siberia across the North Pole into North America and the North Atlantic. To show this, grid point values of SLP and surface temperature (Ts) are correlated with geopotential heights [*Kalnay et al.*, 1996] and snow cover [*Robinson et al.*, 1993] for 1972-1999. Since there could be some issue with the accuracy of reanalysis data over data-sparse regions including the Arctic, therefore, we checked the reanalysis data with in-situ observed data from the International Arctic Buoy Program [*Thorndike and Colony*, 1980] and found them to be nearly equivalent for the Arctic region (monthly root mean square error <1hPa), i.e., less than or equal to expected instrument error in the region of interest.

It has been suggested that the internal mode of variability in the stratosphere is coupled to the surface AO during the winter months [*Baldwin and Dunkerton*, 1999]. To focus on the coupling between the lower stratosphere and mid-troposphere, independent of the surface analysis, we constructed an index using eigenvector decomposition, which maximizes the correlation between the 50 hPa and 500 hPa geopotential heights during DJF. This index, referred to as upper-Arctic Oscillation or uAO, is highly correlated with the surface AO (r=.83). The uAO index is the time series of the leading mode of variability using singular value decomposition (SVD) of the lagged-combined fields of 500 hPa and 50 hPa geopotential height anomalies averaged for the entire winter season. The 50 hPa geopotential height field leads the 500 hPa geopotential height field by about two

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[1] Now at Frontier Research System for Global Change, Tokyo, Japan.

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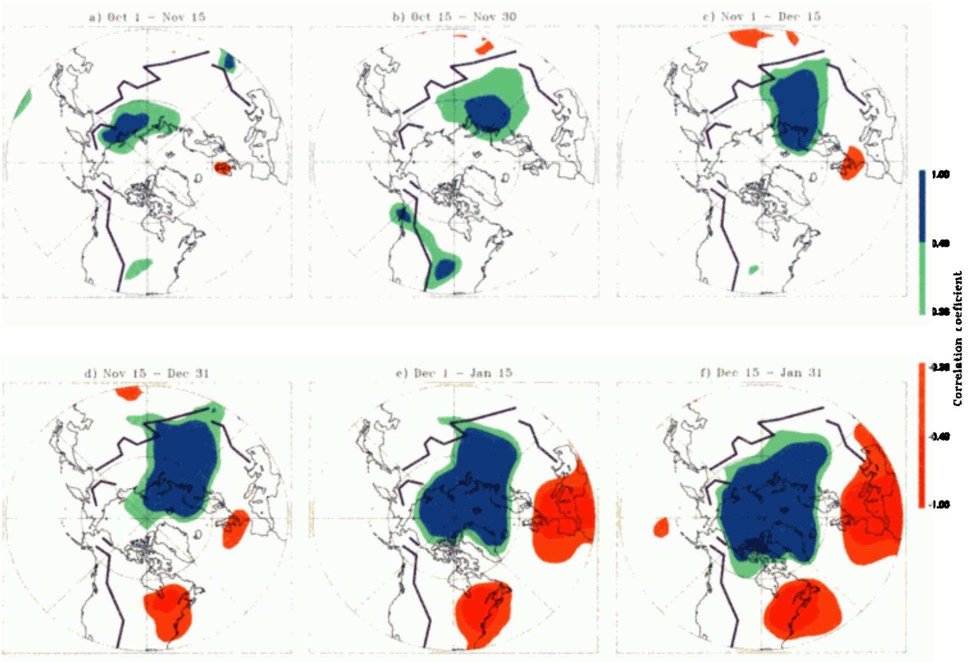

Figure 1. Maps (shading .38 = 95%, .49 = 99% significance) of DJF upper-AO (see text for definition) correlated with approximately 45 day average NH gridded SLP for (a) Oct. 1 - Nov. 15 (b) Oct. 15 - Nov. 30 (c) Nov. 1 - Dec. 15 (d) Nov. 15 - Dec. 31 (e) Dec. 1 - Jan. 15 (f) Dec. 15 - Jan. 31. Solid purple thick lines in a-e are 1000 meter isopleth. 27 years of data included from Sept. 1972-Feb. 1999. Region of light blue shading for correlations greater than .38 and dark blue for correlations greater than .49. Orange for correlations less than -.38 and red for correlations less than -.49.

weeks in the same season. We then correlated the time series of the upper level signal for DJF with gridded 45-day averaged SLP (Figure 1) and Ts (Figure 2), beginning with October 1.

