JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 111, D21104, doi:10.1029/2005JD006975, 2006

Northern Hemisphere snow cover and atmospheric blocking variability

Ricardo García-Herrera[1] and David Barriopedro[1]

Received 12 December 2005; revised 18 May 2006; accepted 7 July 2006; published 4 November 2006.

[1] The interseasonal relationship between Northern Hemisphere (NH) snow cover and regional blocking patterns is explored for a 31-year data set. It is found that snow cover exerts an important influence on regional atmospheric blocking, which, in turn, modulates snow cover extent at subcontinental scales. Observational results provide strong evidence of two primary linkages in the seasonal snow cover-blocking relationship that support an interannual persistence cycle: The first one links winter blocking over the Atlantic and the subsequent spring (summer) Eurasian (North American) snow cover anomalies; the second one implies that spring (summer) Eurasian (North American) snow cover precedes an anomalous winter Atlantic blocking activity. We describe the temporal stages of the snow cover-blocking relationship in the framework of a six-step conceptual model. According to that, an enhanced Atlantic blocking activity in winter favors a later spring snow disappearance through an enhanced cold advection toward western Eurasia. The resulting snow cover anomalies partially force an opposite-sign blocking response over west and central Pacific which is sustained through spring and early summer, presumably because of the persistence of snow cover anomalies. This anomalous pattern seems to play a role in the propagation of snow cover anomalies from Eurasia in spring to the Hudson's Bay region of North America in summer. The excessive snow cover over this region induces an asymmetrical temperature distribution, which, in turn, favors blocking activity over Europe and the West Pacific. The connection between autumn and winter climates is not clear but it could be related with the ability of autumn high ATL blocking activity to determine an early snow cover appearance in October over western Eurasia. This linkage completes a snow cover-blocking cycle of interactions which identifies snow cover as a candidate for the recently observed blocking trends and a contributor to the interannual persistence of winter climate.

Citation: García-Herrera, R., and D. Barriopedro (2006), Northern Hemisphere snow cover and atmospheric blocking variability, *J. Geophys. Res.*, *111*, D21104, doi:10.1029/2005JD006975.

1. Introduction

[2] In the last years, snow cover has been shown as an important forcing mechanism of climatic variability. Atmospheric circulation responses to anomalous Eurasian snow cover have been reported to occur locally and in remote regions and involving a wide range of timescales, from quasi-simultaneous to interseasonal lagged linkages. Several studies have provided strong evidence of the effects of snow cover in anomalous circulation patterns affecting Asia, North America, the North Pacific and the Atlantic regions (see references below). Simultaneously, there has been a renewed interest in the study of atmospheric blocking [e.g., *Barriopedro et al.*, 2006a, hereinafter BP06]. These quasi-stationary large-scale mid troposphere anticyclonic systems modulate midlatitude regional climates of NH by disrupting the westerly winds and suppressing the mean eastward progression of extratropical synoptic disturbances. Regional blocking episodes can persist for several weeks leading to significant anomalies in temperature and precipitation over large areas [*Trigo et al.*, 2004]. However, the snow cover response to persistent anomalous blocking patterns and the linking mechanisms between snow cover and blocking occurrence have not been deeply explored yet.

1.1. Snow Cover as a Forcing Factor of Interannual Atmospheric Variability

[3] Many studies have focused on the impact of autumn and winter snow cover in the subsequent climate since landmasses represent half of the total surface area in the extratropical NH and are extensively covered during cold seasons. Observational and model simulations have documented that an anomalously high Eurasian snow cover in late winter forces a delayed springtime surface air temperature increase, an enhanced spring and summer soil moisture as well as some significant patterns on the downstream side of the continent, as the deepening of the Aleutian low in the

[1]Departamento de Física de la Tierra II, Facultad Ciencias Físicas, Universidad Complutense de Madrid, Madrid, Spain.

0148-0227/06/2005JD006975$09.00

North Pacific Ocean, enhanced east Asian westerly winds, negative 500 hPa height anomalies or the strengthening of the North Pacific storm track induced by local anomalous temperature [*Hahn and Shukla*, 1976; *Yamazaki*, 1989; *Barnett et al.*, 1989; *Walland and Simmonds*, 1997; *Bamzai and Shukla*, 1999; *Clark and Serreze*, 2000].

[4] Atmospheric circulation responses to late winter snow cover anomalies have been attributed to changes in the diabatic (sensible and latent) heat and water (soil moisture and precipitation) balance fluxes in the surface layer occurring throughout the snow cover disappearance season [*Yasunari et al.*, 1991, hereinafter referred to as YA91]. Winter and early spring atmospheric responses have been essentially found in situ because of the reduced ground heating by the local albedo effect [*Namias*, 1985; *Leathers and Robinson*, 1993; *Groisman et al.*, 1994]. In summer, when the snow cover has almost disappeared, atmospheric responses have been documented in remote downstream regions because of the dominant lagged snow hydrological (soil moisture) effect after the snowmelting [YA91].

[5] In addition to other surface forcing factors of long-term memory like ocean surface sea temperatures (SSTs) and sea-ice extent, snow cover is considered a potential precursor of climatic variability in interannual and decadal timescales. Numerous studies have described the inverse relationship between winter-spring snow cover and the succeeding Indian summer monsoon rainfall [*Blanford*, 1884; *Walker*, 1910; *Kripalani and Kulkarni*, 1999]. Most studies have attributed this linkage to the inverse relationship between Eurasian snow cover in winter and the mean 500 hPa ridge location over northwest India in April, which is an important precursor for the seasonal forecast of Indian summer monsoon rainfall [*Hahn and Shukla*, 1976; *Dickson*, 1984].

[6] On the other hand, Eurasian snow cover has also been considered as a likely precursor of the NAO/AO phase. Positive Eurasian snow cover anomalies during warm (April–October) season [*Bojariu and Gimeno*, 2003], summer [*Saunders et al.*, 2003, hereinafter referred to as SA03] and autumn [*Cohen and Entekhabi*, 1999; *Cohen et al.*, 2001] seem to excite a negative NAO/AO like pattern in the subsequent winter. Simultaneously, winter NAO/AO has been identified as a forcing mechanism of spring Eurasian snow cover [*Hori and Yasunari*, 2003] and summer high-latitude climate [*Ogi et al.*, 2003]. However, the linking mechanisms are not clear yet. The predictive winter NAO/AO signal has been described in terms of the extension/contraction and propagation of the Siberian high induced by the seasonal surface heating anomalies of snow cover [*Cohen and Entekhabi*, 1999; *Cohen et al.*, 2001], the upward (downward) vertical propagation of wave activity flux (WAF) in autumn through the polar vortex [*Cohen et al.*, 2001; *Saito et al.*, 2001; *Gong et al.*, 2002] or the anomalous longitudinal differences of summer surface temperature created by June–July snow cover anomalies [SA03]. Eurasian snow cover and NAO oscillations covary at the same quasi-biennial and quasi-decadal timescales suggesting snow cover as a potential forcing of the reddish NAO spectrum [*Saito and Cohen*, 2003].

1.2. Atmospheric Blocking as a Temperature Controller

[7] There are numerous theories about the formation and maintenance of blocking events [*Egger*, 1978; *Reinhold and Pierrehumbert*, 1982; *Simmons et al.*, 1983; *Frederiksen*, 1982; *Tsou and Smith*, 1990]. However, the modulating factors of blocking variability at interannual or interdecadal timescales have not been properly addressed. Some studies have provided a feasible mechanism for interannual blocking variability following the *Charney and DeVore* [1979] conceptual model [*Shabbar et al.*, 2001; BP06]. According to this theory, blocking is a metastable equilibrium state determined by topographical and thermal forcing between continents and oceans, so that the zonally asymmetric thermal distribution forced by the land-sea temperature contrast determines a favorable environment for the blocking occurrence. *Tung and Lindzen* [1979] have linked blocking occurrence to the linear resonance of planetary waves with the surface thermal contrast. *Kikuchi* [1971], using a quasi-geostrophic model, found that land-sea thermal contrast was important in determining those preferred longitudes for blocking formation, in addition to orography. Following the *Charney and DeVore* [1979] conceptual model, BP06 provided observational evidence of a dynamic link between the zonally asymmetrical temperature distributions induced by the main regional NH teleconnection patterns (TCPs) and winter blocking variability over three blocking sectors (Atlantic, Europe, and East Pacific, respectively). However, the West Pacific sector exhibited a distinctive behavior, suggesting that other mechanisms than the regional anomaly pattern, may control the asymmetrical temperature distribution over the Eurasian sector.

[8] Thus the boundary conditions determining asymmetrical thermal patterns may influence blocking occurrence. Some studies have evaluated the impact of Pacific SSTs in blocking variability [*Mullen*, 1989; *Tibaldi et al.*, 1997]. However, the role of surface land temperature modulated by other forcing factors remains still unexplored. Since snow cover modifies local and remote surface temperature patterns, it may be presumed as a potential candidate of blocking variability. Conversely, the role of blocking in altering temperature and precipitation patterns may exert a significant response in snow cover. Despite the vast amount of literature addressing the influence of snow cover on large-scale circulation in different regions and seasons, there is not a comprehensive analysis linking these results within a unified conceptual framework. The objective of this paper is to examine the multiseasonal relationship between snow cover anomalies and regional blocking patterns in the NH in an attempt to build a conceptual model of interseasonal interactions governing the complex snow cover-blocking linkage.

[9] The paper is organized as follows. Snow cover data and the definition of a blocking index are described in the next section. In section 3, statistical analyses are used to explore the snow cover-blocking relationship in all seasons. This section also addresses the physical mechanisms involved in the multiseasonal linkages. The snow cover-blocking relationship is discussed in section 4. Finally, the multiple linkages are presented in a six-step conceptual model

in order to describe an annual cycle of teleconnections, involving the quasi-simultaneous and lagged linkages identified in this paper.

2. Data and Methodology

[10] 31 years (1972–2002) of snow cover data were provided by the Rutgers University Climate Lab (RUCL) [*Robinson et al.*, 1993]. They were originally derived from weekly NOAA visible satellite charts and consist of monthly series of (1) area of snow extent for Eurasia, North America (without Greenland) and NH and (2) 89 $\times$ 89 NH gridded snow matrixes, where each element represents the percentage of snow-covered surface of each cell. Cell resolution ranges from 16 km^2 to 42 km^2. Monthly series of gridded snow cover and regional areal products have been used to obtain seasonal snow cover data sets, with winter, spring, summer and autumn defined as DJF, MAM, JJA and SON, respectively.

[11] The objective blocking detection method designed by BP06 was used to derive a 31-year (1972–2002) blocking database from daily 500 hPa geopotential height fields on a (2.5,2.5)° NH grid of the NCEP-NCAR reanalysis database [*Kalnay et al.*, 1996]. Blocking data provide daily information on blocking occurrence, duration and other flow-related parameters as blocking intensity and extension over four different sectors: Atlantic (260–0)°E (ATL), Europe (0–90)°E (EUR), West Pacific (90–180)°E (WPA) and East Pacific (180–250)°E (EPA). In order to assess the snow cover-blocking relationship, a blocking index (*BI*) was obtained by normalizing the projection of the monthly 500 hPa height field anomaly over the monthly composite blocking pattern computed for each blocking sector:

$$BI_S = \frac{\langle Z_b, Z_m \rangle_S}{\langle Z_b, Z_b \rangle_S} \quad (1)$$

where brackets indicate the normalized projection and Z_b and Z_m are the monthly blocking pattern composite and the current monthly height field anomaly vector, respectively. The blocking pattern (Z_b) was derived by compositing the 500 hPa geopotential height anomalies for those days (falling in the given month) when a blocking was detected over each sector. The superindex S indicates the sector where the blocking occurs.

[12] Finally, the winter AO and NAO indices defined by the Climatic Prediction Center (CPC) of the NOAA have been used to assess the role of the winter dominant mode in the relationship between snow cover and atmospheric blocking (http://www.cpc.ncep.noaa.gov/).

[13] The snow-blocking relationship is examined using the Pearson correlation coefficient (*r* hereinafter). The level of significance has been fixed at $p < 0.05$ for a two-tailed student t-test, taking into account the reduction in degrees of freedom due to time series autocorrelation. The effective number of degrees of freedom has been estimated by including autocorrelation coefficients in both time series with lags of L/2 years, where L is the time series length [*Oort and Yienger*, 1996]. In order to minimize the influence of the linear trends and multiyear signal variability on the magnitude and significance of the correlation analysis time series have previously been detrended by computing the slope in the linear regressions versus time. In any case, the magnitude and significance of the results remained almost unaffected after detrending time series. Those years departing from the 1-σ level above/below the climatic mean have been considered as high/low snow extent and blocking activity periods. Composites of atmospheric fields have also been computed for them. The statistical significance of the composite difference has been estimated from a two-tailed student t-test with a significance level of $p < 0.05$.

[14] This analysis is performed assuming that opposite patterns will occur for opposite phases of blocking activity. Previous studies [*Liu*, 1994] have shown that, despite its nonlinear component, blocking patterns (low zonal flow) do not significantly last more than opposite-blocking patterns (high zonal flow). As a consequence, atmospheric responses in temperature, zonal wind or height geopotential for periods of high/low blocking activity have a quasi-linear nature, with high blocking activity showing opposite related patterns to those occurring for low blocking activity [*Trigo et al.*, 2004].

3. Results

[15] Figure 1a shows the correlation coefficients between winter BI_{ATL} and the upcoming (dark bars) and previous (light bars) seasonal NH snow cover. Two lagged linkages can be inferred. The first one links winter ATL blocking and the contiguous snow cover extent, with the correlation coefficient increasing when season progresses from winter (r = 0.41) to spring (r = 0.56) and summer (r = 0.52). The other one implies a snow-leading relationship, with positive spring (r = 0.54) and summer (r = 0.66) snow cover anomalies preceding a winter ATL blocking-like pattern.

[16] Since the atmosphere does not have memory longer than one month, these lagged correlations should be related to snow cover. Its larger thermal inertia enables the atmospheric influences to be imprinted as snow cover anomalies, which can persist and propagate to remote regions. When the correlations are broken down into Eurasian (Figure 1b) and North American (Figure 1c) landmasses, both regions contribute to the NH snow cover-blocking relationship in a different way:

[17] 1. The linkage between winter ATL blocking and the subsequent spring snow cover is evident over the Eurasian continent, the correlation coefficient reaching 0.48. However, in summer the winter ATL signal is essentially confined to North America (r = 0.61). When monthly lagged correlations are computed, the snow cover response appears in Eurasia two or three months earlier than in North America (not shown). Since the climatological snowmelt season is delayed in high-latitude regions of North America relative to Eurasia, both linkages can be reflecting different timings in the snow cover response to winter ATL blocking.

[18] 2. Simultaneously, the winter ATL blocking response to previous NH spring snow cover is mostly attributable to Eurasia (r = 0.51), while North American snow cover appears as the main contributor to the NH snow cover signal in summer (r = 0.68). Since the snow cover signal is evident from spring to summer when the whole NH is considered, Eurasian and North American linkages may just indicate two regional manifestations of the same global

process consisting of the propagation of snow cover anomalies from Eurasia in spring to North America in summer.

[19] Thus both linkages seem to determine an interannual persistence cycle from one winter to the next governed by seasonal interconnections in the temporal march of the blocking-snow cover relationship. Accordingly, winter ATL blocking modulates the upcoming spring Eurasian snow cover, which in turn might influence the upcoming winter ATL blocking by propagating the snow cover anomaly into North America in summer. The question now arises in determining (1) how the spring snow cover anomalies are generated and related to the previous winter blocking, (2) how the Eurasian spring snow cover anomalies can propagate to North America and (3) how these anomalies may alter the ATL blocking activity in the following winter.

[20] In an attempt to explain these questions, the role of the other blocking sectors has also been investigated by computing seasonal cross correlations between regional snow cover and *BI* indexes. Figure 2 summarizes the interseasonal significant linkages found in this study. In the next subsections the temporal stages of the complex relationships of Figure 2 are discussed in detail. Since the timescales involved in each step were sometimes confined to just one season, monthly correlations have also been computed to discern between snow-leading and snow-lagged relationships. Additionally, Figure 3a summarizes some teleconnections linked to snow cover and physical processes described in the literature, which will help to illustrate the following sections. Finally, Figure 3b shows some key snow-covered regions in the blocking-snow cover relationship.

3.1. Stage 1: Role of Winter Atlantic Blocking in Spring Eurasian Snow Cover

[21] Figure 1b shows a significant linkage between winter BI_{ATL} and spring (r = 0.48) Eurasian snow cover. The evolution of this linkage at monthly timescales reveals that the Eurasian snow cover response remains significant from January to June, with the correlation coefficient reaching a maximum in March and decreasing as the year progresses (not shown). This maximum snow cover response in late winter and early spring is consistent with the findings of *Iwasaki* [1991] who described that the snow cover features are sustained through early winter and then disappear following the climatological disappearance of snow cover. Lag-correlation maps between winter BI_{ATL} and the gridded snow cover in successive seasons (Figures 4a and 4b) show that significant areas are essentially confined to western Europe in winter, whereas the signal increases and progresses northeastward in the next spring to those areas of higher susceptibility to disappearance.

[22] Since snow extent is strongly associated to the timing of disappearance, the Eurasian snow cover response can be attributed to a blocking control in the location of the snow cover line and the timing of snow cover disappearance during the melting season, when snow cover is susceptible to atmospheric control. *Shinoda et al.* [2001] described that

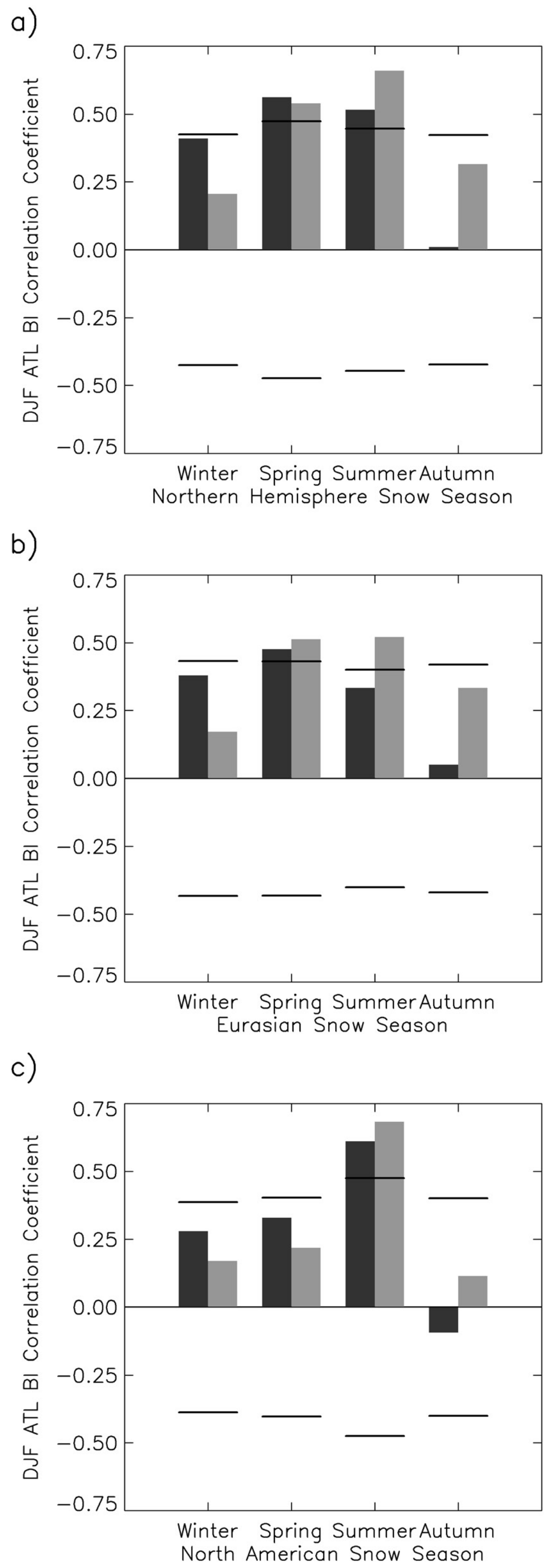

Figure 1. (a) Seasonal evolution of the linkage between winter BI_{ATL} and NH snow cover, (b) as in Figure 1a but for Eurasian snow cover, and (c) as in Figure 1a but for North American snow cover. Dark/light bars indicate correlations with the following/previous seasonal snow cover. Solid lines represent the $p < 0.05$ significance level after correction for autocorrelation.

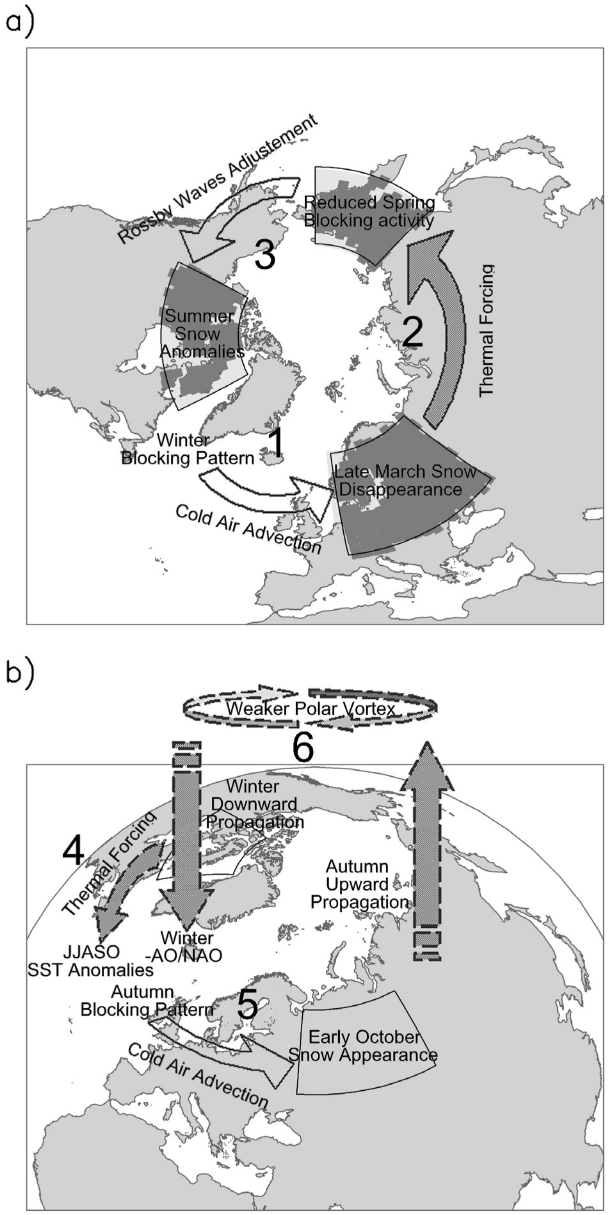

Figure 2. The ATL blocking-snow cover conceptual model. Schematic diagram illustrating the (a) first and (b) second phases of the proposed conceptual model with winter ATL blocking leading to spring Eurasian and summer North American snow cover anomalies (stages 1 to 3 in the text) (Figure 2a) and summer snow cover anomalies determining a blocking-like pattern over the ATL sector in the next winter (stages 4 to 6 in the text) (Figure 2b). Dark/light arrows indicate snow/blocking leading teleconnections. Dashed arrows denote other snow-leading teleconnections reported in the literature that may be involved in the blocking-snow cover relationship. Numbers label the stage number as cited in the text.