Figure 1 shows that starting in October, SLP anomalies first appear in eastern Siberia. The initial SLP anomaly develops over the anomalous snow cover and the resultant heating anomaly, and then propagates to the west with the advancing snowline. The temperature anomaly also propagates west with the snowline but is shifted to the southeast, consistent with the snow-induced heating anomaly being advected downstream. The initial propagation is west because during October and November the waters of the Arctic Ocean in the Barents and Kara Seas are still unfrozen. Starting in late fall and early winter, propagation commences to the north across the now frozen Arctic Ocean [*Gloersen et al.*, 1992]. Due to the shallowness of the dense, cold air associated with the Siberian high, its expansion is

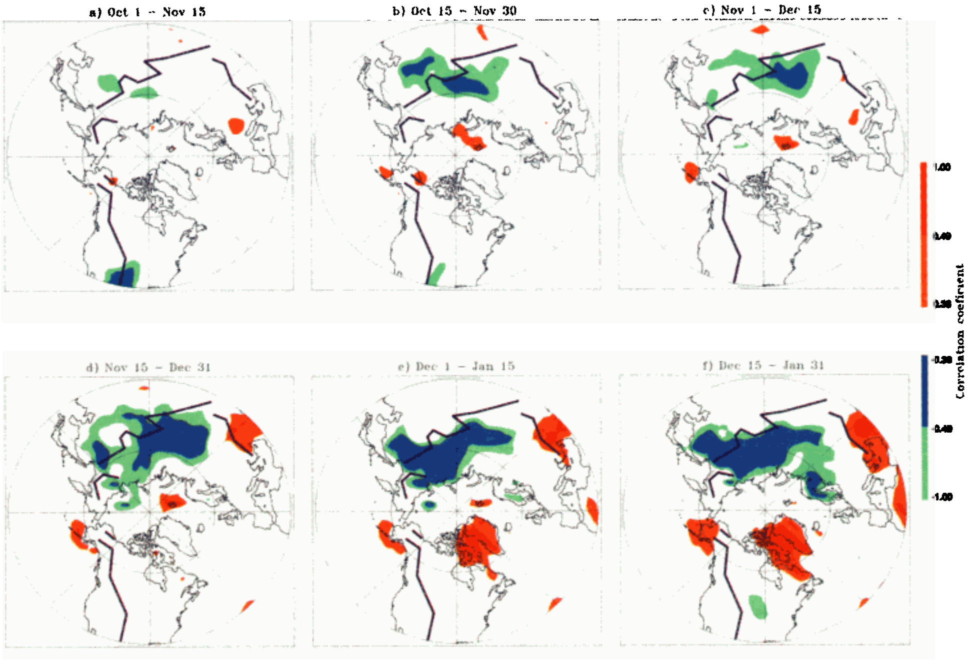

Figure 2. Same as Figure 1 except upper-AO correlated with gridded Ts 45 day averages. Region of orange shading for correlations greater than .38 and red for correlations greater than .49. Light blue for correlations less than -.38 and dark blue for correlations less than -.49.