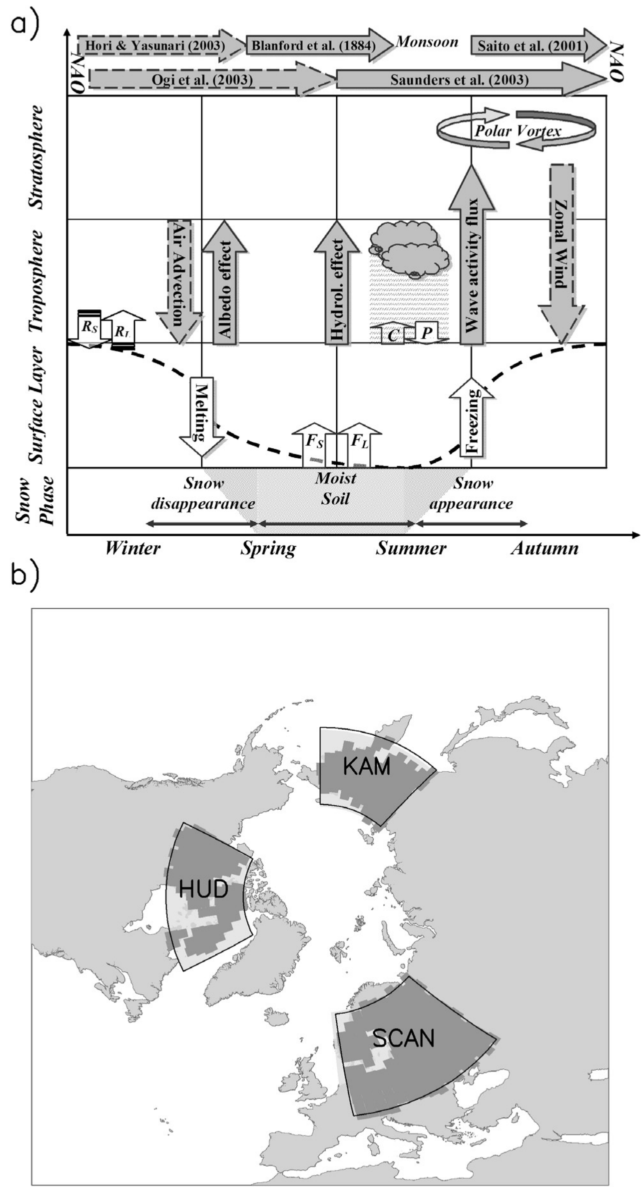

Figure 3. (a) Diagram illustrating some interactions between snow cover and atmosphere described in the literature. Solid/dashed shaded arrows indicate the effect of snow cover/atmosphere on the atmosphere/snow cover. Horizontal arrows represent referenced predictive linkages between snow cover and atmospheric variability (NAO and Indian monsoon). Light arrows indicate processes related to snow cover and the surface water and energy balance fluxes. Dashed line shows the seasonal march of snow cover extent from winter to autumn. R_S (R_L), incoming (outgoing) short (long) wave radiation; F_S (F_L), sensible (latent) heating flux; C (P), convection (precipitation). (b) Some snow-covered key regions as defined in the text: Scandinavian (SCAN), Kamchatka (KAM) and Hudson's Bay (HUD).

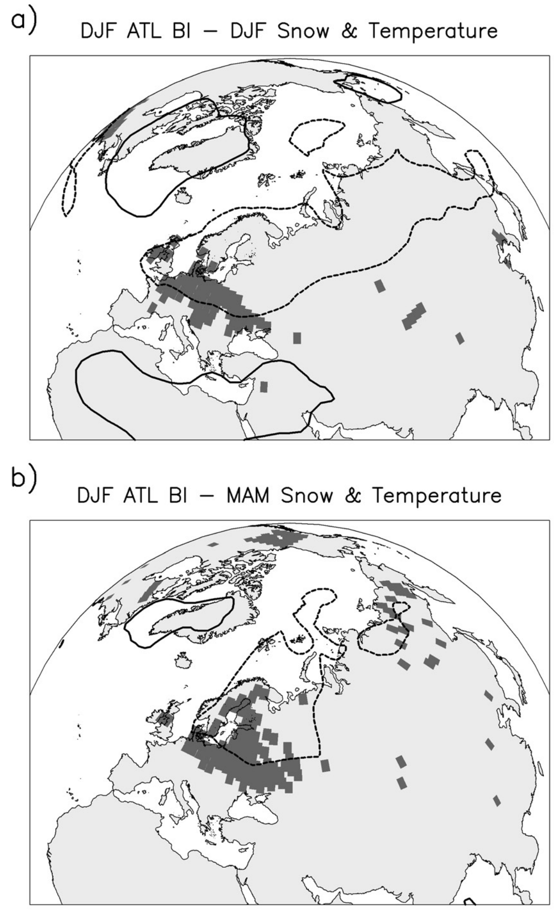

Figure 4. Significant correlation map between winter BI_{ATL} and snow-covered cells (shaded areas) and surface temperature (contour lines) in (a) winter and (b) spring. Shaded areas indicate significant positive correlations between winter BI_{ATL} and snow cover at $p < 0.05$ significance level. The $p < 0.05$ significance level for correlations between winter BI_{ATL} and surface temperature is represented by solid (positive correlations) and dashed (negative correlations) lines.

Eurasian snow disappears within a few weeks, starting in March over western Eurasia (north of the Caspian-Aral Sea). According to that, later snow cover disappearance years are characterized by delays up to three weeks in the disappearance timing and more persisting snow cover over western Eurasia. In contrast, when the snowmelting is accelerated, the snow cover boundary line retreats earlier over western Eurasia.

[23] Several studies have confirmed that the horizontal thermal advection by atmospheric circulation patterns plays a decisive role in determining the melting speeds, with above-zero temperatures favoring rain precipitation instead of snow and a shorter persistence of snow [*Shinoda et al.*, 2001; *Hori and Yasunari*, 2003]. In order to examine the physical mechanisms determining the snow cover response, the anomalous atmospheric patterns associated to winter ATL blocking have been computed (Figure 5a). The most prominent feature is the weakness of the westerlies and an east-west thermal dipole with a dominant warm center south of Greenland and below-normal temperatures over the Scandinavian region. These associated temperature patterns are mainly controlled by the advection of heat by the anomalous mean flow and favor cold advections from northern latitudes into central and western Europe during those phases of high blocking activity [*Trigo et al.*, 2004]. Figure 6 depicts monthly composite differences of snow line and the 0°C surface temperature isotherm (zero-line) for opposite phases of the winter BI_{ATL}. The snow line is defined as the boundary line delimiting those cells being more than 50% snow covered. The difference in snow line reveals that those snow-covered areas sensitive to winter ATL blocking lie between the zero-lines of high and low blocking winters. A higher ATL blocking activity during winter induces a later snowmelting in January over western Europe (Figure 6a). The snow line advances slower into central Europe in February (Figure 6b) and into western Eurasia in March (Figure 6c) and April (not shown).

[24] These results confirm that anomalous positive blocking patterns are likely to determine a slower snowmelting over central Europe and western Eurasia by maintaining surface temperatures below 0°C (see Figure 2a). Recently, *Hori and Yasunari* [2003] and *Saito and Cohen* [2003] have also reported that winter NAO exerts an atmospheric control in the disappearance of western Europe snow cover, with the signal being originated in January and propagating northeastward within the next months. However, it should be noted that, despite the short memory of blocking in monthly timescales, the snow cover response is sustained until early summer. This lagged effect could be partially attributed to the persistence of Eurasian snow cover anomalies. Snow cover fluctuations occur on large timescales, with residence times ranging from about two weeks [*Clark and Serreze*, 2000] to several months [*Iwasaki*, 1991; *Walland and Simmonds*, 1997; *Cohen and Entekhabi*, 1999]. To examine the persistence of snow cover anomalies the number of lagged months with significant autocorrelation has been computed (Figure 7a). The memory effect of March snow cover anomalies persists during the next four months, whereas the autocorrelation falls through late spring. The relatively shorter snow cover memory in January and February points that March snow cover anomalies may exert a major control in the subsequent climate than any other month, probably because Eurasia snow extent exhibits maximum fluctuations from March to April (Figure 7b). After March–April the blocking signature vanishes over western Eurasia and significant correlations are found across eastern Eurasia (not shown). The next sections discuss how the ATL blocking signature leads to snow cover anomalies there.

3.2. Stage 2: Role of the Persistent-March Snow Cover Anomalies in Pacific Blocking

[25] In spring the blocking-snow cover relationship is characterized by an inverse relationship between Eurasian snow cover and BI_{WPA} ($r = -0.66$). At monthly timescales, spring BI_{WPA} is significantly correlated with March–July snow cover, the correlation coefficient peaking in spring months and decreasing as summer approaches (Figure 8). Figure 9a depicts the contemporaneous composite difference pattern associated to high minus low spring Eurasian snow cover. The atmospheric signatures resemble those obtained for negative phases of WPA blocking activity