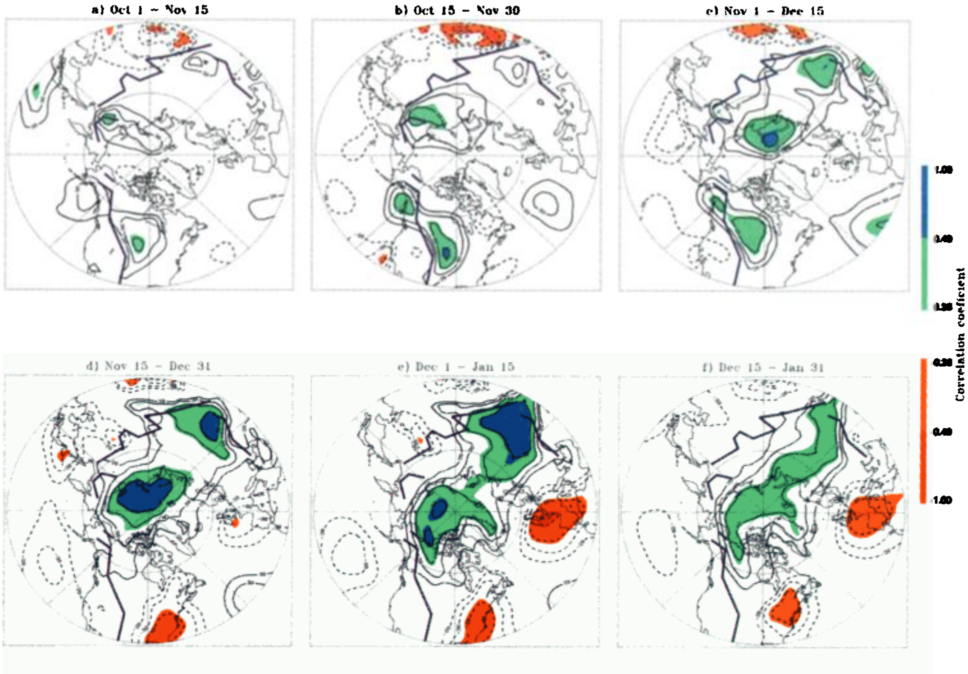

Figure 3. Same as Figure 1 except SON Eurasian snow cover correlated with gridded SLP 45 day averages (contours every .10). Region of light blue shading for correlations greater than .38 and dark blue for correlations greater than .49. Orange for correlations less than -.38 and red for correlations less than -.49.

restricted by high topography. The 1000-meter above sea level topographic contour (thick line in figures) clearly illustrates how the SLP anomaly emanating from the Siberian high is constrained by the high topography not only in Eurasia but even in North America. During winter (simultaneous uAO and SLP) the correlation plot is almost identical to the defined Arctic Oscillation [*Thompson and Wallace*, 1998]. Correlation of the time series from the SLP-derived-AO with the same 45-day periods gives similar results (not shown).

The surface temperature anomaly (Figure 2) also propagates first westward and then northward with the SLP anomaly. The Ts anomaly, similar to the SLP anomaly, is constrained by the high topography in central and western Asia during winter. However, unlike the SLP anomaly, the Ts anomaly propagates southward, lee of the high topography in eastern Asia. This is consistent with the positive correlation of the frequency of East Asian cold surges and the strength of the Siberian high [*Yihui*, 1990].

It has already been shown that September, October, November (SON) Eurasian snow cover anomalies are significantly correlated with DJF SLP and 500 hPa geopotential heights particularly in the Arctic and North Atlantic sectors [*Cohen and Entekhabi*, 1999]. In Figure 3 we provide further statistical evidence that the dominant mode of SLP variability may be partially forced by snow cover variability. To isolate the dominant role of surface heating anomalies due to interannual variability in snow cover across Eurasia, the Eurasian snow cover extent time series in the fall season is used as a climate index. Snow cover extent is defined as the areal snow cover for Eurasia in millions of squared kilometers [*Robinson et al.*, 1993]. In Figure 3 we present a series of plots of the correlation of Eurasian SON snow cover area with the same gridded 45-day averaged SLP as shown in Figure 1. Initially a perturbed region of SLP in eastern Siberia expands westward following the snowline across northern Russian and northern Europe and eventually across the entire Arctic into North America and the North Atlantic. Correlations indicate much of signal in this region to be significant at greater than the 95% confidence value. At lower latitudes in Western Europe and eastern North America SLP anomalies, of opposite sign from the SLP anomaly over the Arctic, also appear. This resembles Figure 1 and the dominant mode of variability for DJF SLP. Tests for field significance show SON snow cover and DJF SLP to be highly significant.

Figures 1-3 demonstrate that: both snow cover and the stratospheric polar vortex are associated with the dominant mode of variability of DJF NH SLP; this mode is inherently

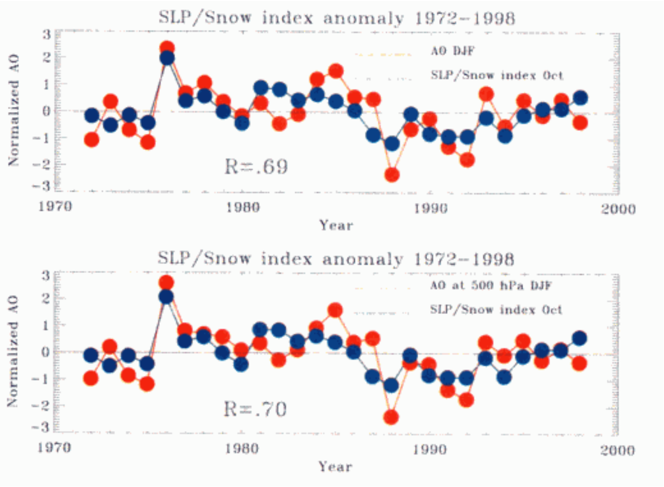

Figure 4. (a) Plot of Arctic Oscillation (AO) for DJF and SLP/snow index for October (see text for definition). Correlation (R) between two time series is also shown. (b) Plot of time series of leading mode of variability of 500 hPa geopotential height field (AO at 500 hPa) for DJF and SLP/snow index for October. Correlation (R) between two time series is also shown.

linked to the expansion and contraction of the Siberian high across the Arctic; and the early season Eurasian snow cover is a forcing mechanism (or at the very least a proxy) for the subsequent NH winter climate. We also note that significant correlations are consistently found in the mid to high-latitudes of the Arctic, Atlantic Oceans and the Eurasian and eastern North American continents but not the Pacific Ocean. This is consistent with the observational evidence and GCM results which show that the dominant mode of NH climate variability is not a zonally symmetric mode [*Deser*, 2000; *Monahan et al.*, 2000].

Conclusions

In this letter we present evidence that for the cool season, underneath the seemingly haphazard passage of weather events, the background unfolds in a systematic and predictable manner across large regions of the NH. As summer turns to fall in northern Eurasia, the Siberian high strengthens and spreads under increased divergence. If anomalously high snow cover is present, colder air temperatures result in strengthened or anomalous high pressure; less snow cover results in anomalous low pressure. Propagation of the SLP anomaly commences to the west towards western Russia and northern Europe. As fall turns to winter propagation to the west is eventually restricted by the strong maritime influence of the North Atlantic and the Gulf Stream in Europe. With the Arctic Ocean becoming nearly completely frozen by early December, the preferred route of expansion changes from west to north, and the Siberian high now expands northward over the frozen Arctic into North America. Not surprisingly, the phase of the Arctic Oscillation is most strongly correlated with sea ice concentration in the same region of the initial propagation of the SLP anomaly from the northern coast of Eurasia into the Arctic Ocean [*Mysak and Venegas*, 1998].

Hence we provide a methodological framework which describes how topography, snow cover, the juxtaposition of land and ocean masses, and sea ice dictate the manner in which cold, dense air spreads out from its source region in Siberia across large tracts of frozen land and ice in the Arctic and North Atlantic sectors. This work even has potential ramifications for paleoclimate studies. It provides a new mechanism for linking the uplift of the Tibetan Plateau with the inception of ice sheet growth in the North Atlantic sector. But besides the theoretical application of our results to different climate disciplines, this work has immediate practical application. As a demonstration of the potential predictive value of the analysis presented, we plot the AO for SLP and 500 hPa geopotential heights for DJF with an index constructed from SLP and snow cover from October in Fig. 4. The index was constructed from the multiple linear regression of Eurasian snow cover extent and the absolute value of SLP centered at $67.5^o N$ and $150^o E$, both for October. Correlation of the SLP/snow index and the AO for SLP and 500 hPa geopotential heights are .69 and .70 respectively. It has already been shown that the use of snow cover anomalies to predict surface temperature and precipitation has demonstrable skill [*Cohen*, 1999].