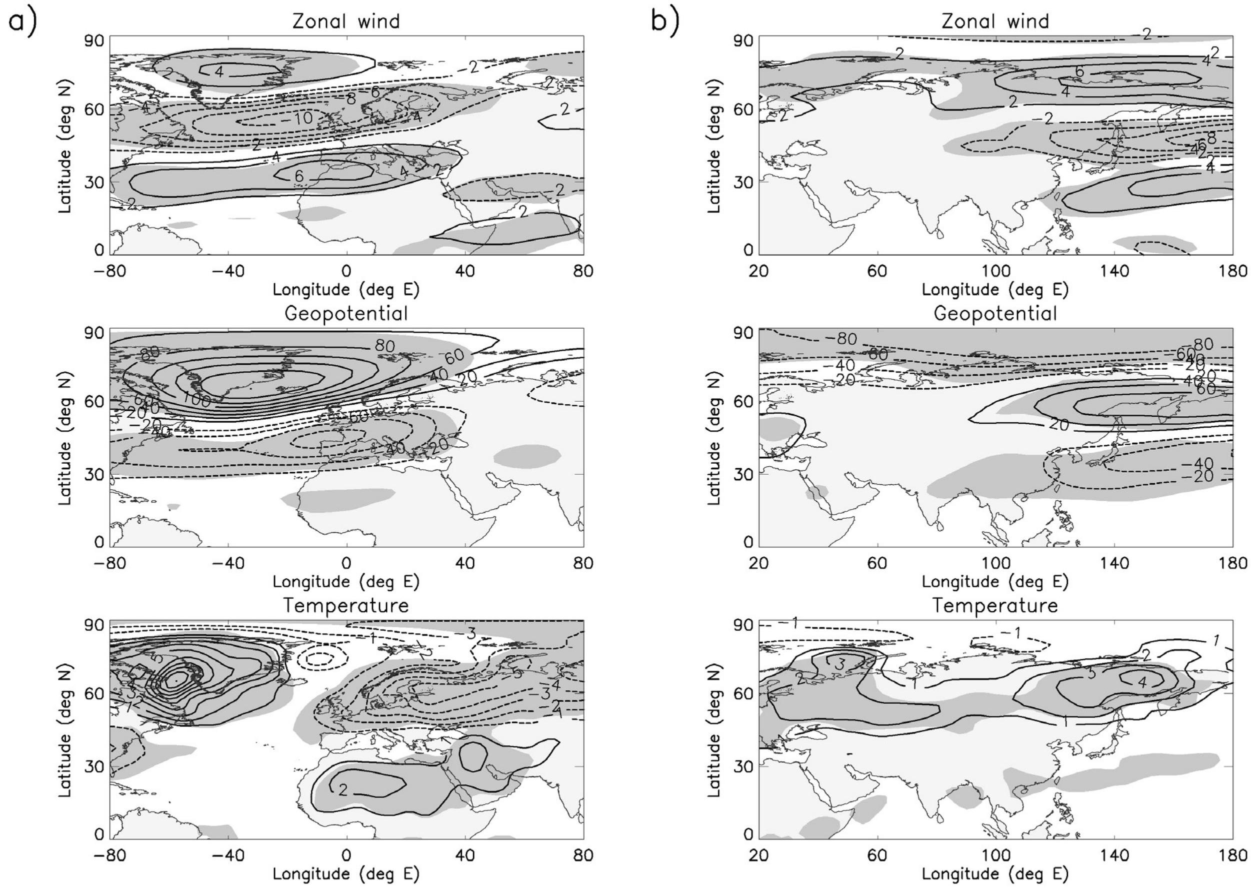

Figure 5. Contemporaneous composite difference of surface temperature (°C) (third panel), 1000–200 hPa geopotential height (gpm) (second panel) and 1000–200 hPa zonal wind (m s^{-1}) (first panel) for high minus low (a) BI_{ATL} winters and (b) BI_{WPA} springs. Solid/dashed lines denote positive/negative differences. Contour line intervals are 1°C, 20 gpm and 2 m s^{-1}, respectively. Shaded areas indicate significant differences at $p < 0.05$ significance level.

(compare with Figure 5b). As mentioned above, spring snow cover anomalies over western Eurasia may result from a previous anomalous winter ATL blocking activity and persist during the snowmelting season. Successive correlation maps of Figure 10 confirm that the linkage between snow and BI_{WPA} originates over the same snow-covered areas modulated by the previous winter ATL blocking activity and progresses northeastward, with the signal embracing a large portion of western Eurasia in March (Figure 10b). Thus the negative WPA blocking pattern in spring may result from the above-normal snow extent over western Eurasia in March due to an enhanced ATL blocking activity in winter. This result implies a WPA blocking dependence on the Eurasian snow disappearance timing.

[26] The local and remote atmospheric circulation patterns associated to snow cover anomalies in spring have been mainly attributed to fluctuations in the surface heat and water balances derived from an anomalous March snow (Figure 3a). Using General Circulation Models (GCMs) experiments, YA91 found that an increase of snow cover in March suppresses the sensible and latent fluxes in spring due to the cooling effect induced by the high albedo. The subsequent reduction of total diabatic heating (albedo effect) and the enhanced heating contrast between continent and ocean produce local and remote atmospheric responses persisting throughout the spring, including a later warming of the Eurasian continent, an enhanced east Asian jet, the weakening (deepening) of the Okhotsk ridge (Aleutian low) or the strengthening of the North Pacific storm track [*Barnett et al.*, 1989; *Kodera and Chiba*, 1989; *Clark and Serreze*, 2000]. YA91 also found that a late snow disappearance in March–April was followed by significant teleconnection patterns downwind of the Eurasian continent persisting until early summer, which is in agreement with the persistence of snow cover anomalies obtained in Figure 6a.

[27] Since blocking occurrence is sensitive to thermal contrasts, it could be expected that the WPA response arises from the snow cover control in land temperatures via the albedo effect (BP06). Correlation maps between monthly surface temperature and spring BI_{WPA} of Figure 10 confirm that spring WPA blocking activity is strongly modulated by March surface temperatures spreading over most of Eurasia. Additionally, the March averaged temperature over the SCAN sector (Figure 3b) shows significant correlations with spring BI_{WPA} ($r = 0.54$). Thus the WPA blocking

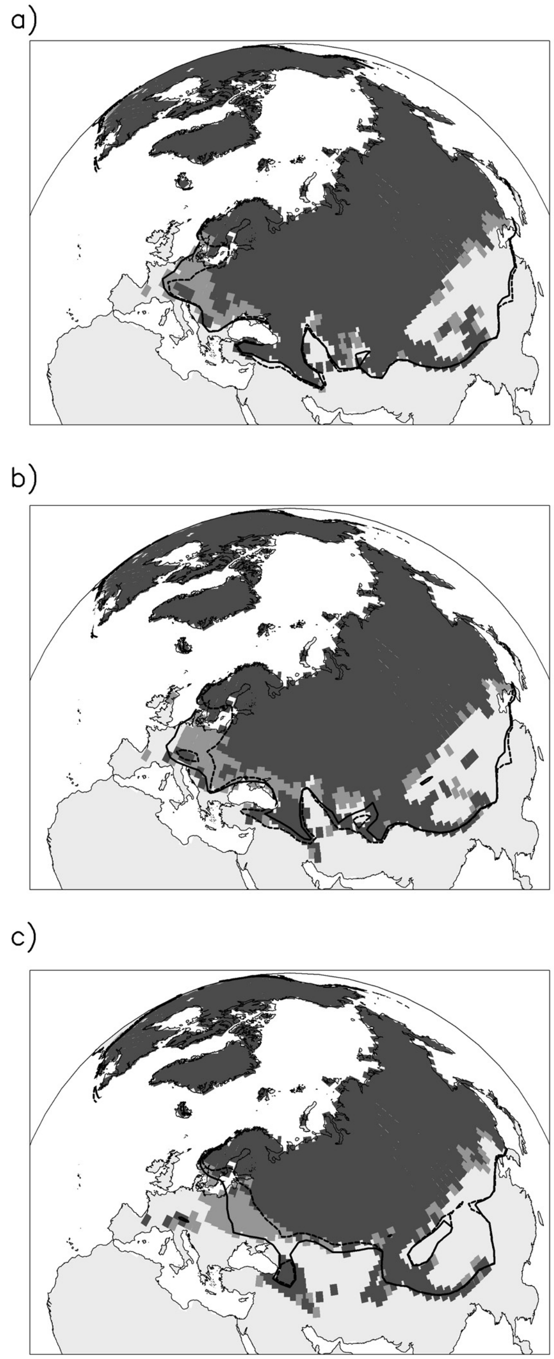

Figure 6. Monthly composite of snow-covered areas and surface temperature for high/low BI_{ATL} winters in (a) January, (b) February, and (c) March. Light/dark cells and solid/dashed lines indicate those cells more than 50% snow covered and the 0°C temperature line for high/low BI_{ATL} winters, respectively.

response could be explained as a remote response to the enhanced meridional thermal gradient over western and central Eurasia resulting from excessive snow cover, which provides an eastward expansion of the east Asian jet and a more zonal flow, inhibiting blocking occurrence over the WPA sector (see Figure 2a).

3.3. Stage 3: Role of Pacific Blocking in the Propagation of Spring Eurasian Snow Cover Anomalies

[28] After May, the anomalous cooling due to the albedo effect is not enough to induce circulation patterns downwind of the continent [YA91]. The weakened westerly winds over the region and the reduced contrast between continent and ocean would make the signal vanish. Instead of this, the correlation coefficient between spring WPA blocking and the Eurasian snow cover anomalies is sustained until July, suggesting a snow cover response to the persistent negative-blocking pattern over the WPA sector (Figure 8). Such signature can be attributed to the lagging snow-hydrological effect. According to YA91, although there is no albedo effect in summer, the atmospheric anomalous patterns derived from excessive March snow can be still evident because of an additional cooling induced by the increase of soil moisture after the snowmelt (Figure 3a).

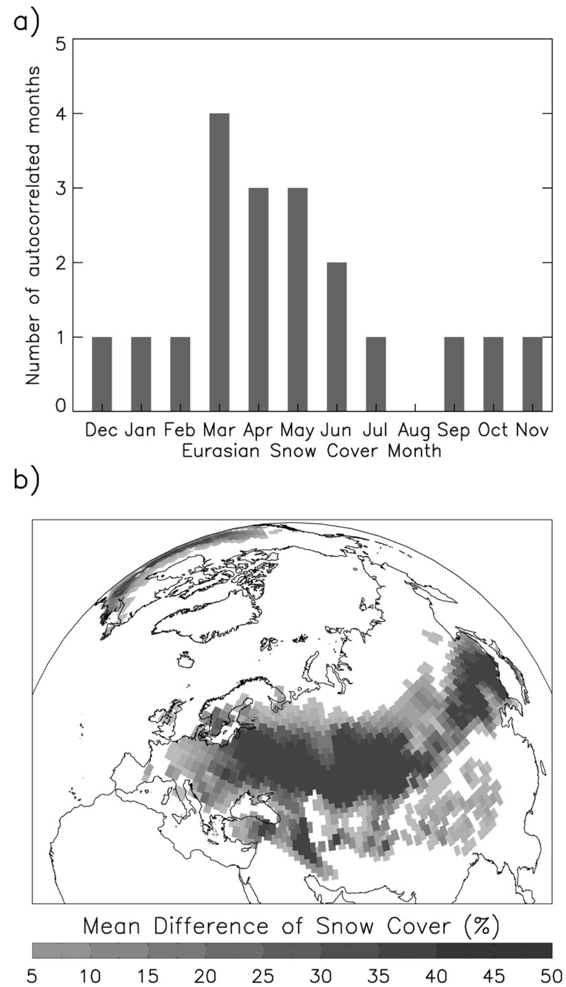

Figure 7. (a) Persistence of monthly Eurasian snow cover anomalies computed as the number of significantly autocorrelated months at $p < 0.05$ significance level and (b) 31-year mean snow cover decrease (in percentage of covered area) from March to April.

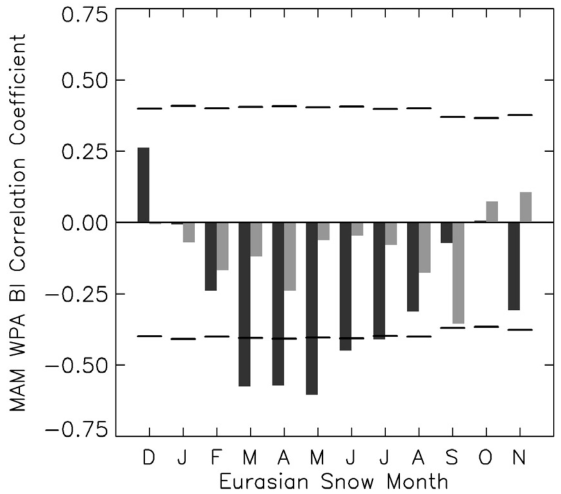

Figure 8. Monthly evolution of the linkage between spring BI_{WPA} and Eurasian snow cover. Dark/light bars indicate correlations with monthly snow cover of the current/previous year. Solid lines represent the p < 0.05 significance level.

Thus the opposite induced blocking-like pattern over WPA may persist during spring and early summer contributing to induce regional snow cover anomalies. Monthly cross correlations between BI_{WPA} and SCAN and KAM snow cover confirm that the interseasonal linkage between WPA blocking and Eurasian snow cover is characterized by a significant WPA blocking response to SCAN from March to April, which, in turn, influences KAM snow cover in May–July. The anomalous circulation patterns associated with opposite phases of spring BI_{WPA} reveal above-normal temperature anomalies over eastern Eurasia with a maximum located over KAM (Figure 5b). The induced warming under the anticyclonic area favors a shorter persistence of snow and an earlier northward retreat of the snow line (not shown). As stated before, the winter ATL blocking influence in snow cover dissipates over western Eurasia in April to reappear in May displaced over eastern Eurasia. This lagged winter ATL blocking signature may be attributed to the WPA blocking ability to propagate SCAN snow cover anomalies into KAM region. Indeed, the spring BI_{WPA} signature in snow cover is almost identical to that obtained for winter BI_{ATL} (not shown), suggesting that spring WPA blocking may participate in the propagation of the winter ATL blocking signal into eastern Eurasia.

[29] By June the winter ATL blocking signal starts to be significant over snow-covered areas of North America, the

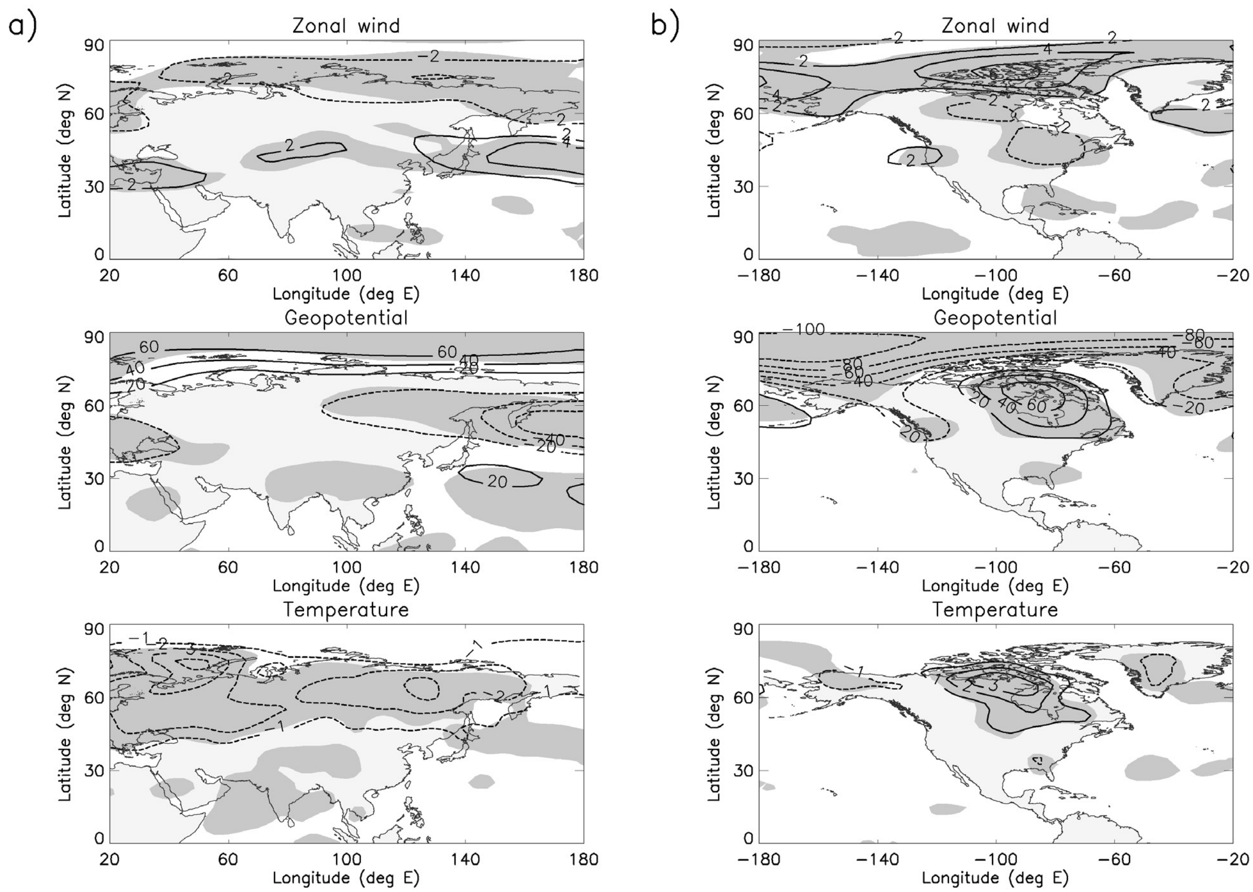

Figure 9. As in Figure 5 but for (a) high minus low Eurasian snow cover springs and (b) positive minus negative phases of summer BI_{WPA}.

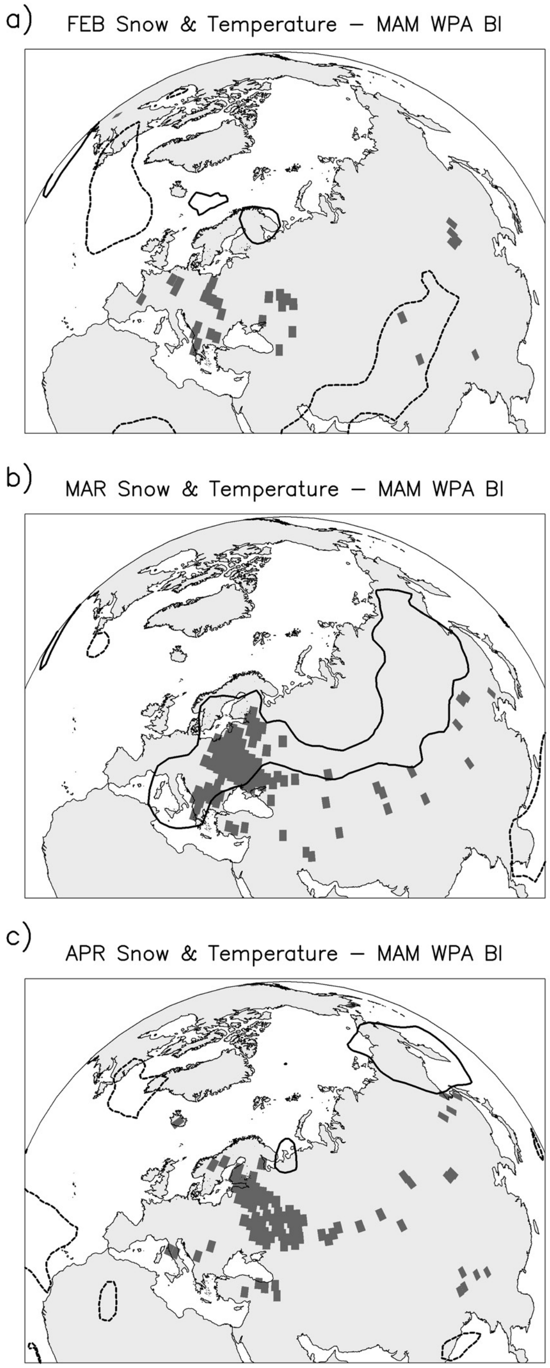

Figure 10. Significant correlation map between spring BI_{WPA} and monthly snow-covered cells (shaded areas) and surface temperature (contour lines) in (a) February, (b) March, and (c) April. Shaded areas indicate significant negative correlations between monthly snow-covered cells and spring BI_{WPA} at $p < 0.05$ significance level. The $p < 0.05$ significance level for correlations between monthly surface temperature and spring BI_{WPA} is represented by solid (positive correlations) and dashed (negative correlations) lines.

Table 1. Teleconnections Between Snow-Covered Regions of Eurasia and North America Computed as Significant Cross Correlations at $p < 0.05$ Significance Level

Sector	Spring KAM	Summer KAM	Summer HUD
Spring SCAN	0.42	0.68	0.54
Spring KAM	· · ·	0.65	0.50
Summer KAM	· · ·	· · ·	0.66

response being confined near the Hudson's Bay (HUD). Correlations of Table 1 reveal that summer snow anomalies over HUD are preceded by fluctuations in snow over the SCAN and KAM regions. This result implies a remote teleconnection between snow-covered areas of Eurasia and North America. Since winter ATL blocking modulates snow cover over SCAN and KAM in spring, summer HUD snow cover anomalies may be attributed to the propagation of the blocking-induced Eurasian snow cover anomalies in spring into the North American continent in summer. Figures 11a and 11b confirm that summer HUD snow cover is related to the previous anomalous cooling induced by spring SCAN snow cover anomalies over western Europe and the resulting weakened polar jet stream. Thus the induced thermal forcing and the anomalous westerlies over and downstream of Eurasia may initiate the snow cover response over HUD. The influence that WPA blocking exerts on the position and intensity of the polar jet stream and the surface temperatures in eastern Eurasia (Figure 5b) provides an additional feasible mechanism in the propagation of snow cover anomalies from Eurasia to North America (see Figure 2a). The BI_{WPA}-related signatures of Figure 9b reveal significant impacts over North America in summer characterized by above-normal temperatures near HUD. These anomalous circulation patterns resemble those obtained from excessive snow in the previous March (see Figures 14, 15 and 17 in YA91). *Kodera and Chiba* [1989] also found that Eurasian snow cover anomalies in spring are related to the stationary summer Okhotsk ridge intensity, which exerts a significant impact in the summer climate of Eurasia and North America. *Déry et al.* [2005] have also reported a connectivity between snow-covered areas of Eurasia in spring and the Canadian snow water equivalent and river discharge, the linkage being attributed to the persistence of the Eurasian snow cover anomalies.

[30] Up to now (stages 1 to 3), we have provided observational evidence of a significant linkage between winter ATL blocking and the succeeding spring Eurasian and summer North American snow cover. These results are consistent with *Ogi et al.* [2003] who described that high-latitude NH summer climate, including snow cover, is strongly influenced by the previous winter climatic variability. Next we will focus our attention in examining how the cycle may continue to modify the ATL blocking activity in the next winter.