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References

Baldwin, M.P. and T.J. Dunkerton. Propagation of the Arctic Oscillation from the stratosphere to the troposphere. *J. Geophys. Res.*, *104*, 30937–30946, 1999.

Barnston, A.G. Linear Statistical short-term climate predictive skill in the Northern Hemisphere. *J. Climate*, *7*, 1513–1564, 1994.

Cohen, J. The potential use of snow cover variability for Northern Hemisphere climate prediction. In T. Barnston and E. O'Lenic, editors, *Proc. 24th Annual Climate Diagnostics and Prediction Workshops*, 240-243, 1999.

Cohen, J. and D. Entekhabi. Eurasian snow cover variability and Northern Hemisphere climate predictability. *Geophys. Res. Let.*, *26*, 345–348, 1999.

Clark, M.P. and M.C. Serreze. Effects of variations in east Asian snow cover on modulating atmospheric circulation over the North Pacific Ocean. *J. Climate*, in Press.

Deser, C. On the teleconnectivity of the "Arctic Oscillation". *Geophy. Res. Let.*, *27*, 779–782, 2000. 297–1300, 1998.

Foster, J., M. Owe and A. Rango. Snow cover and temperature relationships in North America and Eurasia. *J. Clim. Appl. Meteor.*, *22*, 460–469, 1983.

Gloersen, P., W.J. Campbell, D.J. Cavalieri, J.C. Comiso, C.L. Parkinson and H.J. Zwally. Arctic and Antarctic sea ice, 1978-1987: Satellite passive-microwave observations and analysis. *Tech. Rep. SP-511*, 290 pp., National Aeronautics and Space Administration, Washington, D.C., 1992.

Iwasaki., T. Year-to-year variation of snow cover area in the Northern Hemisphere. *J. Meteor. Soc. Jap.*, *169*, 209–217, 1991.

Kalnay, E. and Coauthors. The NCEP/NCAR 40-Year reanalysis project. *Bull. Amer. Meteor. Soc.*, *77*, 437–471, 1996.

Monahan, A.H., J.C. Fyfe and G.M. Flato. A regime view of Northern Hemisphere Atmospheric variability and change under global warming. *Geophy. Res. Let.*, *27*, 1139–1142, 2000.

Mysak, L.A. and S.A. Venegas. Decadal climate oscillations in the Arctic: A new feedback loop for atmosphere-ice-ocean interactions. *Geophys. Res. Let.*, *25*, 3607–3610, 1998.

Robinson, D. A., F. Dewey and R. Heim Jr. Northern Hemispheric snow cover: An update. *Bull. Amer. Meteor. Soc.*, *74*, 1689–1696, 1993.

Sahsamanoglou, H.S., T.J. Makrogiannis and P.P. Kallimopoulos. Some aspects of the basic characteristics of the Siberian anticyclone. *Int. J. Climatol.*, *11*, 827–839, 1991.

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

Thorndike, A.S. and R. Colony. Arctic Ocean Buoy Program Data January 1979 - December 1979. *Tech. Rep. 19*, 131 pp., Polar Science Center, University of Washington, Seattle, Washington, 1980.

Walland, D. J. and I. Simmonds. North American and Eurasian snow cover co-variability. *Tellus*, *49A*, 503–512, 1997.

Yihui., D. Build-up, air mass transformation and propagation of Siberian high and its relation to cold surge in East Asia. *Meteorol. Atmos. Phys.*, *44*, 281–292, 1990.

J. Cohen, Atmospheric and Environmental Research, Inc., 840 Memorial Drive, Cambridge, MA 02139. (email: jcohen@aer.com)

K. Saito, Frontier Research System for Global Change SEAVANS North 7F, 1-2-1 Shibaura, Minato-ku, Tokyo 105-6791 Japan (e-mail: ksaito@frontier.esto.or.jp)

D. Entekhabi, Parsons Laboratory, Massachusetts Institute of Technology, Cambridge, MA 02139. (e-mail: darae@mit.edu)

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