3.4. Stage 4: Role of Summer Snow Cover in Summer and Winter Climate

[31] In summer, NH snow-covered areas are confined to Greenland, the Canadian Archipelago and the highest mountain regions, with the last snow detected during July-August. The HUD region, in North America, the Himalayas, northern Siberia and KAM, in Eurasia, and southern Green-

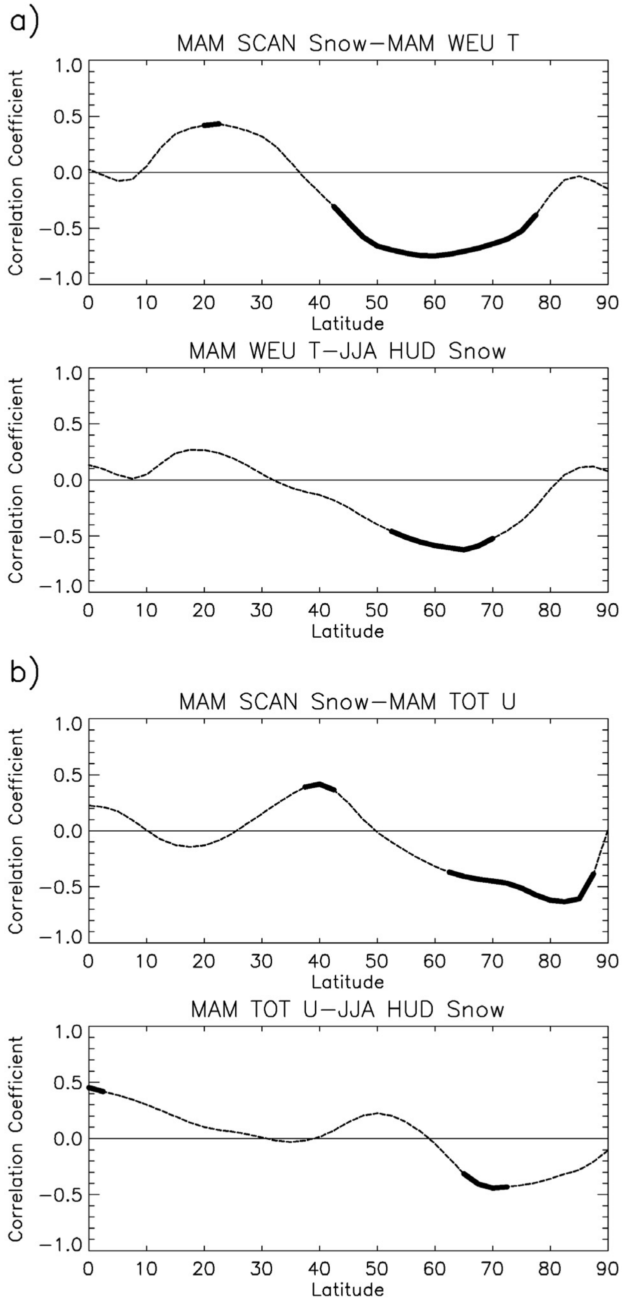

Figure 11. (a) Latitudinal distribution of the correlation coefficient (top) between spring SCAN snow cover and spring surface temperature and (bottom) between spring surface temperature and the upcoming summer HUD snow cover. Temperature has been zonally averaged over the western European sector [0, 60]°E. Solid thick lines indicate significant correlations at $p < 0.05$ significance level. (b) As in Figure 11a but for the NH zonal mean zonal wind.

land exhibit the highest variability in summer snow cover and so the greatest potential for modulating atmospheric variability. The correlation map between winter BI_{ATL} and the subsequent summer snow (Figure 12a) shows significant correlations over northern North America and eastern Eurasia, the pattern being similar to that obtained between summer gridded snow cover and the upcoming winter BI_{ATL} (Figure 12b). Thus, despite that summer Eurasian snow cover is significantly linked with the subsequent winter ATL blocking, the linkage between summer snow-covered areas

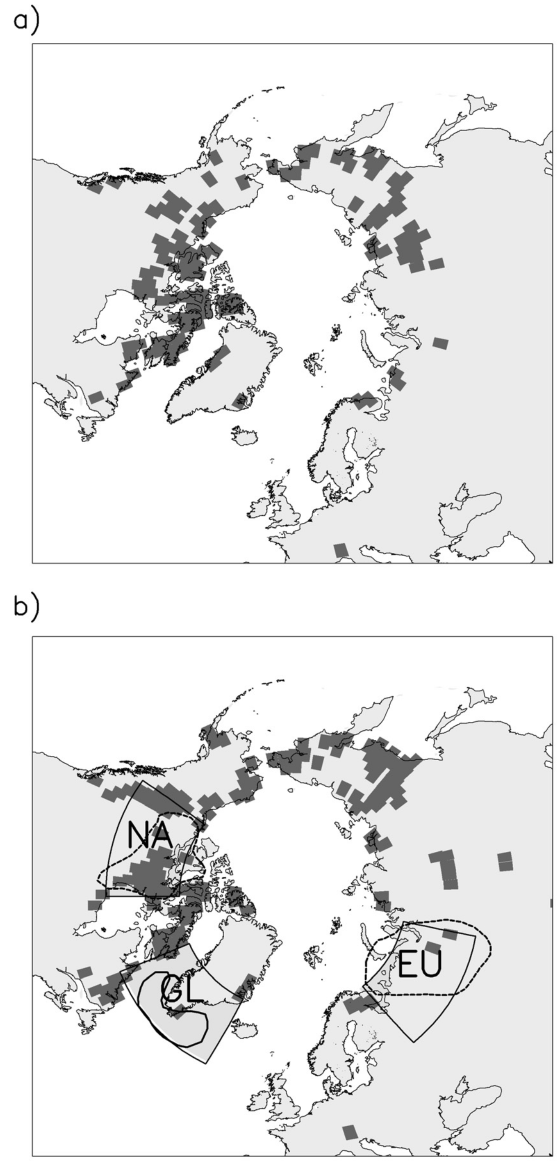

Figure 12. (a) Significant correlation map between winter BI_{ATL} and gridded snow in the upcoming summer. (b) As in Figure 12a but between summer snow-covered cells and the subsequent winter BI_{ATL} (shaded cells) and between summer North American snow cover and summer surface temperature (contour lines). Shaded areas indicate significant positive correlations at $p < 0.05$ significance level. The $p < 0.05$ significance level for correlations between summer North American snow cover and contemporaneous surface temperature is represented by solid (positive correlations) and dashed (negative correlations) lines. Significant temperature correlated areas are marked with black solid lines and are referred to in the text.

Table 2. Significant Cross Correlations Between Some Seasonal Predictors and the Upcoming Winter BI_{ATL} at $p < 0.05$ Significance Level

Season	Spring			Summer		
Region	EUR		NAM	EUR		NAM
Sector	SCAN	KAM	HUD	SCAN	KAM	HUD
r(snow, BI_{ATL})	0.63	. . .	. . .	. . .	0.57	0.57

and the upcoming winter ATL blocking is mostly confined to North America (see Figures 1b and 1c). Table 2 reveals that snow-covered areas over HUD, in North America, are the main summer contributor to the upcoming winter ATL blocking, while KAM provides the highest skill over Eurasia. These results compare favorably with those of SA03 who recently found that June–July (JJ) NH snow cover anomalies favor a negative NAO pattern in the subsequent winter. They demonstrate that the linkage arises from the asymmetrical surface thermal pattern induced by JJ snow cover which, in turn, is a potential precursor of winter climate. To examine the hypothesis of SA03, the correlation map between summer North American snow cover and contemporaneous surface air temperature field has been computed (Figure 12b). The associated thermal pattern resembles that obtained by SA03 between JJ NH snow cover and 2 m air temperature. To assess the role of the asymmetrical thermal pattern we define an index of subpolar temperature difference (equation (2)) from those significantly correlated sectors of Figure 12b, as in SA03. These areas are 40° in longitude and they expand between 57.5°N and 70°N.

$$T_{SP} = \frac{\left(\overline{T}_{NA} + \overline{T}_{EU}\right)}{2} - \overline{T}_{GL} \qquad (2)$$

[32] T_{SP} correlates significantly with the upcoming winter BI_{ATL} (r = −0.70). Additionally, the averaged frequency of ATL blocking days in winter for preceding summers of high T_{SP} was 7.6, significantly lower than those 19.0 obtained for low summer T_{SP} after a t-test (Figure 13a). These results indicate the existence of a dynamical link between summer and winter climates subject to the longitudinal asymmetry of extratropical temperature distributions. Recently, this index has also been identified as the best lagged predictor of the next winter NAO among those found in the literature [*Fletcher and Saunders*, 2006, hereinafter referred to as FS06]. Here, summer T_{SP} also provides correlation coefficients with the next winter BI_{ATL} higher than those obtained with any snow-covered region (see Table 2).

[33] According to FS06, excessive summer snow cover makes the signal feedback into the atmosphere through an anomalous zonal thermal distribution (negative T_{SP} pattern), with above-normal land temperature anomalies over NA and EU and below-normal temperature anomalies over the north Atlantic ocean (GL). Since blocking activity is sensitive to thermal land-sea contrasts [BP06], T_{SP} may exert a significant influence in summer blocking activity. Simultaneous correlations between summer T_{SP} and BI indexes reveal significant linkages over EUR and WPA sectors (r = 0.51 and r = 0.52, respectively). A t-test in blocking days frequency also reveals that blocking occurrence is 47.7% and 38.7% greater over EUR and WPA when summer T_{SP} is positive (Figure 13b). These results indicate that excessive North American snow cover in summer affects summer atmospheric circulation, by forcing anomalous zonal thermal distributions (negative T_{SP}) and opposite blocking

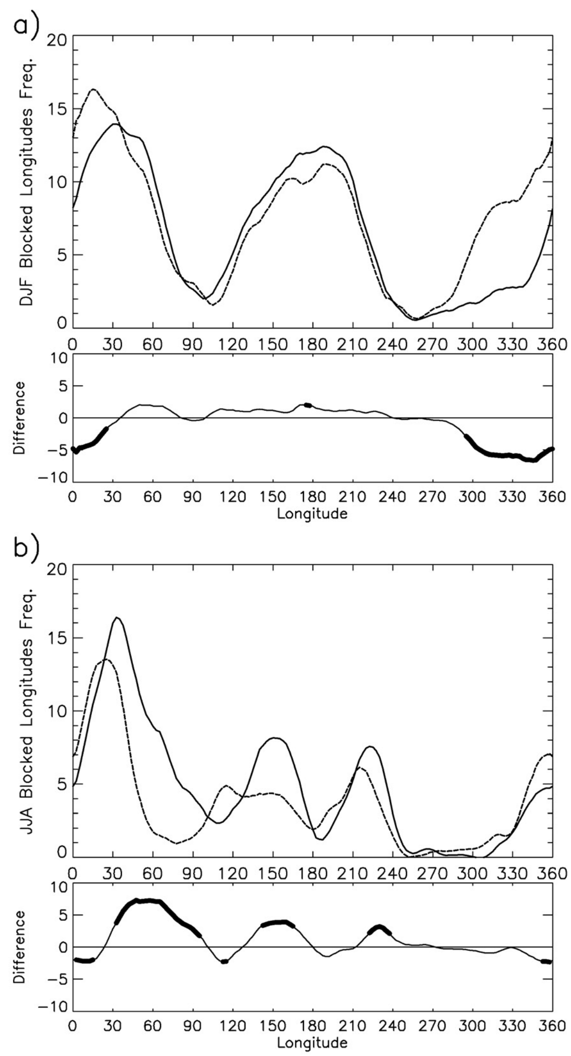

Figure 13. (a) Frequency composites of blocked days in winter for the positive (solid line) and negative (dashed line) phases of the previous summer T_{SP}. The graphic at the bottom displays frequency differences for the positive minus negative phases of T_{SP}. Solid thick lines indicate significant differences at the $p < 0.05$ significance level. (b) As in Figure 13a but in summer.

pattern responses over EUR and WPA. *Barriopedro et al.* [2006b] have reported that, in addition to the summer T_{SP}–winter linkage, T_{SP} is also strongly linked to the summer Northern Annular Mode (NAM) [*Ogi et al.*, 2004], with positive summer T_{SP} inducing an enhanced thermal contrast along the Arctic coast and strengthening the polar vortex. The anomalous westerlies may favor a positive phase of summer NAM by inducing a wave train pattern with ridges located over SCAN, KAM and HUD. Positive summer NAM years are characterized by a double-jet structure which tends to cause atmospheric blocking over the Arctic coasts of Eurasia and North America [*Ogi et al.*, 2004]. Thus the linkage between summer T_{SP} and summer blocking is in agreement with the anomalous induced summer NAM. According to that, positive values of summer T_{SP} force a positive phase of summer NAM and enhance blocking activity over EUR and WPA sectors.

[34] Since the linkage between summer T_{SP} and winter climate vanishes after summer, the zonal temperature gradient should make the predictive signal persist in another long-memory boundary variable that could feedback onto the atmosphere at a later time. By quantifying the relationships between different lagged predictors of winter climate, FS06 found significant linkages between summer T_{SP} and the JJASO EOF2 mode of North Atlantic SSTs [*Saunders and Qian*, 2002]. They proposed that, summer T_{SP} induces contemporaneous midlatitude zonal wind anomalies over the North Atlantic characterized by an enhanced polar jet and a weakened midlatitttude jet, which force persisting North Atlantic SST anomalies with a time lag of one month (see Figure 2b).

3.5. Stage 5: Role of Autumn ATL Blocking in October Eurasian Snow Cover

[35] Although North Atlantic SSTs could be a firm contributor to the summer to winter linkage, the weak SST predictive signal from October to December implies that SST anomalies may not be enough to induce changes in the upcoming winter. Thus persistence of the anomalous SST patterns should force either a direct thermal response in winter or initiate a response via a third variable [FS06]. To explain the signal propagation from autumn to winter, several possibilities have been proposed. The first one implies a direct response from SST anomalies through the upward propagation of anomalous wave activity into subpolar stratosphere [*Polvani and Waugh*, 2004]. Another one postulates that SST could affect either local sea ice outflow [*Kvamsto et al.*, 2004; *Hilmer and Jung*, 2000] and/or induce, by means of atmospheric circulation anomalies, remote snow cover anomalies in early autumn [FS06], which have also been considered a potential forcing factor of the next winter climate variability.

[36] Here we investigate the latter possibility in order to find anomalous circulation patterns capable of propagating autumn signal from the North Atlantic into another slow-varying boundary variable such as Eurasian snow cover. Simultaneous correlations reveal an additional linkage between autumn BI_{ATL} and Eurasian snow cover, being maximum over the SCAN sector, where the correlation coefficient reaches 0.63. On monthly timescales the snow cover response is mainly attributable to September BI_{ATL}, which influences autumn Eurasian snow cover by forcing

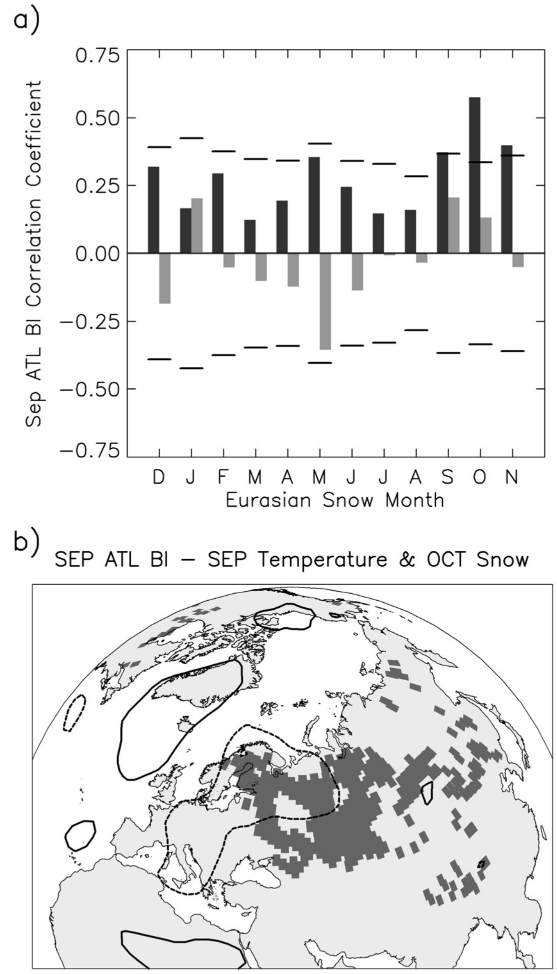

Figure 14. (a) As in Figure 8 but for BI_{ATL} in September and monthly Eurasian snow cover. (b) As in Figure 4 but between September BI_{ATL} and the upcoming October gridded snow cover (shaded areas) and contemporaneous surface temperature (contour lines).

an appreciable snow cover response from September to November, especially evident in October (r = 0.58) (Figure 14a). To examine this relationship, the correlation map between September BI_{ATL} and the coming October snow cover has been computed (Figure 14b). The most significant areas spread from the SCAN sector to most of western and central Eurasia, coinciding with those regions of maximum snow cover variability. According to that, an enhanced blocking activity in early autumn precedes an increase in the expected appearance of Eurasian snow cover in October within a time lag of one month (see Figure 2b). Eurasian snow cover starts to increase in October, partially because temperature remains below 0°C and precipitation occurs as snowfall. As discussed previously, the spatial distribution patterns of both, temperature and precipitation, can be altered by blocking activity. The blocking-associated thermal distribution of Figure 14b suggests that the anomalous cooling induced over western Eurasia could accelerate the snow accumulation by maintaining temperatures below 0°C. This blocking feature provides a potential mechanism by which autumn ATL blocking may modulate the climatological appearance of snow cover and a presumably dynamical atmospheric pathway in the connection between JJASO North Atlantic SST and Eurasian snow cover. However, further investigations are required on the role of North Atlantic SST anomalies in blocking occurrence.

4. Role of the Stratosphere

[37] Recently, autumn snow cover anomalies over Eurasia have been identified as a forcing source of the upcoming winter atmospheric variability, especially in October when the snow cover fluctuations can vary by a factor of five [*Gong et al.*, 2002]. Observational and numeric experiments have shown a linking mechanism between autumn and winter climates [*Saito et al.*, 2001; *Cohen et al.*, 2002; *Gong et al.*, 2002]. According to that, the October snow cover anomalies propagate from the surface weakening the polar vortex via the upward vertical flux of Rossby waves. In early winter, when troposphere and stratosphere are strongly coupled, the forced negative zonal wind anomalies propagate down from the stratosphere to the troposphere determining a negative AO/NAO-like pattern (see Figure 2b).

[38] The significant role of the NAO in winter ATL blocking variability has also been reported, with 67% more blocking days during the negative phase of NAO and about 45% of explained variance [*Shabbar et al.*, 2001; BP06]. Given that AO, NAM and NAO are not separable in winter [*Rogers and McHugh*, 2002; *Ogi et al.*, 2004], the linkage between winter blocking and NAO is equally valid for the winter AO/NAM. In fact, the correlation coefficient between AO and BI_{ATL} in winter reaches -0.71, which is close to that obtained for the winter NAO ($r = -0.79$). The strong association between winter ATL blocking and the AO/NAO suggests that this vertical mechanism via the stratosphere could also force an ATL blocking response in the next winter. Thus excessive October snow cover could determine a negative AO/NAO response in winter then favoring winter blocking occurrence over the ATL sector. According to that, the AO/NAO could act as a linkage mechanism between winter ATL blocking and the preceding snow. Such relationship identifies the stratosphere as a feasible contributor in the linkage between summer snow and winter ATL blocking. This stratospheric connection will be considered the stage 6 in our model (see Figure 2b).

[39] Thus the second phase in the ATL blocking-snow cover relationship (stages 4 to 6) provides a likely dynamical linkage in the propagation of the signal from subpolar areas of North America in summer to the ATL sector in the upcoming winter and completes an interannual persistent cycle of teleconnections. However, the supporting linking mechanisms of the summer to winter linkage may also involve SSTs anomalies, sea-ice and stratospheric processes, which should be subject to future investigations.

5. Discussion

5.1. Snow Cover and the Long-Term Blocking Trends

[40] Recently, BP06 have reported an upward trend in blocking days over the WPA sector, especially significant in spring. It could be partially attributed to concurrent changes in the distribution of anomalous temperature patterns, which are important forcing factors of blocking variability (see the subsection b of the introduction). As snow cover regulates the surface temperature over Eurasia, the downward trend in snow cover extent is a likely precursor of the observed trends in the WPA sector. This study confirms that the association between Eurasian snow cover and spring WPA blocking through the induced formation of anomalous thermal patterns is stronger than that of any extratropical atmospheric TCPs, even when the temporal series are not detrended. These results are suggestive of the potential role of Eurasian snow cover in forcing the recently described blocking trends over the WPA sector.

[41] According to the proposed conceptual model, the interseasonal linkages between snow cover and winter ATL blocking act as a positive feedback to reinforce each other. Thus the snow cover may also regulate the downward trend observed in winter ATL blocking [BP06]. However, there would be negative contributions preventing the cycle from perpetuating. Previous studies have suggested that the observed multidecadal trends in extratropical atmospheric flow, such as the positive trend in the winter NAO, may also be attributable to other variables of long memory such as sea-ice coverage and/or SSTs. Additional efforts should be made in order to address the role of these additional forcing factors in blocking variability.

5.2. On the Propagation of Snow Cover Anomalies

[42] The strong association between snow cover and the next winter climate has increased the interest in snow cover as a predictor tool. However, there is no agreement on the timing and location of the origin of the winter climate signal. Thus April–October Eurasian [*Bojariu and Gimeno*, 2003] and summer NH [SA03] snow cover may both affect the upcoming winter circulation. Our results are consistent with both viewpoints, suggesting that both spring Eurasian and summer North American snow cover are significantly connected (see Table 2). We provide a dynamical link consisting in the propagation of snow cover signal into North American summer through the adjustment of Rossby waves and the anomalous zonal winds resulting from the previous Eurasian spring snow cover anomalies. However, this may not be the only mechanism. The summer Indian Monsoon and SST Pacific anomalies, among others, might also participate in such propagation. Several studies have suggested that Eurasian snow cover and the subsequent decrease of the Indian summer monsoon rainfall could initiate and eastward extension of warmer water in the equatorial Pacific SST field [*Barnett et al.*, 1989; *Yasunari*, 1990].

[43] On the other hand, additional studies based on observations and model outputs have addressed the influence of tropical and midlatitude Pacific SST anomalies in Pacific blocking. *Mullen* [1989], using different simulations from GCMs, demonstrated that some realistic Pacific SST anomaly patterns were able to alter blocking onset in certain geographical regions. *Tibaldi et al.* [1997] suggested that regional Pacific blocking may partially be influenced by the SSTs. Moreover, BP06, based on a 55-year observational study, described that ENSO favors preferred locations for blocking formation. These results identify an additional feasible pathway in the propagation of snow cover anoma-

lies from Eurasia to North America by means of SST-induced atmospheric patterns over the Pacific sector leading to North American snow cover anomalies.

6. Conclusions

[44] In this paper we examine the multiseasonal relationship between NH snow cover and regional atmospheric blocking for a 31-year period. The main conclusion is that snow cover anomalies can modulate regional blocking activity, especially over the ATL and WPA sectors, where atmospheric blocking is sensitive to snow cover fluctuations occurring over Hudson's Bay and western Eurasia, respectively. Simultaneously, persistent blocking patterns over these sectors (ATL and WPA) act as a controlling mechanism of snow cover interannual variability, with the induced snow cover response appearing essentially confined to certain areas (western and eastern Eurasia and Hudson's Bay). These results confirm the importance of regional snow anomalies in driving both, large-scale atmospheric circulation and continental snow cover.

[45] Two primary linkages have been found in the blocking-snow cover relationship at seasonal scales: one, where winter ATL blocking leads spring (summer) Eurasian (North American) snow cover anomalies and a second one with spring (summer) Eurasian (North American) snow cover resulting in an anomalous ATL blocking activity in the next winter. Both linkages are consistent with those reported in the literature between winter NAO and the next summer high-latitude climate [*Ogi et al.*, 2003] as well as with those concerning spring (summer) Eurasian (NH) snow cover and the upcoming winter climate [*Bojariu and Gimeno*, 2003; SA03]. It is found that, in addition to summer snow cover anomalies, spring Eurasian snow cover may also influence the winter climate earlier than summer. Since there is persistence from one winter to the next [*Johansson et al.*, 1998], a portion of the winter predictability arising from spring may result from the influence of the previous winter ATL blocking on spring snow. However, spring snow provides better correlations with the next winter than those obtained with winter ATL blocking leading to spring snow, suggesting that some predictability may reside in spring snow.

[46] A detailed inspection at shorter timescales revealed that these linkages seem to occur through a sequence of teleconnections involving different timings and regional snow-covered regions, depending on the climatological evolution of snow extent. It is suggested that the linking mechanisms result from the associated formation of anomalous thermal patterns. According to that, the snow cover plays a significant role in determining blocking activity over the ATL and WPA sectors by exciting an anomalous land-sea thermal contrast. Simultaneously, the thermal advection of anomalous blocking patterns may act as a forcing mechanism of the subsequent snow cover fluctuations. ATL blocking in winter (autumn) exerts an important influence by modulating the snow disappearance (appearance) timing, while WPA may be more related with the propagation of snow cover anomalies.

[47] Figure 2a (Figure 2b) summarizes the temporal stages in the blocking (snow cover)-leading linkage, with numbers labeling those stages described in the previous sections. These interactions seem to determine a complex atmospheric blocking-snow cover conceptual model. According to that, an enhanced winter ATL blocking activity during early and mid winter is responsible of a reduced warm advection over western Eurasia, determining slower melting speeds, an above-normal March snow extent and hence a later snow disappearance timing (stage 1; Figure 2a). Snow cover anomalies persist during the melting season probably because of the quasi-simultaneous local albedo effect and the following lagged snow-hydrological feedback. The subsequent cooling derived from excessive snow cover and the increased soil moisture after the snowmelt inhibits blocking activity over the WPA sector by inducing a distinctive cold continent-warm ocean pattern (stage 2; Figure 2a). Simultaneously, the persistent negative blocking pattern in spring appears as a feasible contributor in the propagation of snow cover anomalies toward the snow-covered areas of eastern Eurasia and the Hudson's Bay region, which seem to result from the anomalous zonal winds induced by previous spring snow cover anomalies (stage 3; Figure 2a). Once the snow cover anomalies are delivered there in early mid summer, the induced excessive snow cover results in an asymmetrical thermal contrast T_{SP}, favoring blocking-like patterns over EUR and WPA in summer as well as an anomalous North Atlantic SSTs response persisting throughout early autumn [FS06] (stage 4; Figure 2b). The linkage between autumn and winter climates is uncertain but it could be related with anomalous atmospheric patterns capable of propagating the SST signal into another boundary variable of long memory. The ability of autumn ATL blocking to induce snow cover anomalies is a possibility, although subject to further investigations. The enhanced cold advection of the autumn ATL blocking over western Eurasia seems to accelerate the snow cover appearance, determining an anomalous above normal increase of snow cover from September to October and earlier snow appearance timings (stage 5; Figure 2b). As has been reported in the literature, the excessive snow cover signal in October excites the vertical Rossby waves propagation into the stratosphere and returns in December as negative zonal wind anomalies to determine a negative NAO/AO pattern in the subsequent winter, which favors ATL blocking activity (stage 6; Figure 2b), starting a new cycle again.

[48] **Acknowledgments.** The authors wish to thank Rutgers University for providing the snow cover data. We would also like to thank to three anonymous reviewers for their valuable comments. Ricardo Trigo and Roxana Bojariu contributed to improve this manuscript. The Spanish Science and Technology Department supported this study trough the VALIMOD (Climatic VALIdation of Conceptual MODels) project (REN2002-04558-C04-01).

References

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

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.

Barriopedro, D., R. García-Herrera, A. R. Lupo, and E. Hernández (2006a), A climatology of Northern Hemisphere blocking, *J. Clim.*, *19*, 1042–1063.

Barriopedro, D., R. García-Herrera, and E. Hernández (2006b), The role of snow cover in the Northern Hemisphere winter to summer transition, *Geophys. Res. Lett.*, *33*, L14708, doi:10.1029/2006GL025763.

Blanford, H. F. (1884), On the connexion of the Himalaya snowfall with dry winds and seasons of drought in India, Proc, *R. Soc., Ser. A and Ser. B*, *37*, 1–23.

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.

Charney, J. G., and J. G. DeVore (1979), Multiple flow equilibria in the atmosphere and blocking, *J. Atmos. Sci.*, *36*, 1205–1216.

Clark, M. P., and M. C. Serreze (2000), Effects of variations in East Asian snow cover on modulating atmospheric circulation over the North Pacific Ocean, *J. Clim.*, *13*, 3700–3710.

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

Cohen, J., K. Saito, and D. Entekhabi (2001), The role of the Siberian High in Northern Hemisphere climate variability, *Geophys. Res. Lett.*, *28*, 299–302.

Cohen, J., D. Salstein, and K. Saito (2002), A dynamical framework to understand and predict the major Northern Hemisphere mode, *Geophys. Res. Lett.*, *29*(10), 1412, doi:10.1029/2001GL014117.

Déry, S. J., J. Sheffield, and E. F. Wood (2005), Connectivity between Eurasian snow cover extent and Canadian snow water equivalent and river discharge, *J. Geophys. Res.*, *110*, D23106, doi:10.1029/2005JD006173.

Dickson, R. R. (1984), Eurasian snow cover versus Indian monsoon rainfall—An extension of the Hahn-Shukla results, *J. Clim. Appl. Meteorol.*, *23*, 171–173.

Egger, J. (1978), Dynamics of blocking highs, *J. Atmos. Sci.*, *35*, 1788–1801.

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

Frederiksen, J. S. (1982), A unified three-dimensional instability theory of the onset of blocking and cyclogenesis, *J. Atmos. Sci.*, *39*, 969–987.

Gong, G., D. Entekhabi, and J. Cohen (2002), A large-ensemble model study of the wintertime AO/NAO and the role of interannual snow perturbations, *J. Clim.*, *15*, 3488–3499.

Groisman, P. Y., T. R. Karl, R. W. Knight, and G. L. Stenchikov (1994), Changes of snow cover, temperature, and radiative heat balance over the Northern Hemisphere, *J. Clim.*, *7*, 1633–1656.

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

Hilmer, M., and T. Jung (2000), Evidence for a recent change in the link between the North Atlantic Oscillation and Arctic sea ice export, *Geophys. Res. Lett.*, *27*, 989–992.

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.

Iwasaki, T. (1991), Year-to-year variation of snow cover area in the Northern Hemisphere, *J. Meteorol. Soc. Jpn.*, *69*, 209–217.

Johansson, A., A. Barnston, S. Saha, and H. van den Dool (1998), On the level and origin of seasonal forecast skill in northern Europe, *J. Atmos. Sci.*, *55*, 103–127.

Kalnay, E., et al. (1996), The NCEP/NCAR 40–years reanalyses project, *Bull. Am. Meteorol. Soc.*, *77*, 437–471.

Kikuchi, Y. (1971), Influence of mountains and land-sea distribution on blocking action, *J. Meteorol. Soc. Jpn.*, *47*, 564–572.

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

Kripalani, R. H., and A. Kulkarni (1999), Climatology and variability of historical Soviet snow depth data: Some new perspectives in snow–Indian monsoon teleconnections, *Clim. Dyn.*, *15*, 475–489.

Kvamsto, N. G., P. Skeie, and D. B. Stephenson (2004), Impact of Labrador sea-ice on the North Atlantic Oscillation, *Int. J. Climatol.*, *24*, 603–612.

Leathers, D. J., and D. A. Robinson (1993), The association between extremes in North American snow cover extent and United States temperatures, *J. Clim.*, *6*, 1345–1355.

Liu, Q. (1994), On the definition and persistence of blocking, *Tellus, Ser. A*, *46*, 286–290.

Mullen, S. L. (1989), Model experiments on the impact of Pacific sea surface temperature anomalies on blocking frequency, *J. Clim.*, *2*, 997–1013.

Namias, J. (1985), Some empirical evidence for the influence of snow cover on temperature and precipitation, *Mon. Weather Rev.*, *113*, 1542–1553.

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.

Oort, A. H., and J. J. Yienger (1996), Observed interannual variability in the Hadley Circulation and its connection to ENSO, *J. Clim.*, *9*, 2751–2767.

Polvani, L. M., and D. W. Waugh (2004), Upward wave activity flux as precursor to extreme stratospheric events and subsequent anomalous surface weather regimes, *J. Clim.*, *17*, 3548–3554.

Reinhold, B. B., and R. T. Pierrehumbert (1982), Dynamics of weather regimes: quasi-stationary waves and blocking, *Mon. Weather Rev.*, *110*, 1105–1145.

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

Rogers, J. C., and M. J. McHugh (2002), On the separability of the North Atlantic Oscillation and the Arctic Oscillation, *Clim. Dyn.*, *19*, 599–608.

Saito, K., and J. Cohen (2003), The potential role of snow cover in forcing interannual variability of the major Northern Hemisphere mode, *Geophys. Res. Lett.*, *30*(6), 1302, doi:10.1029/2002GL016341.

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

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.

Shabbar, A., J. Huang, and K. Higuchi (2001), The relationship between the wintertime North Atlantic Oscillation and blocking episodes in the North Atlantic, *Int. J. Climatol.*, *21*, 355–369.

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

Simmons, A. J., J. M. Wallace, and G. W. Branstator (1983), Barotropic wave propagation and instability, and atmospheric telleconnection patterns, *J. Atmos. Sci.*, *40*, 1363–1391.

Tibaldi, S., F. D'Andrea, E. Tosi, and E. Roeckner (1997), Climatology of Northern Hemisphere blocking in the ECHAM model, *Clim. Dyn.*, *13*, 649–666.

Trigo, R. M., I. F. Trigo, C. C. DaCamara, and T. J. Osborn (2004), Winter blocking episodes in the European-Atlantic sector: Climate impacts and associated physical mechanisms in the reanalysis, *Clim. Dyn.*, *23*, 17–28.

Tsou, C. H., and P. J. Smith (1990), The role of synoptic/planetary-scale interactions during the development of a blocking anticyclone, *Tellus, Ser. A*, *42*, 174–193.

Tung, K. K., and R. S. Lindzen (1979), A theory of stationary long waves. Part I: A simple theory of blocking, *Mon. Weather Rev.*, *107*, 714–734.

Walker, G. R. (1910), Correlations in seasonal variations of weather II, *Mem. Indian Meteorol. Dep.*, *21*, 22–45.

Walland, D. J., and I. Simmonds (1997), North American and Eurasian snow covariability, *Tellus, Ser. A*, *49*, 503–512.

Yamazaki, K. (1989), A study of the impact of soil moisture and surface albedo changes on global climate using the MRI GCM-I, *J. Meteorol. Soc. Jpn.*, *67*, 123–146.

Yasunari, T. (1990), Impact of Indian monsoon on the coupled atmosphere/Ocean systems in the tropical Pacific, *Meteorol. Atmos. Phys.*, *44*, 29–41.

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.

D. Barriopedro and R. García-Herrera, Departamento de Física de la Tierra II, Facultad Ciencias Físicas, Universidad Complutense de Madrid, E-28040 Madrid, Spain. (rgarciah@fis.ucm.es)